Manuscript Number:	
Article Type:	Article
Section/Category:	Regular Issue
Full Title:	Estimation of earthquake early warning parameters for eastern Sicily
Corresponding Author:	Giuseppina Tusa, PhD Istituto Nazionale di Geofisica e Vulcanologia Catania, ITALY
Corresponding Author's Institution:	Istituto Nazionale di Geofisica e Vulcanologia
Corresponding Author E-Mail:	giuseppina.tusa@ingv.it
Order of Authors:	Giuseppina Tusa, PhD
	Carla Musumeci, PhD
	Domenico Patanè, Dr.
Abstract:	Earthquake Early Warning Systems (EEWS) are becoming a suitable instrument for seismic risk management in real-time. In fact, they are implemented or undergoing testing in many countries around the world, since represent an effective approach to mitigate seismic risk at short time-scale. EEW systems are based on the use of relationships between some parameters measured on the initial portion of seismic signal after the onsets. Here, we address with the first approach to the implementation of EEWS in eastern Sicily, a region that has been hit by several damaging destructive earthquakes. We estimated the peak displacement amplitude of first portion of P- and S-waves, Pd, the ground-motion period parameter, τc, and the peak ground-motion velocity, PGV, from earthquakes with ML ≥ 2.8 recorded by the broadband stations operated by the Istituto Nazionale di Geofisica e Vulcanologia. We found that the Pd is correlated with the size of earthquake and may be used to compute the magnitude for an EEW system in this area. We also derived the relationships between τc and ML, and between Pd and PGV, which can be used to provide onsite warning in the area around a given station and evaluate the potential damaging effects. These relationships may be deemed a useful guide for the future implementation of the earthquake early warning system in the region.
Author Comments:	Dear Editor, Please find the paper "Estimation of earthquake early warning parameters for eastern Sicily" (authors: G. Tusa, C. Musumeci, and D. Patané) for first submission on the Bulletin of Seismological Society of America . Best regards Giuseppina Tusa July, 28, 2016
Suggested Reviewers:	Matteo Picozzi, PhD Università di Napoli 'Federico II' matteo.picozzi@unina.it He is one of the major expert of earthquake early warning at italian level.
	Alfonso Brancato, PhD Istituto Nazionale di Geofisica e Vulcanologia alfonso.brancato@ingv.it He is an expert in statistical data analysis and a very knowledgable about the seismic activity that characterizes the eastern Sicily.
Opposed Reviewers:	

1 **Estimation of earthquake early warning parameters for eastern Sicily**

2 Giuseppina Tusa[1], Carla Musumeci[1], Domenico Patanè[1]

3 [1] Istituto Nazionale di Geofisica e Vulcanologia, Italy

4 Domenico Patanè is also at Instituto Andaluz de Geofísica, Universidad de Granada, Granada,

5 Spain

6

7 <div align="center">Abstract</div>

8 Earthquake Early Warning Systems (EEWS) are becoming a suitable instrument for seismic

9 risk management in real-time. In fact, they are implemented or undergoing testing in many

10 countries around the world, since represent an effective approach to mitigate seismic risk at short

11 time-scale. EEW systems are based on the use of relationships between some parameters measured

12 on the initial portion of seismic signal after the onsets. Here, we address with the first approach to

13 the implementation of EEWS in eastern Sicily, a region that has been hit by several damaging

14 destructive earthquakes. We estimated the peak displacement amplitude of first portion of P- and S-

15 waves, P_d, the ground-motion period parameter, τ_c, and the peak ground-motion velocity, PGV,

16 from earthquakes with $M_L \geq 2.8$ recorded by the broadband stations operated by the Istituto

17 Nazionale di Geofisica e Vulcanologia. We found that the P_d is correlated with the size of

18 earthquake and may be used to compute the magnitude for an EEW system in this area. We also

19 derived the relationships between τ_c and M_L, and between P_d and PGV, which can be used to

20 provide onsite warning in the area around a given station and evaluate the potential damaging

21 effects. These relationships may be deemed a useful guide for the future implementation of the

22 earthquake early warning system in the region.

23

24 **Keywords:** Earthquake ground motion, earthquake early warning, seismic hazard mitigation

Introduction

The structural setting of eastern Sicily is connected to the complex tectonic environment of

the Central Mediterranean basin (Fig. 1a, b) which is subjected to the NNW convergence between

the Eurasian and African plates and to the geodynamic processes due to the opening of the

Tyrrhenian basin (Faccenna *et al*., 2001). This tectonic setting makes the eastern Sicily one of the

most hazardous zones in Italy, characterized by a significant rate of crustal seismicity. The area, in

fact, has undergone some disastrous earthquakes. Among them, we can mention the ones occurring

on 4 February 1169 (moment magnitude M_w=6.4 by the 2011 version of the Catalogo Parametrico

dei Terremoti Italiani (Rovida *et al*., 2011), hereafter CPTI11; I=X on the MSC scale), 10

December 1542 (M_w=6.8 by CPTI11; I=IX–X), 9 and 11 January 1693 (M_w=6.2 and 7.4 by

CPTI11, respectively; I=IX and X–XI, respectively), 20 February 1818 (M_w=6.2 by CPTI11;

I=VIII-IX) and on 28 December 1908 (M_w=7.1 by CPTI11; I=X;), which is one of the strongest

earthquakes ever to have occurred in the central Mediterranean (Boschi *et al*., 1995). In particular,

the 1693 and 1908 earthquakes completely destroyed the cities of Catania and Messina,

respectively, and were followed by large tsunamis, as well. More recently, eastern Sicily has

experienced minor events. In 13 December, 1990 a seismic event (I=VIII) produced several

damages in a wide area despite its magnitude M_s was equal to 5.4 (Amato *et al*., 1995). It caused

severe damage to the cities of Augusta and Catania, and 19 casualties, as well, renewing the

attention in the seismic potential of eastern Sicily and the socio-political consequence of the

earthquakes in the region.

In eastern Sicily, an important role is played by the volcanic area of Mt. Etna (Fig. 1a).

Indeed, its seismicity creates a rather specific scenario, with relatively small earthquakes, very

shallow, that can produce important damage on a local scale. The most recent examples of

destructive seismic events belong to a seismic swarm (max M_L 4.8) occurring at Mt. Etna in

October 2002. The most damaging event of the swarm (M_L = 4.5; I=VIII) occurred on 29 October

50   2002 (Castello *et al*., 2006). It struck a densely urbanized area on the southeastern flank of Mt.

51   Etna, producing . heavy damages even to reinforced concrete structures (Azzaro *et al*., 2006).

52   In the framework of several national projects, different research fields are supported and

53   furthered, including earthquake hazard mitigation in eastern Sicily, as well. In particular, two

54   closely related development projects deal with implementing an Earthquake Early Warning (EEW)

55   system in the considered area as a tool for real-time seismic risk mitigation and management.

56   Indeed, EEW is the current focus of considerable research effort and its potential applicability for

57   the immediate activation of safety measures for critical systems is already undoubted (i.e., Wieland

58   *et al*., 2000; Sato *et al*., 2011). As a matter of fact, the development of these systems is more related

59   to an actual possibility to immediately trigger actions for the protection of strategic sites and

60   lifelines rather than as an instrument giving a massive alarm to the communities (evacuating people

61   from buildings requires warning times so long that rarely are available in urbanized areas).

62   Moreover, due to the recent impulses on the development of earthquake early warning systems

63   worldwide, it is significantly raising the interest toward the potential use of EEW systems for the

64   Structural Control, adopting structure-specific applications (active and semi-active control devices)

65   set up within the leading time so to optimize the expected structural response.

66   Over the last ten years, as a result of the technological evolution in the fields of both

67   computing systems and data transmission, it has been possible to develop more effective techniques

68   to analyse seismic data in real time. Indeed, the Real-Time Seismology (RTS) integrates a real time

69   telemetry system, where the transmission of information takes place with a very low latency, with

70   automatic processing of recorded signals, providing fast and reliable estimates of the main

71   earthquake parameters (location, magnitude) in the first few seconds during its occurrence.

72   On the base of the configuration of the seismic network, EEW systems can be distinguished

73   in two main types: regional and on-site (or site-specific) warning systems (Kanamori, 2005). The

74   regional EEW systems use a dense network of seismic sensors, partially or entirely covering the

75   area where earthquakes are likely to occur, with real-time capability to estimate the source

parameters (event location and magnitude) of earthquake seismic event by using the early portion of recorded seismograms. Therefore, the system uses them to predict a specific peak ground motion at distant target sites through an empirical ground-motion prediction equation. On-site EEW systems, such as the ones installed in Japan (UrEDAS, Nakamura, 1989), in California (ElarmS, Allen and Kanamori, 2003) or in Romania (Wenzel *et al*., 1999; Böse *et al*., 2007), are based on a single seismic station (single-station approach) or an array of seismic stations installed near the target site that needs to be alerted. After detecting the arrival of the faster but weaker P-wave, the system computes the peak amplitude and the predominant period in the very first seconds of the P waves (Wu and Kanamori, 2005a, 2005b) in order to estimate the associated peak ground motion of the more destructive S and surface waves at the target. This approach is relatively simple, but less accurate than the regional approach. It also provides smaller effective "lead-time" (e.g. the time span from the arrival of the damaging waves to the alert notification at a given target site) compared to the regional approach, which also has the advantage that the system is constantly run and tested, and the source parameter estimates gain in accuracy as more data are recorded and analysed. However, the data processing could take a lot of time that the alarm is issued after the ground motion reaches the sites of interest (defining the so-called blind zone, Kanamori, 2005). Meanwhile, regional systems are more effective to applications such as shake maps, very useful for emergency management immediately after the event (Wald *et al*., 1999). At the same time, the site-specific EEW systems are certainly devoted to reducing the exposure of strategic facilities (lifelines, transportation infrastructures, power plants, etc.) in real-time by automated safety actions. The Ignalina nuclear power plant in Lithuania takes advantage of a site-specific EEW system (Wieland *et al*., 2000). The seismic network, composed by six stations, is installed at 30 km from the reactor and ensures an alarm 4 - 8 s before the ground motion affects the power plant. This time is enough to activate the control rods since they need only 2 s to come into use.

The regional and on-site warnings can be combined to give a hybrid EEW system, by using the potential of regional seismic networks to protect different critical systems and/or the population

102   at the same time (Kanamori, 2005; Iervolino *et al*., 2007). Wu and Kanamori (2005) experimented

103   with the method for on-site warning based on the measure of the predominant period (τ_c) of initial

104   portion of P wave by using the data recorded by accelerometric stations of the seismic network in

105   Taiwan. Recently, Zollo *et al*. (2010) developed an integrated regional/on-site early warning

106   method that enables estimating a Potential Damage Zone (PDZ) for the forthcoming earthquake,

107   namely the region in which the most damage is awaited. The method is based on the estimation of

108   the peak ground displacement (P_d) and the predominant period (τ_c), in real-time, at recording sites

109   located at increasing distances from the earthquake epicenter. An alert level is associated to each

110   recording site on the base of critical values of P_d and τ_c. As shown by several authors, these two

111   early warning (EW) parameters are empirically correlated to magnitude (e.g. Wu and Kanamori,

112   2008; Zollo *et al*., 2010; Colombelli *et al*., 2012; Carranza *et al*., 2013) and to peak ground velocity

113   and acceleration (e.g. Böse *et al*., 2007; Zollo *et al*., 2010; Carranza *et al*., 2013).

114   In this work, we determine the two EW parameters, P_d and τ_c, together with the peak ground

115   velocity, *PGV*, by using a set of more than 200 seismic events and explore the use of P_d and τ_c

116   parameters for EEW purposes in the studied area. The aim is to compute specific empirical

117   relationships of these two parameters with earthquake size and peak ground motion parameters for

118   future applications of an EEW system in eastern Sicily. An example of the practical application of

119   the obtained results are performed by the software platform PRESTo (Probabilistic and

120   Evolutionary early warning SysTem, see Data and Resources) by Satriano *et al*. (2011) in

121   simulation mode (from the playback of the actual recorded waveforms). PRESTo integrates

122   algorithms for real-time, rapid earthquake location, magnitude estimation and damage assessment.

123   The code is currently being tested in southern Italy on the Irpinia Seismic Network (ISNet).

124

125   Dataset and record processing

126   Seismic hazard in eastern Sicily is linked to earthquakes occurring in different

127   seismotectonic areas and associated with various types of faulting mechanisms. This should ensure

128   that the effects due to rupture directivity and focal mechanism on peak amplitudes are averaged out.

129   The eastern part of the Sicily is mainly characterized by two active volcanic regions, the

130   Aeolian Archipelago lying in the southern Tyrrenian Sea and the Mount Etna located in the central-

131   eastern Sicily. From a seismological point of view, the northern Sicily and its Tyrrhenian off-shore

132   are characterized by the activity of different tectonic structures associated both to the collision

133   between European and African Plates, and the opening of Tyrrhenian Basin. The seismicity is

134   mainly located in the hinge zone between southern Tyrrhenian and northern Sicily, and is confined

135   in two principal hypocentral sectors (Gueguen *et al*., 2002; Giunta *et al*., 2009). The deep seismicity

136   is essentially connected to the subduction processes of the Ionian lithospheric slab beneath the

137   Calabrian arc and affects the northeastern Sicily. Instead, the shallow seismicity represents the

138   expression of the strain crossing the whole orogeny (Neri *et al*., 1996).

139   Etnean seismic events share their signatures with earthquakes recorded in the near tectonic

140   environments of the Hyblean Plateau and Peloritani-Calabrian Arc (Patanè *et al*., 1997; Patanè and

141   Giampiccolo, 2004). The regional tectonic stresses together with the local stresses connected to the

142   magma migration in the earth's crust, provide the necessary energy for rock failure.

143   The local surface tectonic structures on Mt. Etna, are connected to an intense superficial

144   seismic activity, essentially characterized by earthquakes often clustered in swarms and having

145   focal depths generally less than 3 km (Patanè *et al*., 2004). Although the shallowest events

146   characterize the most central-eastern portion of the volcano, they affect the entire volcanic area,

147   albeit less in number. Occurring in particular geological conditions, they are considered to form a

148   family of events whose characteristic hypocentral location has effects on both seismic scaling laws

149   and wave propagation phenomena. This complex tectonic situation raises questions on the

150   homogeneity of the parent population and on the treatment of the data as a whole. Therefore, we did

151   not include in the analysis any shallow events (H≤5 km) occurring in the Mount Etna area. So doing

we reduce the introduction of heterogeneities in the data set. The selected data set consists of 232 crustal seismic events (Figure 2) recorded between 2006 and 2014 by the stations of the "*Rete Sismica Permanente della Sicilia Orientale*" (RSPSO) operated by *Istituto Nazionale di Geofisica e Vulcanologia (INGV) Sezione di Catania– Osservatorio Etneo* (Fig. 1c). The seismic network is located in a region between the Hyblean Plateau and the volcanic archipelago of Aeloian Islands, and comprises about eighty digital stations equipped with Nanometrics Trillium broadband seismometers having an eigenperiod of 40 s. The signals are digitized at each station with 24-bit resolution at 100 samples/s.

The seismic events considered in the analyses have been extracted from the "Catalogo dei terremoti della Sicilia Orientale - Calabria Meridionale, INGV, Catania" (Gruppo Analisi Dati Sismici, 2016) within an area defined by the rectangle of latitude N $35.90 - 38.85$ and longitude E $13.45 - 16.85$, local magnitude (M_L) greater than 2.8 (maximum M_L=4.8; about 35% of the earthquakes have a magnitude $\geq$3.5) and focal depth up to 35 km (Fig. 2). We have chosen this interval of magnitude for the quality of data and homogeneity of the instrumental chain. The seismic events were located by using the Hypoellipse code (Lahr, 1989) in a seven layer, one-dimensional velocity structure (Hirn *et al*., 1991). A constant V_p/V_s ratio of 1.73 for travel time calculations was assumed. Figure 3a shows the distribution of records of our data set with respect to local magnitude and hypocentral distance. The starting data set has hypocentral distances from 6 to 377 km. However, the 99% of total data have been acquired at a distance less than about 200 km, therefore by considering a range up to this distance we obtain a spatial sampling rather homogeneous and dense. Looking at Figure 3a, we can expect that no bias is introduced, since no evident trend between magnitude and hypocentral distance exists. Finally, Figure 3b shows the signal-to-noise (S/N) ratio for the P phases estimated over windows of 2 s wide. It evidences that the 97% of windows have an S/N ratio greater than 5 dB, therefore suggesting a good quality of our data set.

177   As a first step in the processing, the data were checked visually at all stations to exclude

178   traces with electronic glitches, and with phenomena of amplitude saturation. The digitized velocity

179   time histories of the three components of ground motion were first baseline corrected, by removing

180   the offset and the linear trends, and thus instrument corrected. Therefore, we identified the P-phases

181   and their picks on the unfiltered vertical velocity components. As a final step, we integrated the

182   velocity records to obtain the displacement time series and, in order to remove the low frequencies

183   introduced by numerical integration, a high-pass Butterworth filter with corner frequency of 0.075

184   Hz was applied to the data.

185

186   Empirical correlation laws for eastern Sicily

187   *Peak ground displacement, magnitude and distance*

188   Nakamura (1984, 1988), with the EEW system known as UrEDAS, was the pioneer in

189   considering the first few seconds of recorded P-waves to estimate the magnitude of a a seismic

190   event. The method of Nakamura consists in computing continually in real-time the predominant

191   period from the first 2 to 4 s of P waves for estimating the magnitude of the event. An alternative

192   technique has been proposed by Wu and Zhao (2006), based on the use of the peak displacement

193   amplitude, P_d, measured on the three seconds window starting from the P-wave arrival time

194   picking. They investigated the relationship between P_d, the hypocentral distance and the local

195   magnitude in southern California and Taiwan . They found that P_d can be used to estimate the

196   magnitude of seismic events and can have a practical application in the EEW system.

197   Independently, Zollo *et al*. (2006, 2007) showed that the peak displacement amplitude of first few

198   seconds of P- and S-wave seismic signal scales with the earthquake magnitude for $4 \leq M_w \leq 7$; it can

199   be used for the estimation in real-time of the earthquake size in EEW applications. Two important

200   differences distinguish the approach by Zollo *et al*. (2006, 2007) from that by Wu and Zhao (2006):

201   (1) the time window is not fixed to three seconds and (2) the initial S-waves are taken into account,

202   as well. Indeed, when a dense seismic network is placed around the potential source area, the S-

203   phases data, which are available at the stations closest to the epicenter, can be used to improve the

204   magnitude estimation before the strong ground shaking reaches the distant target sites.

205   Following Lancieri and Zollo (2008), P- (P_dP) and S-waves (P_dS) peak displacement

206   amplitudes are measured on the modulus of displacement defined as:

207

208                         $$D = \sqrt{Z^2 + EW^2 + NS^2}$$ (1)

209

210   where Z, EW and NS are the vertical, east-west and north-south components of ground motion,

211   respectively. Unlike P-waves picking, the onset of S-waves has been estimated from travel time of

212   P waves by assuming a ratio $V_p/V_s = 1.73$ (V_p and V_s are the P- and S- wave velocities, respectively).

213   Therefore, we measure P_d on time windows of 2 s and 4 s of P-waves (denoted as *2P* and *4P*,

214   respectively) and 2 s for the S-wave (denoted as *2S*) starting from the P- and S-waves picked

215   arrivals.

216   The logarithmic P_d is generally assumed to be related to magnitude (*M*) and hypocentral

217   distance (*R*) through the standard attenuation expression (Wu and Zhao, 2006; Zollo *et al*., 2006;

218   Lancieri and Zollo, 2008):

219

220                         $$Log_{10}(P_d) = a + bM + cLog_{10}(R)$$ (2)

221

222   where *b* and *c* are the coefficients describing the magnitude dependence and the exponent in the

223   distance dependent amplitude decay (that is the geometrical attenuation, assumed constant in all the

224   investigated distance range), respectively. The model in equation (2) does not include the term

225   representing the anelastic attenuation, that is the linear R term. This is generally removed from the

226   model since it was found not to be statically significant (Wu and Zhao, 2006). Before testing this

227   assumption for our data, we need to make a number of points which emerge from the distribution of

228    $Log_{10}(P_d)$ as a function of hypocentral distance. In Figure 4 the $Log_{10}(P_d)$ measured on time

229   windows of 2 s (2P and 2S) and 4 s (4P) are plotted versus the hypocentral distance for three narrow

230   ranges of magnitude, 2.8-3.0, 3.6-3.8, and 4.6-4.8, respectively. So doing we reduce the scatter in P_d

231   amplitudes due to the different source sizes. At distances less than about 60 km, essentially

232   corresponding to attenuation of direct waves and where the effect of the anelastic attenuation should

233   be smaller, the $Log_{10}(P_d)$ values decay at a rate higher than those shown by $Log_{10}(P_d)$ values at

234   larger distances, for all three considered magnitude intervals. In fact, beyond about 60 km the rate

235   of decay of the $Log_{10}(P_d)$ is less severe due to the arrival of energy refracted and reflected from the

236   deeper parts of the crust. This means that if we consider a wide range of hypocentral distance, the

237   coefficient c in equation (2) cannot be supposed constant. Therefore, in order to estimate the

238   relationship between peak ground displacement, local magnitude, and distance, only records with a

239   maximum hypocentral distance of 60 km have been used. The number of records considered thus

240   drops to 3,928 three-component seismograms. This choice is also based on the general observation

241   that the crustal seismic events have high-frequency direct body waves with dominant amplitude at

242   distance from the receiver comparable with earthquake rupture length (Zeng *et al*., 1993). Figure 5

243   describes the distribution of the number of three component seismograms as a function of

244   magnitude that we used to perform the further analysis.

245   We investigate the attenuation due to geometrical spreading and anelastic attenuation

246   reformulating the equation of $Log_{10}(P_d)$ as follows

247

248                          $$Log_{10}(P_d) = a + bM + cLog_{10}(R) + dR \qquad (3)$$

249

250   where dR represents the anelastic attenuation. In Table 1, the regression coefficients, together with

251   their 95% confidence interval, for 2P, 4P, and 2S are reported for both attenuation models. Note that

252   P_d is in meters and R is in kilometers. The coefficient of elastic attenuation, d, was found not to be

253   statically significant, with a positive value very close to zero. Additionally, its introduction in the

254 attenuation model does not improve the fit of the data as evidenced by the root mean squared errors

255 and the coefficient of determination values (see Table 1 A and B for comparison). It was therefore

256 removed from the model.

257   A residual analysis has been performed in order to verify whether the regression models are

258 able to explain as much variation as possible in the dependent variable, assuming that the random

259 error is uniquely distributed over the data set. In particular, the regression analysis can be

260 considered successful in explaining the variation of the dependent variable if the residuals are

261 unstructured and small. Otherwise, the validity of regression is questionable since the residuals are

262 correlated to two independent variables. Figure 6 (a, b, and c) shows the residuals as a function of

263 the hypocentral distance and magnitude. It is clear from the figure that the residuals do not show

264 any significant trends both vs. magnitude and hypocentral distance. Additionally, more than about

265 90% of residuals is in the range -0.5 ÷ 0.5 for the three considered time windows.

266   The equation (2) has been used to correct for the distance effects the observed peak

267 amplitudes, by normalizing them to a reference distance of 30 km, a value chosen since

268 approximates the mean epicentral distance for the considered data set. As shown in Figure 7, there

269 is an evident positive correlation between the logarithmic peak displacement normalized to 30 km

270 [$Log_{10}(P_d)^{30km}$)] and the local magnitude for 2P, 4P, and 2S time windows in the whole investigated

271 magnitude range.

272   For each magnitude value, we first calculated the mean and the standard deviation of

273 $Log_{10}(P_d)^{30km}$ for both P- and S-waves. Therefore, a linear regression line was computed for the

274 means of $Log_{10}(P_d)^{30km}$ weighted by the inverse of standard deviation (σ) as (Zollo *et al*., 2006):

275

276 $$Log_{10}\left(P_d\right)^{30km} = a' + b'M$$ (4)

277

278 The means of $Log_{10}(P_d)^{30km}$ are shown in the Figure 7 (black dots), whilst the estimated coefficients

279 a' and b' are listed in Table 2, together with the calculated weighted standard errors (*WSE*) that has

280 been computed as:

281

282
$$WSE = \sqrt{\frac{\sum_i w_i \left[Log_{10}\left(P_{di}\right)^{30km} - \left(a'+b'M_i\right)\right]^2}{\sum_i w_i}}$$
(5)

283

284 where $w_i = 1/\sigma_i$ for the ith value of magnitude. The peak amplitudes are log linearly correlated with

285 magnitude in the considered magnitude range ($2.8 \leq M_L \leq 4.8$) for both P- and S-waves, with

286 correlation coefficients greater than 0.98 even for very short time windows from P-wave arrivals.

287

288 *τ_c and magnitude*

289 Nakamura (1988) was the first to develop a method for rapidly estimating the magnitude of

290 an earthquake for early warning purposes by using the frequency content of the first P-wave train.

291 Nakamura's approach is based on the computation of the predominant period for the first P-wave

292 train taking into account the vertical component of ground-motion. It has been widely applied to

293 data from both broad-band and strong-motion stations in several seismic regions, demonstrating that

294 the predominant period scales with seismic event magnitude (Allen and Kanamori, 2003; Olson and

295 Allen, 2005; Lockman and Allen, 2007), and up to a few hundreds of kilometers from the seismic

296 source is independent from the epicentral distance (Allen and Kanamori, 2003; Allen *et al.*, 2009a,

297 b).

298 An alternative method has been developed by Wu and Kanamori (2005a), based on the

299 computation of the characteristic period, τ_c, defined as:

300

$$\tau_c = 2\pi \sqrt{\frac{\int_0^{\tau_0} u^2(t)dt}{\int_0^{\tau_0} v^2(t)dt}}$$ (6)

301

302

303   where $u(t)$ and $v(t)$ are the ground-motion displacement and velocity, respectively. The time

304   window of integration starts at the P wave onset time and has a duration equal to τ_0, generally set a

305   3 s. τ_c is considered to represent the average period of P-wave signal and several studies have

306   shown that it reflects the sizes of earthquakes (Kanamori, 2005; Wu and Kanamori, 2005b).

307   Moreover, τ_c is less affected by the filter parameters and pre-event noise than the predominant

308   period for the first P-wave train, since it is estimated on the actual P-wave window (Shieh *et al.*,

309   2008).

310   For the estimation of τ_c, we considered the ground motion filtered (high-pass filtered at

311   0.075 Hz) displacement, $u(t)$, and velocity, $v(t)$, from the vertical component of ground motion. τ_0

312   has been set to 3 s.

313   In the model considered here, the parameter τ_c is depending only on the source

314   characteristics and not on the distance. In order to verify this assumption, we plot in Figure 8a the τ_c

315   as a function of the hypocentral distance. Looking at the figure, we observe that τ_c does not show

316   any significant trend with the distance, at least up to 60 km, as further confirmed by t-test with

317   significance threshold equal to 0.05.

318   As before, a linear regression was estimated for the means of $Log_{10}(\tau_c)$ computed for each

319   value of magnitude (Fig. 8b), weighted by the inverse of standard deviation. It has the following

320   equation:

321

322    $$Log_{10}(\tau_c) = 0.143(\pm 0.070)M_L - 0.853(\pm 0.266)$$ (7)

323

where τ_c is measured in seconds, suggesting that the average $Log_{10}(\tau_c)$ values increase with

increasing magnitude. The uncertainties associated to the two coefficients of the model are the 95%

confidence intervals. In Figure 9, we compare our relationship with those obtained by Zollo *et al*.

(2010) considering the data from south of Italy, Taiwan, and Japan, land Carranza *et al*. (2013) with

data from the south of the Iberian Peninsula, SE Iberia and north Africa. The regression through our

data yielded similar results to the findings of Zollo *et al*. (2010), even though the investigated

ranges of magnitude are different, whilst the relationship by Carranza *et al*. (2013) suggests a closer

dependence of the period parameter τ_c on the magnitude. These differences could be attributed to

the characteristics of the used dataset. In particular, Zollo *et al*. (2010) selected waveform records of

events essentially occurring at a depth less than 50 km and acquired at less than 60 km hypocentral

distance, as in our case. However, their magnitude range spanned 4 to 8.5. Conversely, Carranza *et*

al. (2013) considered seismic events with magnitude ranging from 3.8 to 5.9 but recorded at

regional distances (up to 700 km).

Peak ground velocity, PGV, versus P_d

Wu and Kanamori (2005a) showed that P_d is correlated with the peak ground-motion

velocity, *PGV*, at the same site, and when $P_d > 0.5$ cm, the event is most likely able to produce

damages. Therefore, in real-time, the measured P_d and τ_c can be used to calculate the level of

shaking (that is *PGV*) at the target sites, and *M*, respectively, even though *M* is not directly used for

onsite early warning purposes.

In Figure 10, the PGV values, measured as the maximum amplitude between the two

unfiltered horizontal components of ground motion velocity, are plotted as a function of peak

displacement P_d, measured from the low-pass filtered displacement records over a 3 s time window

after the P-wave pick. The figure shows that overall the PGV values increase logarithmically with

P_d in the investigated range of magnitude, in agreement with findings of several previous studies

349   (e.g., Wu *et al*., 2007; Zollo *et al*., 2010). Again, considering a maximum distance of 60 km, we

350   obtain the following best-fit regression line:

351

352    $$Log_{10}(PGV) = 1.36(\pm0.05) + 0.91(\pm0.02)Log_{10}(P_d) \tag{8}$$

353

354   where the units of *PGV* are cm/s and of P_d are cm. The standard deviation of $Log_{10}(PGV)$ is 0.27,

355   whilst the coefficient of determination is 0.80. In Figure 11, we compare our *PGV-P_d* relationship

356   against others calibrated for several areas worldwide by Wu *et al*. (2007), Zollo *et al*. (2010), and

357   Carranza *et al*. (2013). The comparison suggests that our data distribution is consistent with the

358   empirical regression lines obtained by these authors, independently of the considered maximum

359   distance and magnitude ranges.

360

361   Discussion

362   We have estimated empirical scaling relationships between the EEW parameters, P_d and τ_c,

363   and both magnitude and peak ground velocity (*PGV*), by using the broadband velocity seismograms

364   of earthquakes occurring in eastern Sicily. The data have been acquired by the stations of the

365   currently operating network in the area whose distribution ensures a good distance and azimuthal

366   coverage (Fig. 1c), making this case study a good test for the application of EEW methodologies.

367   The τ_c-M_L empirical scaling law estimated with our data distribution proved very similar

368   with that obtained by Zollo *et al*. (2010) by using strong motion data recorded in south of Italy,

369   Taiwan, and Japan in spite of the differences in the magnitude ranges covered by the data (see Fig.

370   9). The estimated *PGV* versus P_d relationship is nearly identical to those obtained from data of

371   various regions around the world (see Fig. 11). This suggests that, despite the scatter of the data

372   around the mean, the correlation between *PGV* and P_d is independent of effects such as source,

373   attenuation, site response or tectonic regime (Zollo *et al*., 2010; Carranza *et al*., 2013). The

15

374   uncertainty bounds associated to the regression lines take into account the potential differences due

375   to the regional context or earthquake mechanisms.

376   In order to understand how the estimated scaling laws work, we selected some events that

377   are not included in the dataset used to discover the predictive relationships, as test data. In

378   particular, we considered 20 events ranging from magnitude M_L 2.8 to 4.3 (see Table 3) and

379   compared the observed P_d, τ_c, and *PGV* values with the predicted ones. In Figure 12, we compare

380   the base-10 logarithms of observed P_d ($P_{d\text{-}obs}$) values for 2P, 4P, and 2S time windows with the

381   predicted P_d ($P_{d\text{-}pred}$) by equation (2). We note that there is a general decreasing trend of P_d values

382   with the hypocentral distance. This was to be expected because of the combined effect of the

383   geometrical spreading and anelastic attenuation on the amplitude peak ground motion. Figure 12

384   shows that the $P_{d\text{-}pred}$ accords well with the observed one, since the data points are well aligned

385   along the straight line with slope equal to 1.

386   The comparison between the average of the observed τ_c values and the predicted ones by

387   equation (7) for the 20 earthquakes belonging to the test dataset is shown in Figure 13, where the

388   range of one standard deviation is also reported. Taking into account the uncertainty of magnitude

389   estimation (on average ± 0.2 units), as well, we can see that the average of the observed τ_c values

390   are within the predictive uncertainty bounds (±1SVD).

391   Finally, we compare the predicted and observed *PGV* of the 20 test events in Figure 14,

392   where the 45° line is also shown. It can be seen that the data are distributed fairly close to the 45°

393   line, suggesting that the values determined by the equation (8) are reliable estimates of *PGV*. For

394   example, for the two earthquakes of M_L=3.5 (see Table 3), equation (8) predicts $Log_{10}(PGV)$ (with

395   PGV in cm/s) from -2.3 to -0.9 with a mean value of -1.8±0.26. Within the uncertainty, these values

396   are consistent with the observed $Log_{10}(PGV)$ from -2.5 to -0.9, with a mean of -1.8±0.31. Similarly,

397   for the events in Table 3 with M_L=4.1 the $Log_{10}(PGV)_{pred}$ estimates from -1.6 to -0.1 (mean equal to

398   -1.0±0.4) accord well with the $Log_{10}(PGV)_{obs}$ from -1.8 to -0.1 (mean equal to -0.9±0.4). These

399 results confirm the robustness of P_d as a predictor of *PGV* for regional earthquake monitoring

400 purposes and EEW operations in our region.

401   An example of application of the obtained results in terms of EEW in the studied area has

402 been performed by the software platform PRESTo in simulation mode. It is mainly based on the

403 RTLoc (Satriano *et al.*, 2008) and RTMag (Lancieri and Zollo, 2008) algorithms for the earthquake

404 location and magnitude estimation in real-time, respectively. As regards the code, we do not go into

405 details here and refer the reader to the original paper by Satriano *et al.* (2011). It is worth noting that

406 the module RTMag implemented inside PRESTo makes use of empirical correlation laws between

407 $Log_{10}(P_d)$ and magnitude, such as equation 3. Playing back the recorded traces into PRESTo we can

408 to follow up the accuracy of the earthquake parameter predictions through the estimated empirical

409 laws and, at the same time, we can evaluate the speed of convergence for both location and

410 magnitude estimates for a potential development of EEW system in the investigated area.

411   As a case study, we present a simulation of a M_L 4.6 (23 June 2011) earthquake, one of the

412 strongest events in our dataset. It occurred close to the village of Tortorici (see star in Fig. 2), at

413 about 45 km away from a key refinery in Milazzo. Looking at the plots in Figure 15, we can

414 quantitatively analyse the results of the computation and see how quickly the system produces

415 reliable and stable estimation of the event parameters on the basis of the information coming from

416 the actual seismic network. The first input parameters are the P-phase picks that are required by the

417 phase association algorithm and the location module. In particular, before declaring an event at least

418 3 picks within 5 s must be acquired by the system. This condition is reached 5.09 s after the event

419 origin time (Fig. 15b), when the stations MUCR, MSFR, and ECAN trigger. 0.2 s later, the first

420 location is available with a difference of about 8 km with respect to the reference epicentral location

421 and focal depth (namely the one from the seismic catalog). At the same time, the first magnitude

422 estimate is 5.0 with an uncertainty (Fig. 15a), defined as the confidence interval between 5% and

423 95%, from 3.6 to 7.5. This estimation is associated to the 2P window available at the nearest station

424 (MUCR, at about 8 km from the epicenter). At 6.2 s from the origin, another 2P window is available

425 (at station MSFR) and the estimated magnitude reaches 4.0 with smaller error (from 3.7 to 5.2,

426 range that still contains the true value). As more P_d measurements are acquired by the system, the

427 magnitude estimates through the empirical regression model is more and more refined and the

428 uncertainty decreases. In fact, at about 8 s (15 P picks) from the origin time the magnitude settles at

429 4.6, with an error equal to ± 0.4 units of magnitude, which is reasonable for an early warning

430 application. Therefore, with the actual network configuration, both location and magnitude

431 estimations are fairly stable and reliable after about 8 s from the origin time, when the system gives

432 awaited lead times equal to 7 s at Milazzo, 13 s at Catania and 14 s at Messina. This means that

433 when the event is declared the expected lead times are about 3 s longer.

434   In the platform PRESTo the on-site approach to EEW is addressed through the threshold-

435 based method by Zollo *et al*. (2010) for the independent definition of alert levels at each recording

436 sites. This approach has the advantage that the potential damaging effects of the earthquake are

437 evaluated without requiring accurate real-time location of the event. This is particularly useful when

438 the distribution of the stations is very sparse and does not guarantee an optimal azimuthal coverage

439 to obtain an early and reliable location of the epicenter. In the threshold approach, the parameters P_d

440 and τ_c, measured in a 3 s time window after the first P-arrival time at each station, are compared

441 with a priori selected threshold values that define four alert levels (from 0 to 3, see Zollo *et al*.,

442 2010 for details) inside a decisional table. These levels are connected to both the expected on-site

443 damages and the damages at distance. Considering a threshold value of macroseismic intensity I_{MCS}

444 (Mercalli-Cancani-Sieberg Intensity Scale) for damage effects equal to VII, we can estimate the

445 PGV expected by using the regression relationships by Faenza and Michelini (2010). Note that the

446 MCS Intensity scale is used in Italy to describe the effects of the earthquake ground shaking on the

447 built environment and communities. For I_{MCS}=VII the Faenza and Michelini (2010) empirical law

448 predicts $PGV = 6$ cm/s, that we can convert into P_d threshold value by using the equation (8) and

449 taking 1 standard deviation, obtaining $P_d = 0.1$ cm. The τ_c threshold can be estimated through the

450 equation (7) for a minimum magnitude value fixed as threshold. On the basis of seismic history of

the region, we selected M_L 5, estimating a $\tau_c = 0.3$ s obtained, again considering a 1 standard

deviation. Figure 16 shows the time evolution of P_d and τ_c measurements for the M_L 4.6 seismic

event we considered above, and the alert levels that have been associated to the recording sites. The

first measurements of P_d and τ_c are available after 6.1 s from the origin time, or after 3.1 s from the

first P-pick at station MUCR which is the nearest station to the epicenter. According to P_d and τ_c

values, the system associates an alert level 3 (damage expected nearby and far from stations, Zollo

et al. (2010)) since at MUCR both measurements are higher than the threshold values. The values of

τ_c remain stable with time (or with distance) (Fig. 16a) and consistent from one station to another.

Conversely, P_d shows a decreasing trend with time or distance, as expected, and the values rapidly

drop under the threshold value (alert level 1) (Fig. 16c). In this case study, only at station MUCR

the threshold values are reached and surpassed. On the contrary, if there are a certain number of

near source stations where P_d and τ_c exceed the threshold values, the real-time mapping of alert

levels can be used to predict the Potential Damage Zone (PDZ) (Zollo *et al.*, 2010). This is

particularly important to guarantee an efficient planning of rescue operations during emergency

phases immediately after an earthquake.

Concluding remarks

Today, the development of earthquake early warning systems represents one of the most

useful strategy to mitigate seismic risk in short time-scales and many countries worldwide are

promoting and developing such systems. In the frame of seismic risk management, it is considered a

reasonable costly solution for the loss reduction. Additionally, the developments of the real time

seismology are opening new scenarios in the framework of interaction of EEW and earthquake

engineering applications (i.e., Fujita *et al.*, 2011; Kubo *et al.*, 2011; Maddaloni *et al.*, 2011;

Nakamura *et al.*, 2011).

475   The great number of data we used in this study, acquired by the current seismic network

476   deployed in eastern Sicily, ensured an appropriate sample size for the robustness and accuracy of

477   the empirical laws we estimated. This is an important aspect since the reliability of the predicted

478   ground shaking depends, first of all, on the accuracy of the attenuation law applied to estimate it.

479   Moreover, with its present configuration, the network can be useful for evaluating how to develop

480   the seismic early warning system and select the most appropriate approach (regional, on-site, or

481   mixed) for the area. For instance, it is dense enough to allow the PRESTo platform to converge to a

482   stable estimation of the location and magnitude 2 to 3 s from the event declaration.

483   Evaluating the practicability of an EEW system in this area is justified by its high level of

484   seismic hazard. Moreover, three big oil refineries and power plants are installed along the eastern

485   cost of Sicily in the cities of Milazzo, Augusta and Priolo Gargallo, which have as main activity the

486   refining of crude oil and its derivatives. For them suitable safety measures such as, for example, the

487   automatic blocking of pipelines or gas lines in order to prevent fire hazards or the automatic

488   shutdown of the manufacturing operations to avoid the equipment casualty, could be adopted for the

489   damage reduction. Therefore, the introduction of an EEW system within the practices of frame of

490   real-time seismology that has been regularly carried out for years in eastern Sicily is a worthwhile

491   objective, since it can be effectively used to reduce damage caused by strongest earthquakes.

492   Taking into account the distribution of the major earthquakes in eastern Sicily, it could be

493   expected that in many cases the regional approach will not give enough time to process the data and

494   divulgate the alarm. Some problems may arise, above all, for the events occurring offshore. For

495   them, more coastal stations might be needed to better constrain the earthquake location. However,

496   for these events the on-site threshold-based EEWs approach can issue an alert rapidly to the in-land

497   target sites and estimate a potential damage zone within very few seconds (2-3 s) from the origin of

498   the seismic event, increasing the lead-time and reducing the blind zone.

499   In conclusion, hopefully it will be the development in eastern Sicily of a seismic network

500   that includes the real-time processing of the seismic recordings since it can be used also as a tool to

501   predict in real-time the ground motion measure and allow emergency response to be carried out

502   quickly.

503

504   Data and Resources

505   Seismic recordings used in this work are not accessible to the public. The Web site for the software

506   PRESTo is www.prestoews.org (last accessed March 2015).

507

519

520   References

521   Allen, R. M. (2007). The ElarmS earthquake early warning methodology and application

522   across California, In *Earthquake Early Warning System*, P. Gasparini, G. Manfredi, J. Zschau (Eds),

523   Springer, p. 21-44.

524   Allen, R. M., and H. Kanamori (2003). The potential for earthquake early warning in

525   Southern California, *Science*, **300**, 786-789.

526   Allen, R. M., H. Brown, M. Hellweg, O. Khainovski, P. Lombard, and D. Neuhauser

527   (2009a). Real-time earthquake detection and hazard assessment by ElarmS across California,

528   *Geophys. Res. Lett*., **36**, L00B08, doi: 10.1029/2008GL036766.

529   Allen, R. M., P. Gasparini, O. Kamigaichi, and M. Böse (2009b). The status of earthquake

530   early warning around the world: an introductory overview, *Seism. Res. Lett*., **80**, doi:

531   10.1785/gssrl.80.5.682.

532   Amato, A., R. Azzara, A. Basili, C. Chiarabba, M. Cocco, M. Di Bona, and G. Selvaggi

533   (1995). Main shock and aftershocks of the December 13, 1990 Eastern Sicily earthquake. *Ann.*

534   *Geophys*., **38**, 255-266.

535   Azzaro, R., S. D'Amico, A. Mostaccio, L. Scarfì, and T. Tuvè (2006). Terremoti con effetti

536   macrosismici in Sicilia orientale nel periodo gennaio 2002-dicembre 2005. *Quad. Geofis*. **41**,

537   INGV, Roma (in Italian).

538   Boschi, E., G. Ferrari, P. Gasperini, E. Guidoboni, G. Smriglio, and G. Valensise (1995),

539   Catalogo Dei Forti Terremoti in Italia Dal 461 A. C. Al 1980, Ist. Naz. di Geofis., Rome.

540   Böse, M., C. Ionescu, and F. Wenzel (2007). Earthquake early warning for Bucharest,

541   Romania: novel and revised scaling relations, *Geophys. Res. Lett*., **34**, L07302, doi:

542   10.1029/2007GL029396.

543   Carranza, M., E. Bufon, S. Colombelli, and A. Zollo (2013). Earthquake early warning for

544   southern Iberia: a P wave threshold-based approach, *Geophys. Res. Lett.*, **40**, 4588-4593.

545   Castello B., G. Selvaggi, C. Chiarabba, and A. Amato (2006). CSI Catalogo della sismicità

546   italiana 1981-2002, versione 1.1. INGV-CNT, Roma. http://www.ingv.it/CSI/

547   Colombelli, S., A. Zollo, G. Festa, and H. Kanamori (2012). Early magnitude and potential

548   damage zone estimates for the great M_w 9 Tohoku-Oki earthquake, *Geophys. Res. Lett*., **39**,

549   L22306, doi: 10.1029/2012GL053923.

550   Espinosa-Aranda, J. M., A. Jimenez, G. Ibarrola, F. Alcantar, and A. Aguilar (1995).

551   Mexico City seismic alert system, *Seism. Res. Lett*., **66**, 42-53.

552   Faccenna, C., T. W. Becker, F. P. Lucente, L. Jolivet, and F. Rossetti (2001). History of

553   subduction and back-arc extension in the Central Mediterranean, *Geophys. J. Int.*, **145**, 809-820.

554   Faenza, L., and A. Michelini (2010). Regression analysis of MCS intensity and ground

555   motion parameters in Italy and its application in ShakeMap, *Geophys. J. Int*., **180**, 1138-1152.

556   Gueguen, E., E. Tavarnelli, P. Renda, and M. Tramutoli (2002). The geodynamics of the

557   southern Tyrrhenian Sea margin as revealed by integrated geological, geophysical and geodetic

558   data. *Boll. Soc. Geol. It.,***1**, 77-85.

559   Heaton, T. H. (1985). A model for a seismic computerized alert network, *Science*, **229**, 987-

560   990.

561   Fujita, S., K. Minagawa, H. Shimosaka, and G. Tanaka (2011). Intelligent seismic isolation

562   system using air bearings and earthquake early warning, *Soil Dyn. Earth. Eng*., **31**, 223-230.

563   Hirn, A., A. Nercessian, M. Sapin, F. Ferrucci, G. Wittlinger (1991). Seismic heterogeneity

564   of Mt. Etna: structure and activity, *Geophys. J. Int*., **105**, 139–153.

565   Kanamori, H. (2005). Real time seismology and earthquake damage mitigation, *Ann. Rev.*

566   *Earth Planet Sci.*, **33**, 195-214.

567   Kubo, T., Y. Hisada, M. Murakami, F. Kosuge, and K. Hamano (2011). Application of an

568   earthquake early warning system and a real-time strong motion monitoring system in emergency

569   response in a high-rise building, *Soil Dyn. Earth. Eng*., **31**, 231-239.

570   Lahr, J. C. (1989). HYPOELLIPSE/VERSION 2.0: A computer program for determining

571   local earthquake hypocentral parameters, magnitude, and first motion pattern, *U.S. Geol. Surv.*

572   *Open File Rep., 89/116*, 81 pp.

573   Lancieri, M., and A. Zollo (2008). A bayesian approach to the real-time estimation of

574   magnitude from the early P and S wave displacement peaks, *J. Geophys. Res*., **113**, B12302, doi:

575   10.1029/2007JB005386.

576   Lockman, A., and R. M. Allen (2005). Single station earthquake characterization for early

577   warning, *Bull. Seism. Soc. Am*., **95**, 2029-2039.

578   Lockman, A., and R. M. Allen (2007). Magnitude-period scaling ralations for Japan and the

579   Pacific Northwest: implications for earthquake early warinig, *Bull. Seism. Soc. Am.*, **97**, 140-150.

580   Maddaloni, G., N. Caterino, and A. Occhiuzzi (2011). Semi-active control of the benchmark

581   highway bridge based on seismic early warning systems, *Bull. Earth. Eng.*, **9**, 1703-1715.

582   Musumeci, C., L. Scarfì, M. Palano, and D. Patanè (2014). Foreland segmentation along an

583   active convergent margin: New constraints in southeastern Sicily (Italy) from seismic and geodetic

584   observations, *Tectonophysics*, **630**, 137-149, doi: 10.1016/j.tecto.2014.05.017.

585   Nakamura, Y. (1984). Development of the earthquake early-warning system for the

586   Shinkansen, some recent earthquake engineering research and practical in Japan. The Japanese

587   National Committee of the International Association for Earthquake Engineering, 224-238.

588   Nakamura, Y. (1988). On the urgent earthquake detection and alarm system (UrEDAs),

589   *Proceedings of the 9th World Conference on Earthquake Engineering*, Tokyo-Kyoto, Japan.

590   Nakamura, Y., J. Saita, and T. Sata (2011). On a earthquake early warning system (EEW)

591   and its applications, *Soil Dyn. Earth. Eng.*, **31**, 127-136.

592   Neri, G., D. Caccamo, O. Cocina, and A. Montalto (1996). Geodynamic implications of

593   earthquake data in the Southern Tyrrhenian Sea. *Tectonophysics*, **258**, 233-249.

594   Patanè, D., O. Cocina, S. Falsaperla, E. Privitera and S. Spampinato (2004). Mt Etna

595   Volcano: A Seismological Framework. In: "Etna Volcano Laboratory", Bonaccorso A., Calvari S.,

596   Coltelli M., Del Negro C., Falsaperla S. (Eds), AGU (Geophysical monograph series), 147-165.

597   Olson, E., and R.M. Allen (2005). The deterministic nature of earthquake rupture, *Nature*,

598   **438**, 212-215.

599   Patanè, D. and E. Giampiccolo (2004). Faulting processes and earthquake source parameters

600   at Mount Etna: State of the Earth and perspectives, In: A. Bonaccorso, S. Calvari, M. Coltelli, C.

601   Del Negro and S. Falsaperla (eds.), Etna Volcano Laboratory, AGU Geophysical monograph series,

602   167-189.

603   Rovida, A., R. Camassi, P. Gasperini, and M. Stucchi (eds.) (2011). CPTI11, the 2011

604   version of the Parametric Catalogue of Italian Earthquakes. Istituto Nazionale di Geofisica e

605   Vulcanologia, Milano, Bologna. doi: http://doi.org/10.6092/INGV.IT-CPTI11.

606   Sato, S., K. Ashiya, S. Yamamoto, N. Iwata, M. Korenaga, and S. Noda (2011). Practical

607   use of the Earthquake Early Warning (EEW) system for Shinkansen, *Proceedings of the 9th World*

608   *Congress on Railway Research*, Lille, France.

609   Satriano, C., A. Lomax, and A. Zollo (2008). Real-time evolutionary earthquake location for

610   seismic early warning, *Bull. Seism. Soc. Am.*, **98**, 1482-1494, doi: 10.1785/01200601590.

611   Satriano, C., L. Elia, C. Martino, M. Lancieri, and A. Zollo (2011). PRESTo, the earthquake

612   early warning system for southern Italy: concepts, capabilities and future perspectives, *Soil Dyn.*

613   *Earth. Eng.*, **21**, 137-153.

614   Selvaggi, G. and C. Chiarabba (1995). Seismicity and P-wave velocity image of the

615   Southern Tyrrhenian subduction zone. *Geophys. J. Int*., **121**, 818-826, doi: 10.1111/j.1365-

616   246X.1995.tb06441.x

617   Shieh, J., Y.M. Wu, and R. Allen (2008). A comparison of τ_c and τ^p_{max} for magnitude

618   estimation in earthquake early warning, *Geophys. Res. Lett*., **35**, L20301, doi:

619   10.1029/2008GL035611.

620   Wieland, M., M. Gresser, and C. Kuendig (2000). Seismic early warning system for a

621   nuclear power plant, *Proceedings of the 12th World Conference on Earthquake Engineering*,

622   Auckland, New Zeland.

623   Wu, Y. M., and H. Kanamori (2005a). Rapid assessment of damage potential of earthquakes

624   in Taiwan from the beginning of P waves, *Bull. Seism. Soc. Am*., **95**, 1181-1185.

625   Wu, Y. M., and H. Kanamori (2005b). Experiment on an onsite early warning method for

626   the Taiwan early warning system, *Bull. Seism. Soc. Am*., **95**, 347-353.

627   Wu, Y. M., and L. Zhao (2006). Magnitude estimation using the first three seconds P-wave

628   amplitude in earthquake early warning, *Geophys. Res. Lett.*, **33**, L16312,

629   doi:10.1029/2006GL026871

630   Wu, Y. M., H. Y. Yen, L. Zhao, B. S. Huang, and W. T. Liang (2006). Magnitude

631   determination using initial P waves: a single-station approach, *Geophys. Res. Lett.*, **33**, L05306,

632   doi:10.1029/2005GL025395.

633   Wu, Y.M., H. Kanamori, R. M. Allen, and E. Hauhsson (2007). Determination of earthquake

634   early warning parameters, τ_c and P_d, for southern California, *Geophys. J. Int.*, **170**, 711-717, doi:

635   10.1111/j.1365-246X.2007.03430.x.

636   Wu, Y. M., and H. Kanamori (2008). Development of an earthquake early warning system

637   using real-time strong motion signals, *Sensors*, **8**, 1-9.

638   Zeng, Y., K. Aki, and T. Teng (1993). Mapping of the high-frequency source radiation for

639   the Loma Prieta Earthquake, California. *J. Geophys. Res.*, **98**, 11,981-11,993.

640   Zollo, A., M. Lancieri, S. Nielsen (2006). Earthquake magnitude estimation from peak

641   amplitudes of very early seismic signals on strong motion records, *Geophys. Res. Lett.*, **33**, L23312,

642   doi:10.1029/2006GL027795.

643   Zollo, A., M. Lancieri, S. Nielsen (2007). Reply to comment by P. Rydelek et al. on

644   "Earthquake magnitude estimation from peak amplitudes of very early seismic signals on strong

645   motion records", *Geophys. Res. Lett.*, **34**, L20303, doi:10.1029/2007GL030560.

646   Zollo, A., O. Amoroso, M. Lancieri, Y. M. Wu, and H. Kanamori (2010). A threshold-based

647   earthquake early warning using dense accelerometer networks, *Geophys. J. Int.*, **183**, 963-947.

648

649

650

651

652

653

654

Tables

Table 1 - Coefficients of equations 2 and 3 for the prediction of $Log_{10}(P_d)$.

	a	CI95 for a	b	CI95 for b	c	CI95 for c	d	CI95 for d	RMSE	R^2
Equation 2										
2P	-5.865	0.115	0.990	0.022	-1.915	0.068	-	-	0.3231	0.7140
4P	-5.904	0.239	1.007	0.026	-1.860	0.142	-	-	0.3151	0.7331
2S	-5.437	0.109	1.069	0.023	-2.016	0.062	-	-	0.3395	0.7331
Equation 3										
2P	-5.442	0.324	0.991	0.022	-2.327	0.303	0.0057	0.004	0.3228	0.7146
4P	-5.735	1.620	1.007	0.026	-2.004	1.378	0.0015	0.014	0.3152	0.7331
2S	-5.425	0.246	1.069	0.023	-2.028	0.233	0.0002	0.003	0.3396	0.7332

CI95 indicates the confidence intervals at 95% confidence level.

RMSE indicates the root mean squared errors.

R^2 is the coefficient of determination.

Table 2 - Coefficients of equation 3 for the prediction of $Log_{10}(P_d)^{30km.}$

	a'	CI95 for a'	b'	CI95 for b'	WSE	R^2
2P	-8.817	0.216	1.024	0.057	0.078	0.982
4P	-8.757	0.205	1.036	0.054	0.077	0.984
2S	-8.494	0.227	1.092	0.060	0.086	0.982

CI95 indicates the confidence intervals at 95% confidence level.

RMSE indicates the root mean squared errors.

R^2 is the coefficient of determination.

Table 3 - List of events used as test dataset.

Date (yyyy/mm/dd)	Time (hh:min:ss)	Latitude (°)	Longitude (°)	Depth (km)	M_L
2013/11/05	05.06.39.880	37.709	14.919	15.6	2.9
2014/01/01	01.59.50.610	36.644	14.954	8.9	3.0
2014/04/17	21.52.25.630	38.207	15.216	14.3	2.8
2014/06/27	02.56.47.770	37.803	14.590	18.6	3.1
2014/10/09	22.58.26.540	38.485	14.730	14.5	4.3
2014/10/10	16.27.12.920	38.056	15.105	33.1	2.9
2014/10/22	22.41.51.740	38.473	14.799	17.3	3.1
2015/01/05	07.27.02.830	37.167	15.289	20.8	3.1
2015/02/08	19.39.21.980	37.339	15.195	19.0	2.9
2015/02/11	03.57.00.110	38.034	14.754	9.3	3.1
2015/08/08	22.46.24.960	38.458	14.272	11.5	4.1
2015/09/20	22.27.59.120	37.170	15.507	21.9	4.1
2015/10/09	00.35.29.540	37.729	15.115	6.6	3.0
2015/10/10	21.37.44.980	37.839	14.879	22.9	2.8
2015/12/20	09.46.07.470	38.354	13.892	32.4	4.3
2015/12/22	05.35.09.180	37.773	15.431	21.8	3.5
2016/02/08	15.35.42.720	37.002	14.802	6.0	4.3
2016/02/08	17.57.37.320	37.002	14.806	6.0	3.6
2016/02/11	01.38.50.520	37.835	15.376	30.0	3.5
2016/03/01	16.47.50.470	38.490	14.613	12.4	3.1

663

664

665

666

Figure 1. (a, b) Simplified tectonic map (after Musumeci *et al*., 2014). (c) Geographic distribution of the permanent stations of the "*Rete Sismica Permanente della Sicilia Orientale*" (RSPSO).

Figure 2. Epicentral map of the 232 earthquakes selected for this study. The green star indicates the event used to perform the test with the code PRESTo (see text for details).

Figure 3. (a) Plot of local magnitude (M_L) vs. hypocentral distance for the whole data set shown in Figure 2; (b) histogram showing the signal-to-noise ratio (dB) of P-waves used for the analysis (~7000).

Figure 4. Plots of peak displacement amplitude versus hypocentral distance for three different range of magnitude.

Figure 5. Number of 3C seismograms vs. local magnitude for the data set used to solve equation 2 (see text for details).

Figure 6. Distribution of residuals of the $Log_{10}(P_d)$ as a function of the independent variables hypocentral distance and local magnitude for the three time windows (2P, 4P, and 2S).

Figure 7. Logarithm of P_d parameter normalized at a reference distance of 30 km versus the local magnitude (a and b) in the time windows 2s and 4s from the P-onset and (c) 2s from the S-onset (grey diamonds). Black dots indicate the mean of $Log_{10}(P_d)^{30km}$ for each M_L value plotted with its standard deviation. The solid and dashed lines indicate the best fit and $\pm 1WSE$ error bounds, respectively.

Figure 8. (a) The τ_c parameter versus hypocentral distance. (b) Correlation between the logarithm of period parameter τ_c value and local magnitude (grey diamonds). Black dots indicate the mean of $Log_{10}(\tau_c)$ for each local magnitude value plotted with its standard deviation. The best fit regression line (solid line), along with $\pm 1WSE$ error bounds (dashed lines), is shown. *R* is the correlation coefficient.

Figure 9. τ_c-magnitude relations obtained from this study, from Zollo *et al*. (2010), and Carranza *et al*. (2013). The grey area indicates the $\pm 1SDV$ error bounds.

Figure 10. Peak ground velocity (*PGV*) versus peak initial displacement amplitude (P_d) for hypocentral distances up to 60 km. Solid and dashed lines show the least squares fit and the range of one standard deviation, respectively. *R* is the correlation coefficient and *SDV* is the standard deviation.

Figure 11. Comparison of *PGV*-P_d relationship obtained in this study with those estimated by Wu *et al*. (2007) (data from Taiwan and Southern California), Zollo *et al*. (2010) (data from Italy, Japan, and Taiwan), and Carranza *et al*. (2013) (data from south of Iberian Peninsula).

Figure 12. (Left) Plots of observed values of $Log_{10}(P_d)$ versus the hypocentral distance. (Right) Plots of predicted values of $Log_{10}(P_d)$ versus the observed ones (the line 1:1 is also shown) from the three time windows (2P, 4P, and 2S) for the 20 selected events (see also Table 3) of the test dataset.

Figure 13. The observed τ_c parameter versus the local magnitude (M_L) for the 20 seismic events belonging to the test dataset. The predicted τ_c (black line) by equation (7) and $\pm 1SVD$ (black dashed lines) are also shown.

Figure 14. Predicted *PGV* (*PGV_pred*) versus observed *PGV* (*PGV_obs*) for the 20 selected events of test dataset (see text for details). The line 1:1 is also shown.

Figure 15. PRESTo timeline for the M_L 4.6, 23 June 2011 earthquake occurring near the village of Tortorici. The dashed line in the plots (a) and (f) represent the reference magnitude and focal depth, respectively, obtained from the seismic catalog.

Figure 16. Evolution in time of the ground parameters (a) τ_c, (b) P_d, and (c) the corresponding alert level at the different stations for the M_L 4.6, 23 June 2011 earthquake as estimated by the code PRESTo.
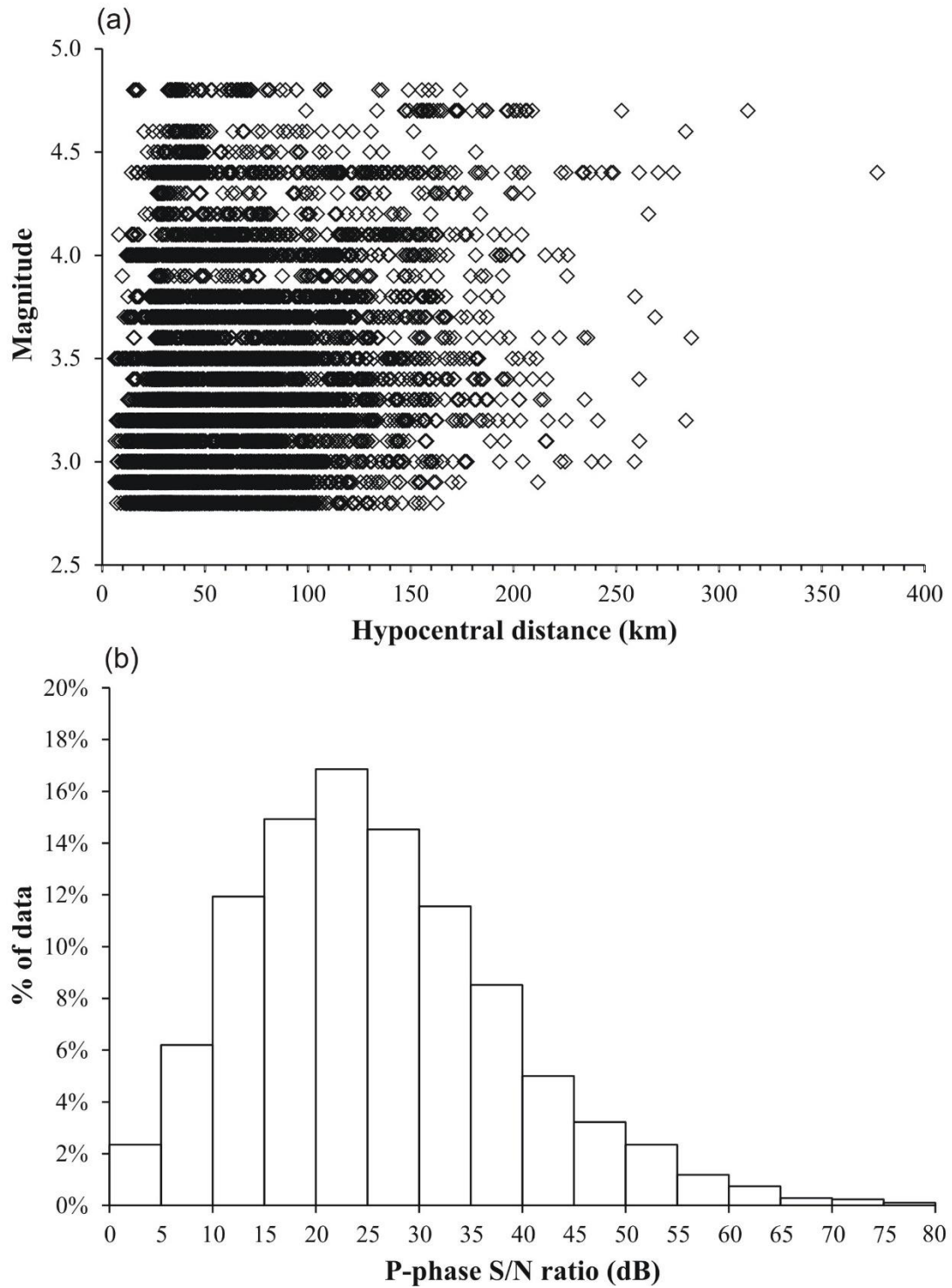
Figure 1. (a, b) Simplified tectonic map (after Musumeci *et al*., 2014). (c) Geographic distribution of the permanent stations of the "*Rete Sismica Permanente della Sicilia Orientale*" (RSPSO).
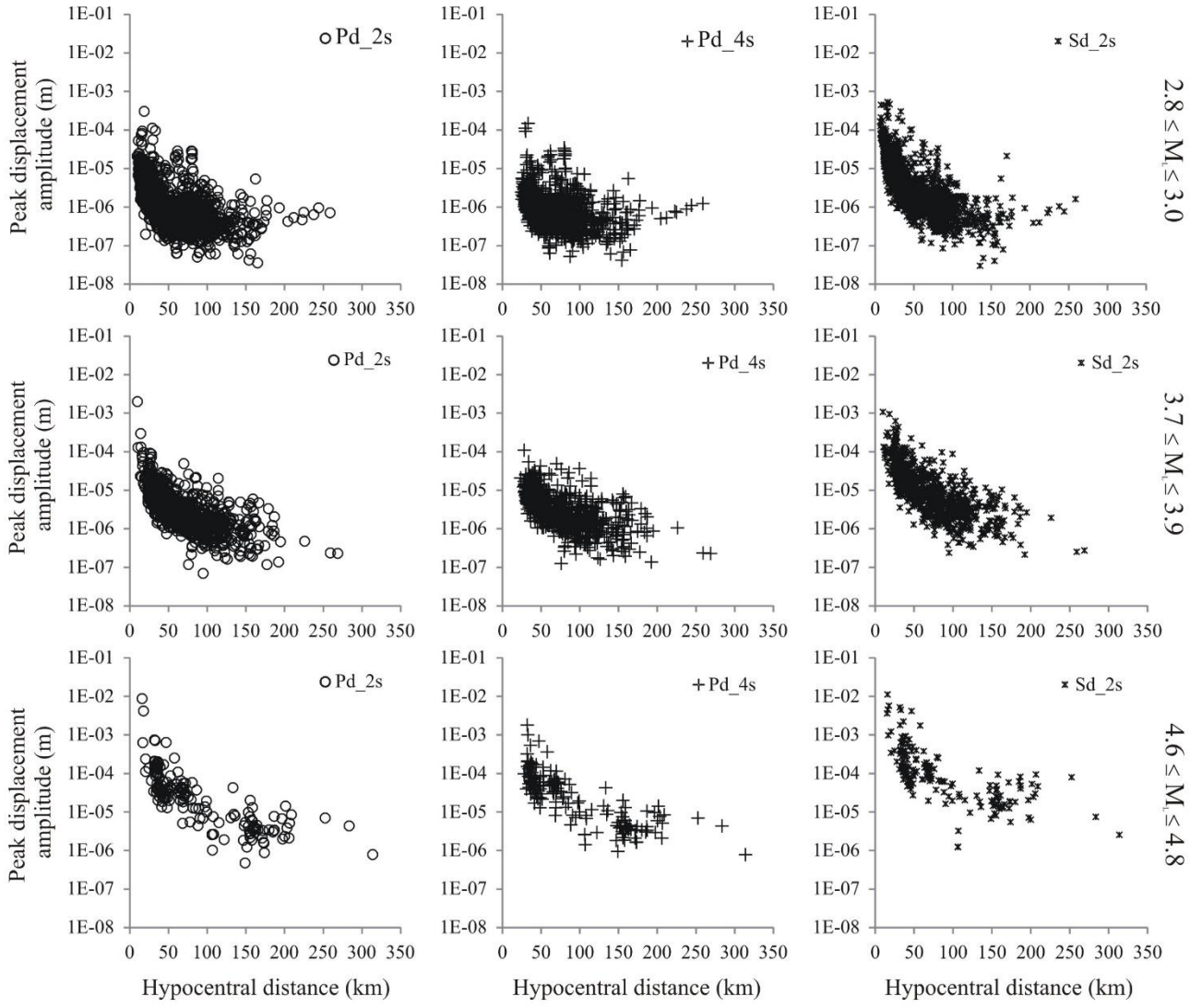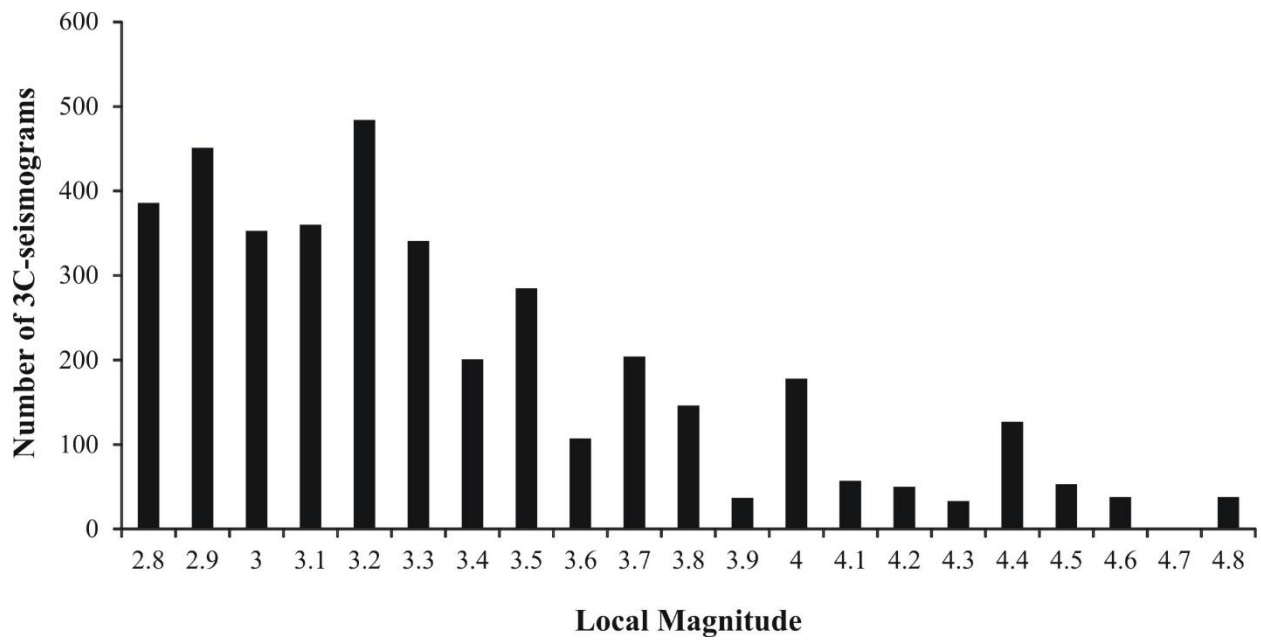
Figure 2. Epicentral map of the 232 earthquakes selected for this study. The green star indicates the event used to perform the test with the code PRESTo (see text for details).
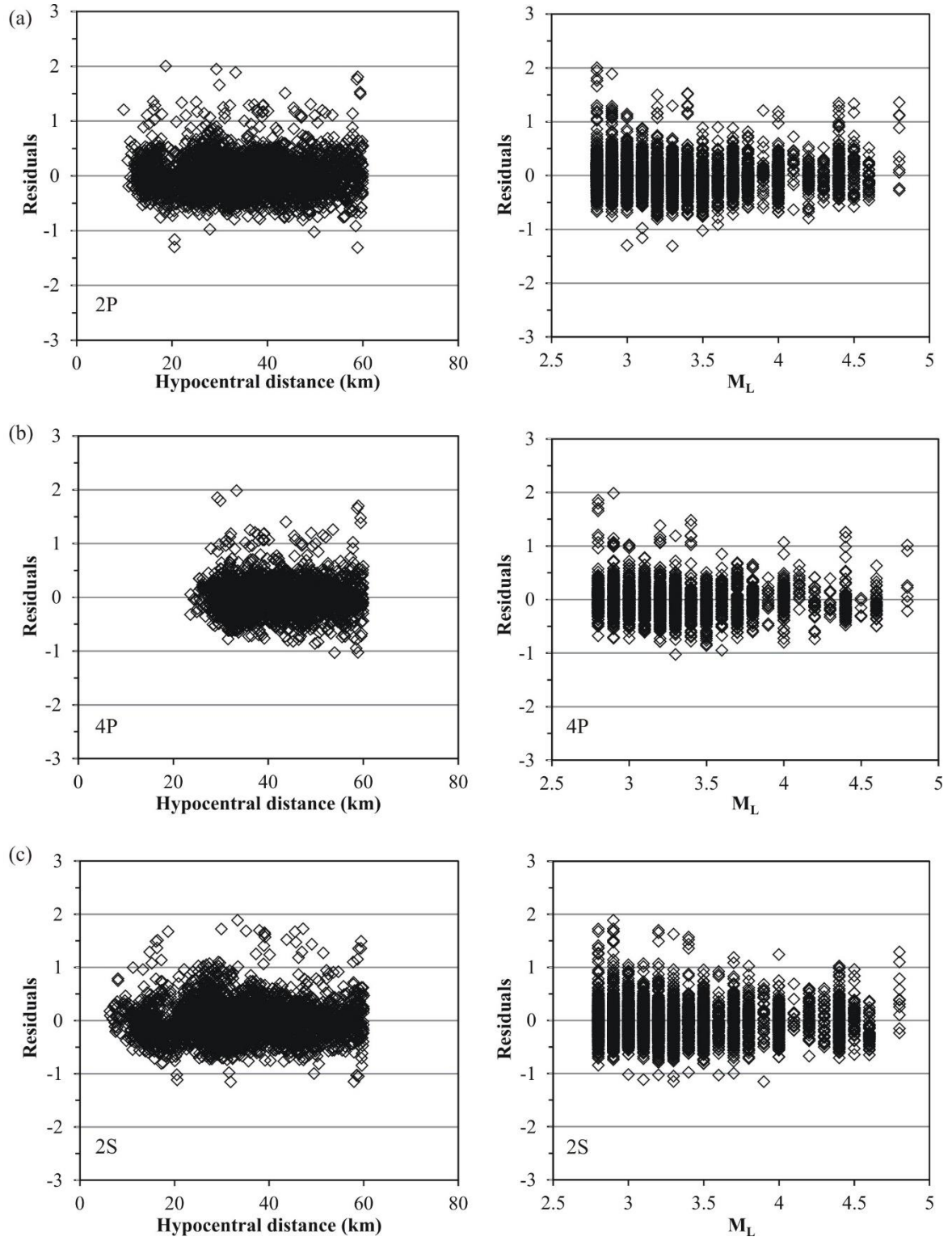
Figure 3. (a) Plot of local magnitude (M_L) vs. hypocentral distance for the whole data set shown in Figure 2; (b) histogram showing the signal-to-noise ratio (dB) of P-waves used for the analysis (~7000).

Figure 4. Plots of peak displacement amplitude versus hypocentral distance for three different range of magnitude.

Figure 5. Number of 3C seismograms vs. local magnitude for the data set used to solve equation 2 (see text for details).

Figure 6. Distribution of residuals of the $Log_{10}(P_d)$ as a function of the independent variables hypocentral distance and local magnitude for the three time windows (2P, 4P, and 2S).

Figure 7. Logarithm of P_d parameter normalized at a reference distance of 30 km versus the local magnitude (a and b) in the time windows 2s and 4s from the P-onset and (c) 2s from the S-onset (grey diamonds). Black dots indicate the mean of $Log_{10}(P_d)^{30km}$ for each M_L value plotted with its standard deviation. The solid and dashed lines represent the best fit and $\pm 1WSE$ error bounds, respectively.

(a)

(b)

$$Log_{10}(\tau_c)=0.143(\pm0.070)M_L-0.853(\pm0.266)$$
$$R=0.643$$
$$SDV=0.35$$

Figure 8. (a) The τ_c parameter versus hypocentral distance. (b) Correlation between the logarithm of period parameter τ_c value and local magnitude (grey diamonds). Black dots indicate the mean of $Log_{10}(\tau_c)$ for each local magnitude value plotted with its standard deviation. The best fit regression line (solid line), along with $\pm1WSE$ error bounds (dashed lines), is shown. R is the correlation coefficient.

Figure 9. τ_c-magnitude relations obtained from this study, from Zollo *et al*. (2010), and Carranza *et al*. (2013). The grey area indicates the *±1SDV* error bounds.
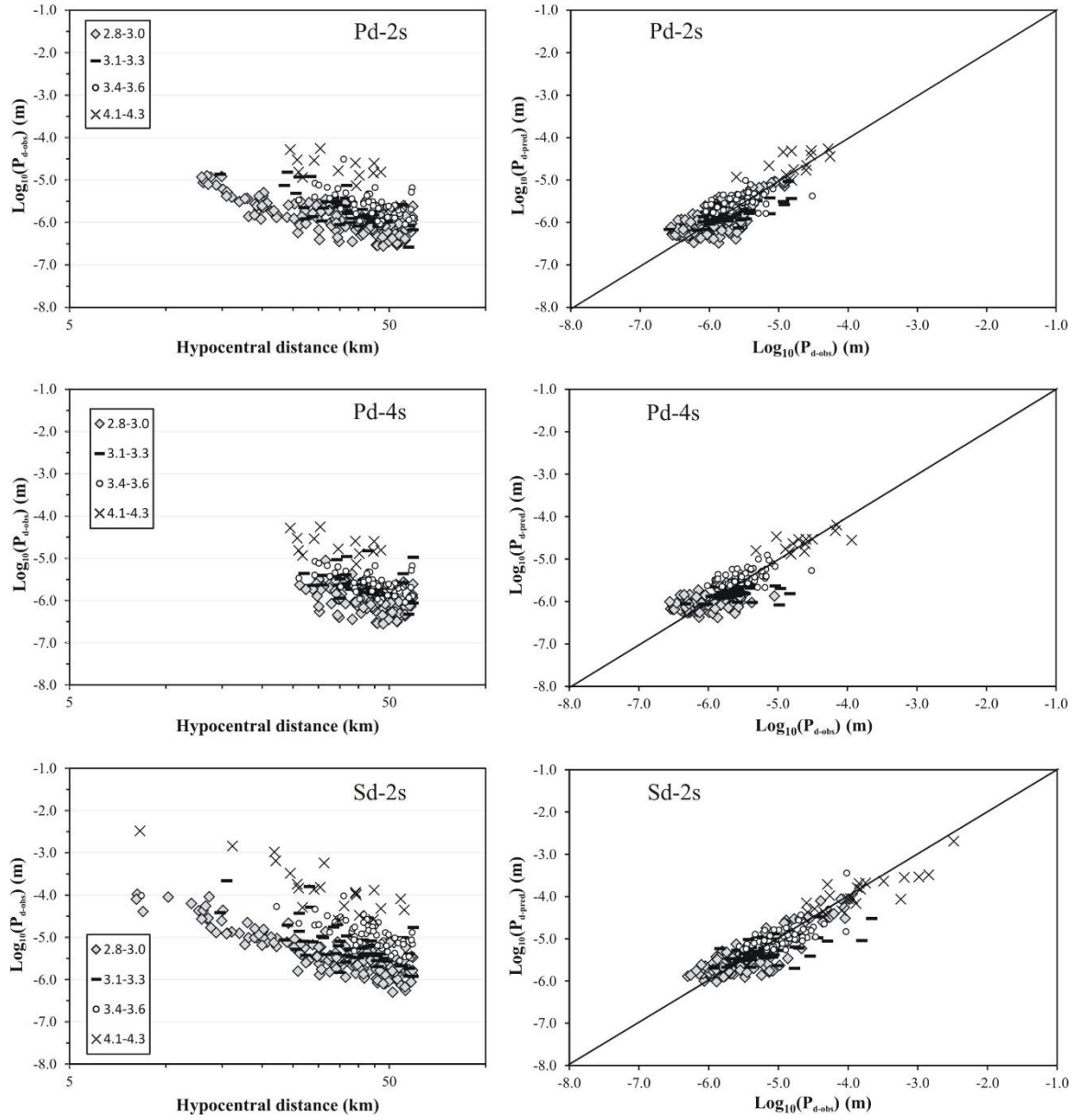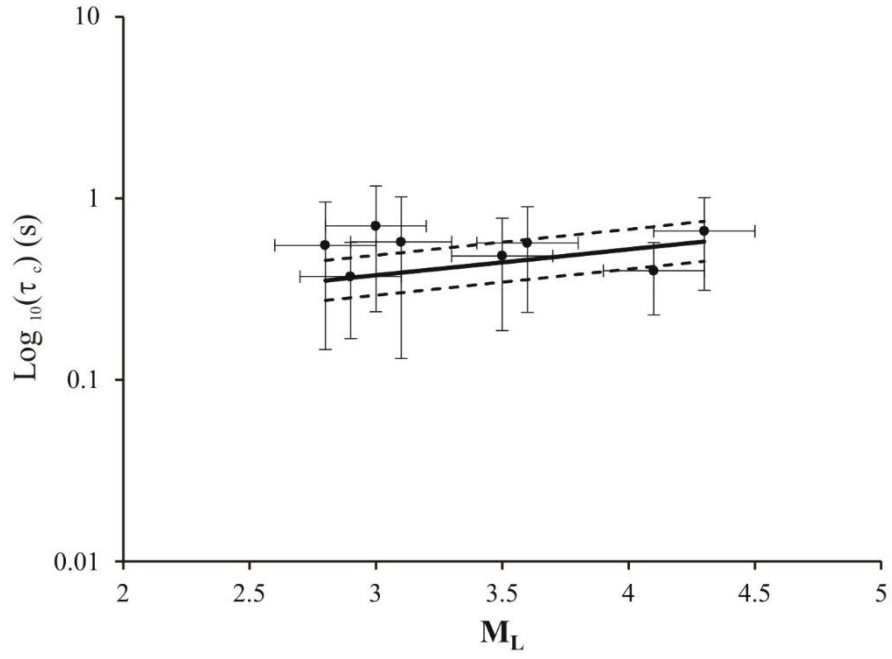
Figure 10. Peak ground velocity (*PGV*) versus peak initial displacement amplitude (P_d) for hypocentral distances up to 60 km. Solid and dashed lines show the least squares fit and the range of one standard deviation, respectively. *R* is the correlation coefficient and *SDV* is the standard deviation.
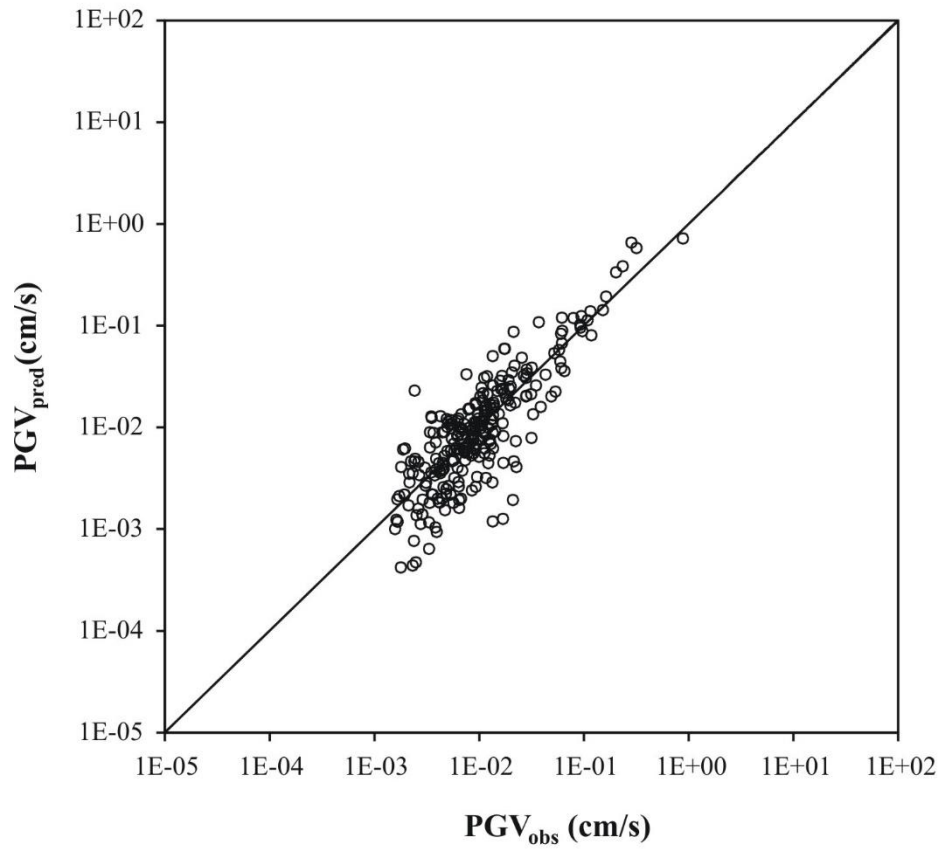
Figure 11. Comparison of *PGV-P$_d$* relationship obtained in this study with those estimated by Wu *et al*. (2007) (data from Taiwan and Southern California), Zollo *et al*. (2010) (data from Japan, Taiwan and Italy), and Carranza *et al*. (2013) (data from south of Iberian Peninsula).
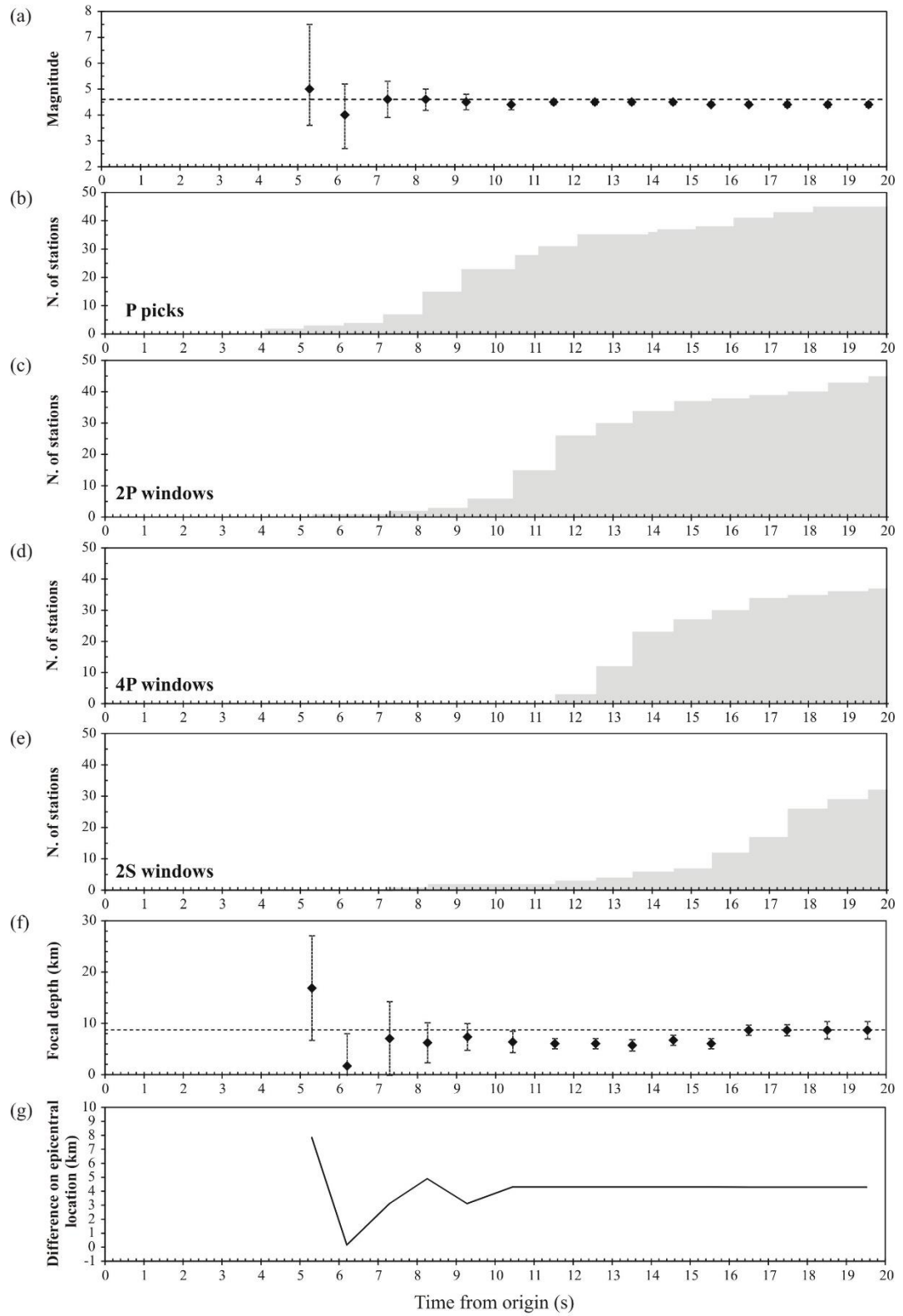
Figure 12. (Left) Plots of observed values of $Log_{10}(P_d)$ versus the hypocentral distance. (Right) Plots of predicted values of $Log_{10}(P_d)$ versus the observed ones (the line 1:1 is also shown) from the three time windows (2P, 4P, and 2S) for the 20 selected events (see also Table 3) of the test dataset.
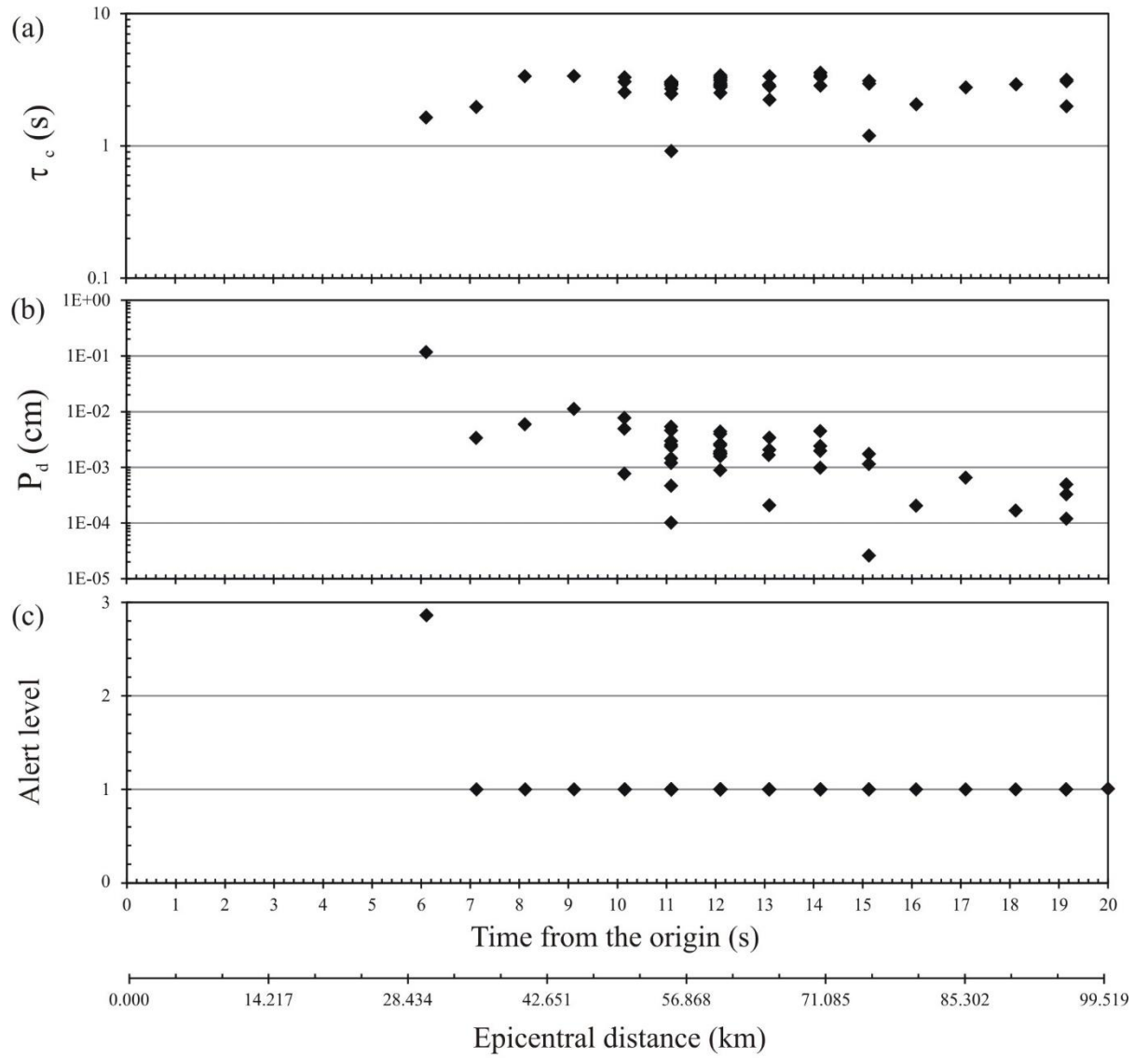
Figure 13. The observed τ_c parameter versus the local magnitude (M_L) for the 20 seismic events belonging to the test dataset. The predicted τ_c (black line) by equation (7) and $\pm 1SVD$ (black dashed lines) are also shown.

Figure 14. Predicted *PGV (PGV$_{pred}$)* versus observed *PGV (PGV$_{obs}$)* for the 20 selected events of test dataset (see text for details). The line 1:1 is also shown.

Figure 15. PRESTo timeline for the M_L 4.6, 23 June 2011 earthquake occurring near the village of Tortorici. The dashed line in the plots (a) and (f) represent the reference magnitude and focal depth, respectively, obtained from the seismic catalog.

Figure 16. Evolution in time of the ground parameters (a) τ_c, (b) P_d, and (c) the corresponding alert level at the different stations for the M_L 4.6, 23 June 2011 earthquake as estimated by the code PRESTo.

Bulletin of the Seismological Society of America

COPYRIGHT/PUBLICATION-CHARGES FORM

PLEASE FILL OUT AND SUBMIT THIS FORM ONLINE WHEN SUBMITTING YOUR PAPER
OR FAX IT TO FAX NUMBER 503 405 7190

Manuscript Number: BSSA-D-_____ [leave blank for new submissions]

Title: _ESTIMATION OF EARTHQUAKE EARLY WARNING PARAMETERS FOR EASTERN Sicily_

Authors: _GIUSEPPINA TUSA, CARLA TUSUMECI, DOMENICO PATANE_

COPYRIGHT

In accordance with Public Law 94-533, copyright to the article listed above is hereby transferred to the Seismological Society of America (for U.S. Government employees, to the extent transferable) effective if and when the article is accepted for publication in the *Bulletin of the Seismological Society of America*. The authors reserve the right to use all or part of the article in future works of their own. In addition, the authors affirm that the article has not been copyrighted and that it is not being submitted for publication elsewhere.

To be signed by at least one of the authors (who agrees to inform the others, if any) or, in the case of "work made for hire," by the employer.

_____   _GIUSEPPINA TUSA_   _July 19, 2016_
Authorized Signature for Copyright   Print Name (and title, if not author)   Date

PUBLICATION CHARGES

The Seismological Society of America requests that institutions supporting research share in the cost of publicizing the results of that research. The Editor has the discretion of waiving publication charges for authors who do not have institutional support. In addition to regular publication charges there is a nominal fee for publishing electronic supplements. Current rates are available at http://www.seismosoc.org/publications/journal-publication-charges/.

Color options: Color figures can be published (1) in color both in the online journal and in the printed journal, or (2) in color online and gray scale in print. Online color is free; authors will be charged for color in print. You must choose one option for all of the color figures within a paper; that is, you cannot choose option (1) for one color figure and option (2) for another color figure. You cannot submit two versions of the same figure, one for color and one for gray scale. You are responsible for ensuring that color figures are understandable when converted to gray scale, and that text references and captions are appropriate for both online and print versions. Color figures must be submitted before the paper is accepted for publication.
    Art guidelines are at http://www.seismosoc.org/publications/bssa/bssa-art-submission-guidelines/

Will publication charges be paid? Check one:

_____ **BOTH PUBLICATION CHARGES AND COLOR CHARGES WILL BE PAID**, and all color figures for this
        paper will be color both online and in print. This option requires full payment of publication & color charges.

__X__ **ONLY PUBLICATION CHARGES WILL BE PAID**, and all figures for this paper will be gray scale in print.
        Color figures, if any, will be color online.

_____ **REQUEST A REDUCTION IN PUBLICATION CHARGES.** Send a letter of request and explanation to the
        Editor-in-Chief at BSSA@seismosoc.org. Color figures, if any, will be color online but grey scale in print.

Send Invoice to: _ISTITUTO NAZIONALE DI GEOFISICA E VULCANOLOGIA._
SEZIONE DI CATANIA, PIAZZA ROMA 2, 95125, CATANIA, ITALY

If your paper is accepted for publication, SSA requires that you fill out and submit an online billing/offprint form.

Questions regarding billing should be directed to the SSA Business Office,
400 Evelyn Avenue, Suite 201 Albany, CA 94706 USA Phone 510 525-5474 Fax 510 525-7204

Rev. 2016-06-01