

Near real-time automatic moment magnitude estimation

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Abstract In this paper we describe a stable automatic method to estimate in real time the seismic moment, moment magnitude and corner frequency of events recorded by a network comprising broad-band and accelerometer sensors. The procedure produces reliable results even for small-magnitude events $M_W \approx 3$. The real-time data arise from both the Trans-frontier network at the Alps-Dinarides junction and from the Italian National Accelerometric Network (RAN). The data is pre-processed and the S-wave train identified through the application of an automatic method, which estimates the arrival times based on the hypocenter location, recording site and regional velocity model. The transverse component of motion is used to minimize conversion effects. The source spectrum is obtained by correcting the signals for geometrical spreading and intrinsic attenuation. Source spectra for both velocity and displacement are computed and, following Andrews (1986), the seismic moment and the first estimate of the corner frequency, f_0 , derived. The procedure is validated using the recordings of some recent moderate earthquakes (Carnia 2002; Bovéc 2004; Parma 2008; Aquila 2009; Macerata 2009; Emilia 2012) and the recordings of some minor events in the SE Alps area for which independent seismic moment and moment magnitude estimates are available. The results obtained with a dataset of 843 events recorded by the Transfrontier and RAN networks show that the procedure is reliable and robust for events with $M_W \geq 3$. The estimates of f_0 are less reliable. The results show a scatter, principally for small events with $M_W \leq 3$, probably due to site effects and inaccurate locations.

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1 Introduction

Several magnitude scales have been introduced to characterize seismic events. This is the case even for local or regional earthquakes. [Richter \(1935\)](#) introduced the local magnitude, M_L , to quantify medium-size earthquakes in Southern California. This magnitude was computed from the maximum displacement amplitude of recordings obtained by a Wood-Anderson torsion seismometer. Later on, other local magnitudes have been proposed (e.g. [Gutenberg and Richter 1942, 1956](#); [Kanamori and Jennings 1978](#); [Eaton 1992](#)), their measurements based on the S-wave amplitude, duration or coda characteristics. For teleseismic events more magnitudes have been introduced (for an overview see [Kanamori 1983](#)), the most known ones being M_S and m_b . These magnitudes are not a physical source-related quantity and all of them saturate at a certain magnitude value. [Kanamori \(1977\)](#) introduced the moment magnitude to overcome the magnitude saturation. This magnitude scale is linked to the physical quantity describing the seismic source, the seismic moment, introduced a decade before by [Aki \(1966\)](#) and computed from the flat low-frequency portion of the source spectrum ([Brune 1970, 1971](#)). For medium-sized earthquakes, the moment magnitude scale should be similar to the Richter value; however this has the side effect that the scales diverge for smaller earthquakes ([Hanks and Kanamori 1979](#)).

Many studies about seismic moment and moment magnitude estimates for local and regional events have been performed. This paper will not give a complete review, but only a few interesting and relatively recent approaches will be pointed out. [Mayeda \(1993](#); [Mayeda and Walter 1996](#); [Mayeda et al. 2003](#)) developed a completely empirical calibration method for obtaining seismic source moment-rate spectra derived from local and regional coda envelopes using broadband stations. [Lancieri and Zollo \(2008\)](#) estimate magnitude from measures of peak displacement. [Ottemoller and Havskov \(2003\)](#) use an automatic procedure to estimate moment magnitudes from displacement spectra of local and regional earthquakes. They use two inversion methods: a converging search and a standard genetic algorithm. [Edward et al. \(2010\)](#) compute moment magnitude using a spectral fitting method.

In this paper a new procedure is described that estimates moment magnitude using spectral analysis of S-waves, following [Andrews \(1986\)](#). This method does not depend on empirical relationships, but physical quantities from which the moment magnitude can be directly estimated.

We have tested the procedure on broad-band and strong-motion waveforms of recent earthquakes that have occurred in Italy and Slovenia. At present the procedure is fully implemented and routinely analyzes the events recorded by the Transfrontier network ([Costa et al. 2010](#)) and the Italian strong motion network (RAN) ([Gorini et al. 2010](#), [Zambonelli et al. 2011](#)) and provides a stable estimation of moment magnitude within few minutes from the first earthquake detection. Thanks to these two networks a large number of data for each event is obtained, an essential pre-requisite to obtain accurate and robust estimates of earthquake source parameters. The real-time estimate of earthquake related parameters (e.g., waveforms, epicenter coordinates, ground motion parameters, Fourier spectra, response spectra and other parameters to engineering related parameters) is one of the issues of most interest to the Civil Defence for an immediate assessment of the effects and characteristics of an earthquake.

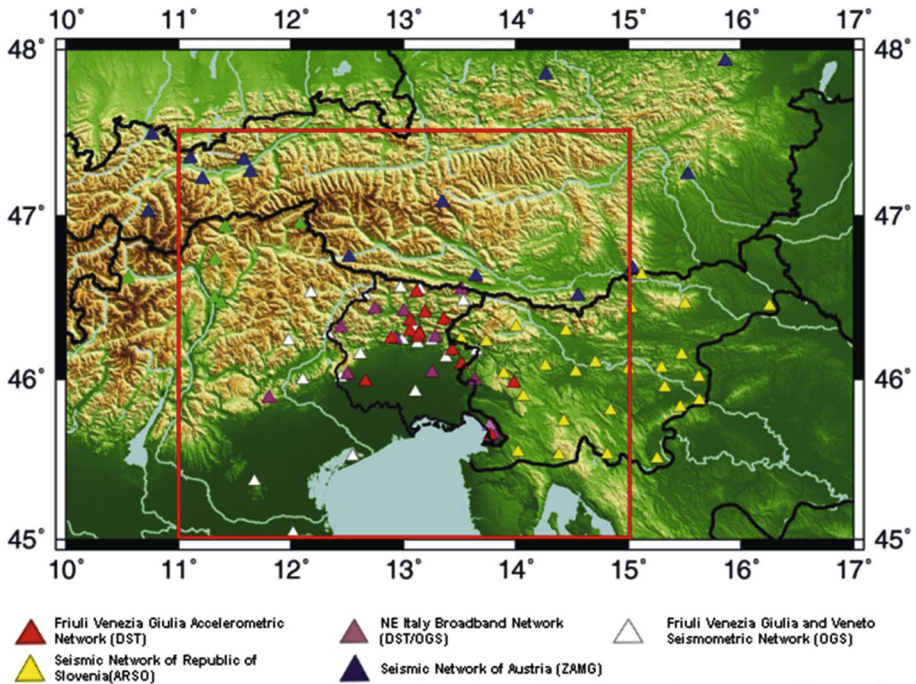


Fig. 1 The Transfrontier network. In *red* the Friuli Venezia Giulia Accelerometric Network (RAF); in *yellow* the Republic of Slovenia Seismic Network; in *pink* the NE Italy Broadband Network; in *blue* the Seismic Network of Austria; in *white* the Friuli Venezia Giulia and Veneto Network. The area delimited by the *red square* represents the main zone in which Mw is at present routinely estimated

2 Networks and data

The procedure for near real-time seismic source parameters has been routinely running at the Mathematics and Geosciences Department (DMG) of University of Trieste using Transfrontier network data and since one year has been also applied to events recorded by the RAN managed by Civil Protection (DPC) in Rome.

The first stations of the Transfrontier network were installed within the European project Interreg IIIa Italy-Austria “Seismological networks without borders in southeastern Alps”, based on a close collaboration between several partners: the Civil Defense of the Friuli Venezia Giulia region, Palmanova (PCFVG); the Seismological Research and Monitoring (SeisRaM) group of the Department of Earth Sciences [now of Mathematics and Geosciences], Trieste; the Zentralanstalt für Meteorologie und Geodynamik (ZAMG) in Vienna, Austria; the Environmental Agency of the Republic of Slovenia (ARSO) in Ljubljana, Slovenia; the National Institute of Oceanography and Experimental Geophysics (OGS) of Trieste. The creation of this integrated network proved to be very important, because it has since allowed for quick access to many well-distributed high-quality data, recorded also in the vicinity of small events. At present (January 2013), the Transfrontier network (Fig. 1) consists of more than 100 instruments including broadband, accelerometric and short-period stations, which cover the total area of SE Alps with a good station geometry, especially in Friuli Venezia Giulia and Slovenia.

In Friuli Venezia Giulia the Accelerometric network (RAF) is operated (Costa et al. 2010) by the SeisRaM group. The first RAF stations were installed in the period 1993–1995 by the former Department of Earth Sciences and from the beginning they were equipped with digital GPS time-synchronized three-component instruments. The new high-dynamic (>120 db) instrumentation is formed by three-component digital broadband feedback accelerometers coupled with 18 or 24 bits acquisition systems. These free-field stations are mainly installed on rock. Today RAF is configured taking into account also other network stations operating in Italy and it is part of the Transfrontier network that records accelerations at several important sites in the SE Alps area. This allows an immediate estimate of peak ground acceleration as well as a first evaluation of possible damage (Costa et al. 2010).

The data from the stations of the Transfrontier network are collected in near real-time by the Antelope[®] software (BRTT, Boulder). This software acquires and processes data from the network stations; at the end it archives all data, both raw data retrieved from stations as the related estimated parameters, into a database.

The National Strong Motion Network, RAN, is distributed on the national territory to record earthquakes of medium and high intensity, managed by the Seismic Monitoring Service of the Territory of the Seismic and Volcanic Risks Office of the Civil Protection Department (DPC) in Rome (Gorini et al. 2010; Zambonelli et al. 2011). RAN has about 400 digital stations equipped with a GSM modem or GPRS, connected to the RAN data capture Centre of Rome (*last update: 20 May 2011*). Since one year the same procedure implemented at RAF has been also applied to events recorded by the RAN. The same Antelope[®] software and the procedure to determine moment magnitude and all seismic source parameters in near real-time, is now installed also at DPC for managing and analyzing recorded data.

All these networks provide a large number of high-quality data. In particular, the results of our analyses are better when using events recorded by RAN stations, as we will show and discuss in the following. In fact, the RAN stations consist of the same type of instruments and all instrumental updates are available in real time and this leads to improved moment magnitude estimates. For the Transfrontier network, on the other hand, instrumental and other updates might not be known in real time, thus introducing potential errors in the moment magnitude estimations.

To validate our procedure we used the recordings of some recent moderate earthquakes that occurred in Slovenia and in Italy: Carnia 2002; Bovec 2004; Parma 2008; L'Aquila 2009; Macerata 2009; Emilia 2012 (Fig. 2). In particular, in the L'Aquila (2009) sequence case we consider the main event and 7 aftershocks; in the Emilia (2012) earthquake sequence we consider the two main events of May 20 and May 29, plus several aftershocks with magnitudes ranging from 3 to 5 for which we have moment magnitude estimates provided by other Institutions.

We have also tested our procedure using 100 events recorded by RAN in 2012 from January to June with $M_L > 3$. Additionally, we have analyzed a data set of about 743 events recorded by the Transfrontier network in 2012 and automatically located with $M_L > 1$. This test was needed in order to validate and to improve the reliability of our procedure.

3 Method

3.1 Data processing

Our procedure determines automatically and in near real-time the moment magnitude and other seismic source parameters by spectral analysis following Andrews' method (1986).

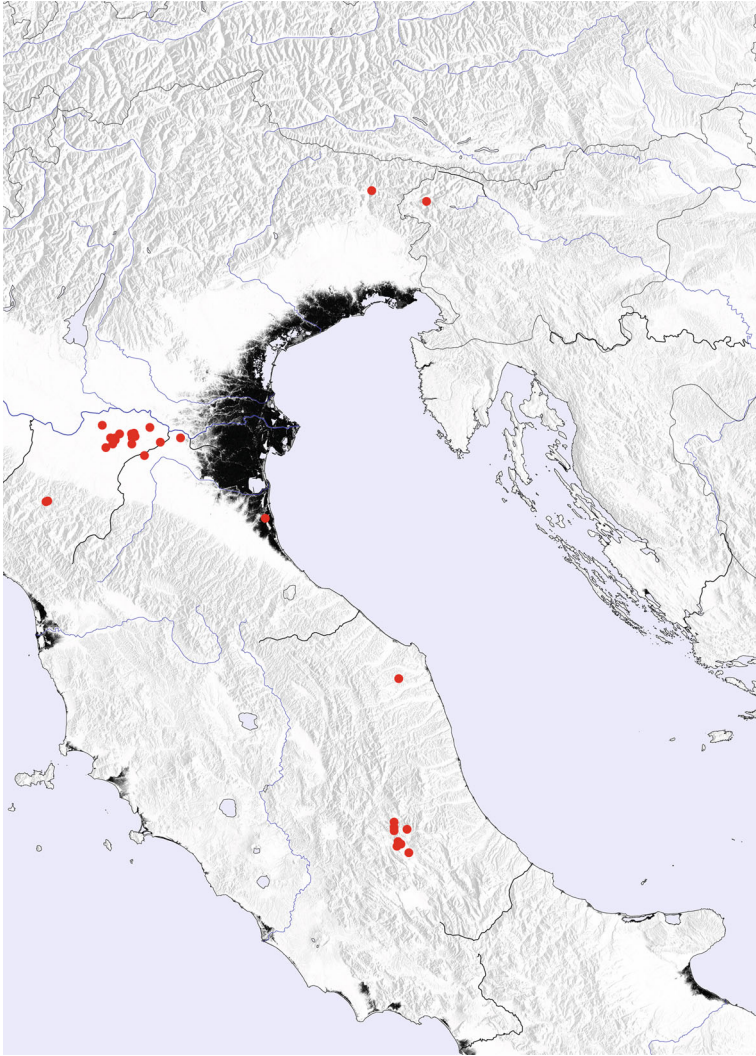


Fig. 2 Maps of recent moderate events used for the validation of the moment magnitude estimation procedure

From the Antelope database we retrieve the seismic waveforms and those event parameters, such as azimuth, epicentral distance, signal type (acceleration or velocity) and sampling frequency, which are necessary for our algorithm. We have optimized this software by creating new tables in the database in order to interface it with our algorithm.

We remove the average, trend, spike and instrumental response. We set the epicentral distance up to the maximum of 70 km. We have checked the quality of each record and selected the band-pass corner frequencies, using signal-to-noise (SNR) ratio, that provide the frequency window in which the signal is analyzed. The “noise” window lasts for ten seconds before the P arrival. The “signal” window starts from the S arrival and its duration is a function of event-epicenter distance, in particular is equal to the distance divided by 3 km/s, an empirically determined value. P and S phases are detected by Antelope® software.

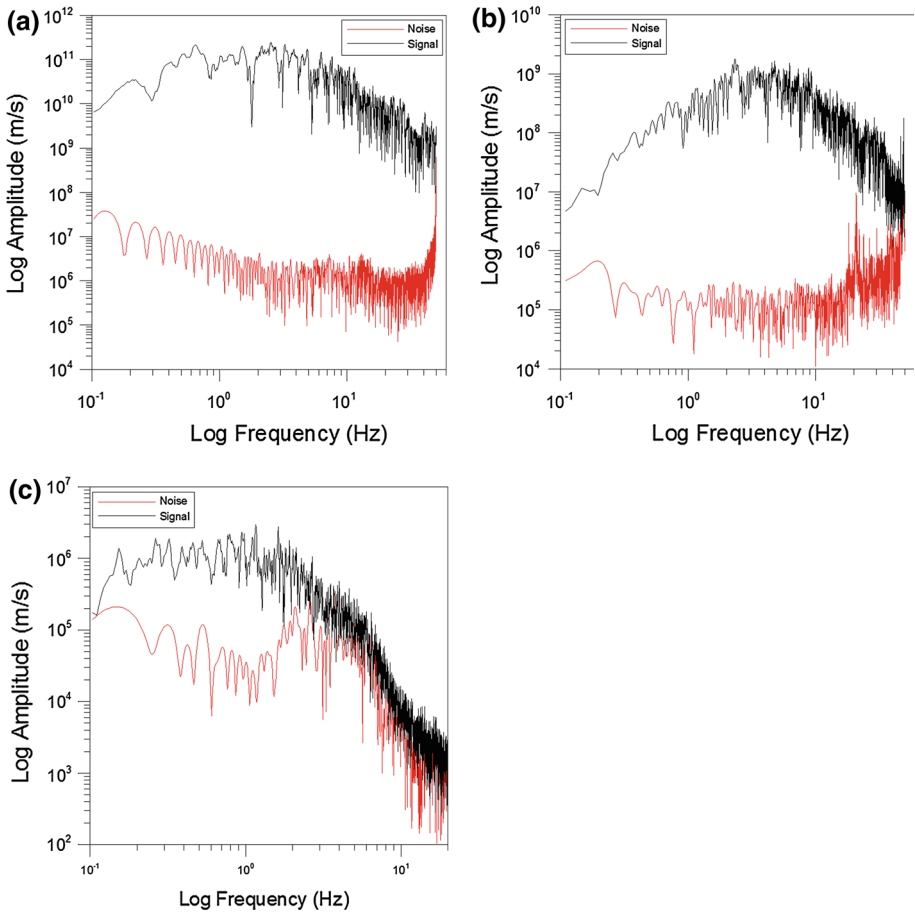


Fig. 3 Examples of signal and noise acceleration spectra used in the procedure to define the frequency windows in which the signal is processed. **a** Excellent SNR in all frequency range. **b** Good SNR at low frequency up to 20 Hz. **c** Good SNR in frequency range $0.2 < f < 1.6$

Since the noise spectra might be of the same order as the level of the signal and even dominate the spectral shape at low and high frequencies, it is essential to determine the frequency range within which the signal spectral level is significantly higher than the noise one. We apply the following procedure to the investigated frequency range (usually 0.1 and 50 Hz), that allows to select the frequency window within which $\text{SNR} > 3$. Starting from the lowest frequency and using a 50-points sliding window, we calculate the minimum frequency for which $\text{SNR} > 3$ for all frequencies within the first half of the smoothed (3 point running average) spectrum; we apply the same procedure to the second half of the investigated range, this time starting from the highest frequency and moving towards lower frequencies (Fig. 3).

We apply a Butterworth band pass filter, then we obtain accelerations, velocities and displacements doing the derivative or the integral of the signal. We apply the Fast Fourier Transform (FFT) to obtain the signal spectra. Then we correct them for geometrical spreading and intrinsic attenuation to retrieve the source spectra.

3.2 Seismic moment and corner frequency estimation

For seismic source parameters estimation, we rotate the horizontal components of the signal to obtain the transverse component of motion. We choose to use SH waves for several reasons: P waves have the advantages of being the first to arrive and have therefore usually a clear onset, but they are soon overlapped by the subsequent arrivals of converted waves. This time limit affects the minimum detectable frequency for the P phase. Also SV waves are contaminated with converted phases and interfere with the vertical component of P waves.

The signal spectrum can be expressed as a function of the frequency, f , by means of the following formula:

$$A(f) = D(f) E(f) G(R) \quad (1)$$

where R is the hypocentral distance, $D(f)$ represents the source spectrum, $E(f)$ is the attenuation function and $G(R)$ is the geometrical spreading.

The source term used is a simple ω^2 model proposed by [Brune \(1970\)](#):

$$D(f) = \frac{M_0}{4\pi k \rho v^3} \left[1 + \left(\frac{f}{f_0} \right)^2 \right]^{-1} \quad (2)$$

where M_0 is the seismic moment, $k = (\sqrt{2} \cdot 0.6)^{-1} = 0.91$ is a correction factor for free-surface reflection (factor 2) and for displacement radiation pattern (factor 0.6), ρ is the density, v is the S-wave velocity at the source, and f_0 is the corner frequency.

For the attenuation term $E(f)$ we use:

$$E(f) = e^{-\left(\frac{\pi T f}{Q(f)}\right)} \quad (3)$$

where T is the travel-time, f the frequency and Q the frequency-depend attenuation factor.

Following [Console and Rovelli \(1981\)](#):

$$Q(f) = 80 \cdot f^{1.1} \quad (4)$$

The authors obtained attenuation parameter for Friuli region from strong motion spectra in the frequency range of 0.1 and 10 Hz with the maximum distance of 200 Km. For the wider SE Alps area various studies on the attenuation model have been done and all of them have observed a frequency-dependent Q ([Console and Rovelli 1981](#); [Castro et al. 1996](#); [Malagnini et al. 2002](#)). We choose the Console and Rovelli relationship because we considered only locally recorded events and direct wave and did not take in account a complex geometrical spreading. We report the considerations about in discussion section.

For the geometrical spreading term we assume the one for body-wave theory:

$$G(R) = \frac{1}{R} \quad (5)$$

Following [Andrews \(1986\)](#), the velocity power spectrum $V^2(f)$ and the displacement power spectrum $D^2(f) = V^2(f) / (2\pi f)^2$ are used to calculate the following integrals:

$$S_{D2} = 2 \int_0^{\infty} D^2(f) df \quad (6)$$

$$S_{V2} = 2 \int_0^{\infty} V^2(f) df \quad (7)$$

The values of these two integrals go from zero to infinity whereas the spectrum is effectively computed for a limited frequency range. Di Bona and Rovelli (1988) discuss about this limitation. They notice that two cut-off frequencies limit the spectral range over which the two integrals have to be computed: the low-frequency cut-off is due to the finite duration of time history of motion, the high-frequency cut-off is given by the Nyquist frequency, $f_N = 1/2\Delta t$, where Δt is the sampling rate. In practice, the available bandwidth is limited by the instrumental response and the presence of noise sources at low and high frequency. So the two integrals (6) and (7) become:

$$S_{V2} = 2 \int_{f_{inf}}^{f_{sup}} V^2(f) df \quad (8)$$

$$S_{D2} = 2 \int_{f_{inf}}^{f_{sup}} D^2(f) df \quad (9)$$

and are exactly solvable:

$$S_{D2} = \frac{1}{4} \Omega^2 (2\pi f_0) \quad (10)$$

$$S_{V2} = \frac{1}{4} \Omega^2 (2\pi f_0)^3 \quad (11)$$

Also the corner frequency is computed in terms of these integrals as

$$f_0 = \frac{1}{2\pi} \sqrt{S_{V2}/S_{D2}} \quad (12)$$

For the Brune spectrum the low-frequency spectral level is related to the integrals as

$$\Omega^2 = 4S_{D2}^{3/2} S_{V2}^{-1/2} \quad (13)$$

As we know, this value is related to the seismic moment:

$$M_0 = 4\pi\rho\beta^3\Omega k \quad (14)$$

Finally, the moment magnitude is determined according to the Kanamori (1977) formula:

$$M_W = \frac{2}{3} \log_{10} M_0 - 6.1 \quad (15)$$

At the end of the procedure all results are stored and shown in a database table and a report is generated within 10 min from the event.

4 Validation

The recent moderate events used to validate the procedure (Fig. 2) are listed in Table 1 in which the location, the local (Richter) magnitude M_L calculated by Antelope[®] software, the moment magnitude estimation are reported. The error of our moment magnitude estimation is given by its standard deviation and is linked to the number of stations within the pre-defined distance range (0–70 km).

Table 1 List of the results regarding the recent main events occurred in Italy and Slovenia used for validating our M_W procedure: Carnia (2002), Bovec (2004), Parma sequence (2008), L'Aquila sequence (2009), Macerata (2009), Emilia sequence (2012)

	Latitude (°N)	Longitude (°E)	Depth (km)	Date (gg/mm/aa)	Time (hh:mm)	M_L (Antelope)	M_w (SeisRaM)	M_o (N*m)
CARNIA	46.38	13.16	10.0	12/02/02	3:18		4.6 ± 0.3	$3.03E + 16$
BOVEC04	46.32	13.63		12/07/04	13:04	5.1	5.0 ± 0.1	$1.99E + 17$
PARMA	44.50	10.38	30.8	23/12/08	15:24	5.1	5.0 ± 0.1	$5.03E + 16$
	44.50	10.37	30.0	23/12/08	21:58	4.7	4.90 ± 0.04	$3.03E + 16$
AQUILA	42.35	13.38	9.5	06/04/09	1:32	5.8	5.9 ± 0.2	$2.27E + 18$
	42.36	13.41	9.3	07/04/09	9:26	4.7	4.9 ± 0.2	$3.93E + 16$
	42.31	13.48	20.5	07/04/09	17:47	5.3	5.4 ± 0.2	$1.98E + 17$
	42.38	13.39	6.0	07/04/09	21:34	4.2	4.7 ± 0.2	$1.91E + 16$
	42.47	13.35	14.6	09/04/09	0:53	5.1	5.3 ± 0.2	$1.67E + 17$
	42.46	13.47	11.1	09/04/09	4:32	4.0	4.5 ± 0.2	$1.01E + 16$
	42.50	13.36	13.3	09/04/09	19:38	4.9	5.1 ± 0.1	$8.43E + 16$
	42.45	13.36	14.2	22/06/09	20:58	4.5	4.6 ± 0.2	$1.49E + 16$
MACERATA	43.41	13.39	37.7	20/09/09	3:50	4.6	4.4 ± 0.0	$1.32E + 16$
EMILIA	44.96	11.26	7.9	20/05/12	2:03	6.1	6.1 ± 0.2	$4.55E + 18$
	44.90	11.13	10.4	20/05/12	3:02	5.3	5.3 ± 0.2	$3.38E + 17$
	44.89	11.11		20/05/12	9:13	4.4	4.7 ± 0.1	$2.40E + 16$
	44.89	11.52	2.5	20/05/12	13:18	5.3	5.2 ± 0.1	$3.41E + 17$
	44.87	11.35	10.0	20/05/12	17:37	4.8	4.8 ± 0.2	$4.58E + 16$
	44.87	11.35	10.0	21/05/12	16:37	4.3	4.6 ± 0.2	$1.35E + 16$
	44.87	11.35	6.0	23/05/12	21:41	4.5	4.5 ± 0.3	$1.24E + 16$
	44.92	11.12	6.2	25/05/12	13:14	4.2	4.4 ± 0.3	$9.46E + 15$
	44.78	11.21	10.2	27/05/12	18:18	4.3	4.5 ± 0.3	$1.54E + 16$
	44.92	11.10	4.0	29/05/12	7:00	5.8	5.9 ± 0.2	$2.29E + 18$
	44.89	10.96	11.4	29/05/12	8:25	4.8	4.8 ± 0.2	$2.36E + 18$
	44.86	11.10		29/05/12	9:30	4.1	4.2 ± 0.2	$4.47E + 17$
	44.88	10.96	4.9	29/05/12	10:55	5.4	5.5 ± 0.3	$3.40E + 19$
	44.89	10.92	12.1	29/05/12	18:28	4.0	4.3 ± 0.2	$4.67E + 17$
	44.83	10.88	14.8	31/05/12	14:58	4.3	4.2 ± 0.2	$3.93E + 17$
	44.86	10.94	15.0	31/05/12	19:04	4.5	4.5 ± 0.3	$1.69E + 18$
	44.92	11.00	8.0	03/06/12	19:20	5.0	5.2 ± 0.2	$1.07E + 18$
	44.40	12.25	12.2	06/06/12	4:08	4.6	4.3 ± 0.2	$4.51E + 17$
	44.97	10.85	8.0	12/06/12	1:48	4.8	4.6 ± 0.2	$1.39E + 18$

Table 2 List and comparison of moment magnitudes of the recent events occurred in Emilia Romagna region, Italy, as estimated by various sources (SeisRaM: this paper; INGV: <http://cnt.rm.ingv.it/>; Pondrelli et al. 2012; Saraò and Peruzza 2012; Malagnini et al. 2012; <http://www.globalcmt.org/CMTsearch.html>; <http://geofon.gfz-potsdam.de/eqinfo/list.php>)

Event	SeisRaM	No of used stations	INGV	Pondrelli et al. (2012)	Saraò and Peruzza (2012)	Malagnini et al. (2012)	CMT	GFZ
19/05/2012 23:13	4.40 ± 0.01	3	4.0	4.0	4.2			4.1
20/05/2012 2:03	6.1 ± 0.2	13	5.9	6.1	6.1	5.6	6.1	6.1
20/05/2012 3:02	5.3 ± 0.2	12	4.9	5.1	5.0	4.7	5.1	
20/05/2012 9:13	4.7 ± 0.1	7	4.1			4.1		
20/05/2012 13:18	5.2 ± 0.1	11	5.0	5.2	5.1	4.8	5.2	5.1
20/05/2012 17:37	4.8 ± 0.2	6	4.3	4.5	4.4			
21/05/2012 16:37	4.6 ± 0.2	8	4.0	4.1	4.1			
23/05/2012 21:41	4.5 ± 0.3	16	3.9	4.1	4.1	3.7		
25/05/2012 13:14	4.4 ± 0.3	15	3.7		3.9	3.6		
27/05/2012 18:18	4.5 ± 0.3	16	3.8		3.9	3.7		
29/05/2012 7:00	5.9 ± 0.2	25	5.7	6.0	5.9	5.4	5.9	5.8
29/05/2012 8:25	4.8 ± 0.2	12	4.4	4.7	4.6	4.3		
29/05/2012 9:30	4.2 ± 0.2	13	3.7		4.0	3.5		
29/05/2012 10:55	5.5 ± 0.3	23	5.3	5.5	5.5	5.1	5.5	5.5
29/05/2012 18:28	4.3 ± 0.2	11	3.8		4.0			
31/05/2012 14:58	4.2 ± 0.2	14	3.7					
31/05/2012 19:04	4.5 ± 0.3	16	3.9		3.9			
03/06/2012 19:20	5.2 ± 0.2	25	4.8	4.9	4.8		5.0	4.7
06/06/2012 4:08	4.3 ± 0.2	17		4.3	4.1			
12/06/2012 1:48	4.6 ± 0.2	21	3.9	4.2	4.1			

In particular for the Emilia sequence several moment magnitude estimations are available for comparison with our results. In Table 2 we report our estimates, the magnitudes estimated by the Istituto Nazionale di Geofisica e Vulcanologia (INGV) along with ones obtained by other authors. The moment magnitudes are obtained by the others agencies and authors using the well-known time domain moment-tensors inversion code TDMT_INVc and its changes. Dreger (2003) developed his method to determine seismic moment tensors by inversion of three-component broadband waveforms.

For events with $M_W > 5$, our estimates are in agreement with those by Pondrelli et al. (2012) and Saraò and Peruzza (2012) and are a bit larger than INGV ones. For events with $M_W < 5$ the agreement is quite fair, but clearly there are discrepancies principally with the estimations by Malagnini et al. (2012) and INGV (Fig. 4). The reasons for discrepancies for some values could be connected to different initial assumptions to compute moment magnitude, such as, e.g.: velocity model, epicentral distance and frequency range (Malagnini et al. 2012; Pondrelli et al. 2012; Saraò and Peruzza 2012; Scognamiglio et al. 2012).

We have also validated our procedure applying it to the recordings of some minor ($2 < M_W < 4$) events in the SE Alps area for which independent seismic moment and moment magnitude estimates were available. Figure 5 shows the comparison between the results for all events used to validate our procedure, taking into account only the moment magnitude estimated by CRS (Centro Ricerche Sismologiche) of OGS and INGV.

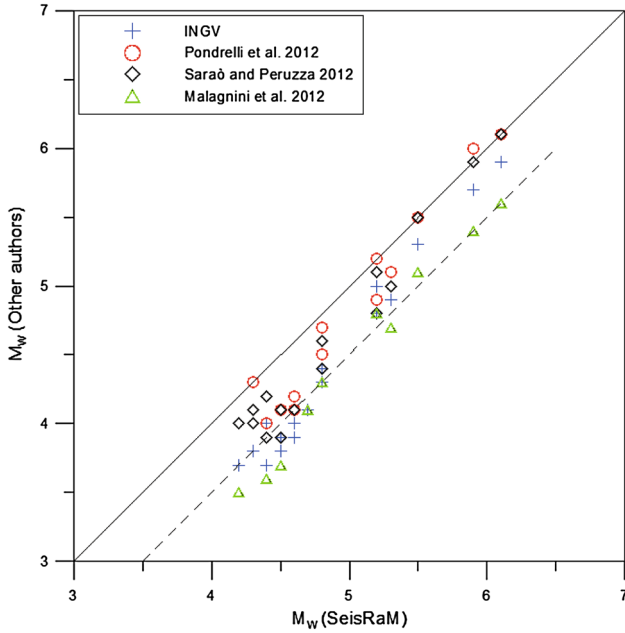


Fig. 4 Comparison between the moment magnitude estimated by the procedure described in this paper (SeisRaM) and other Agencies for Emilia 2012 earthquakes

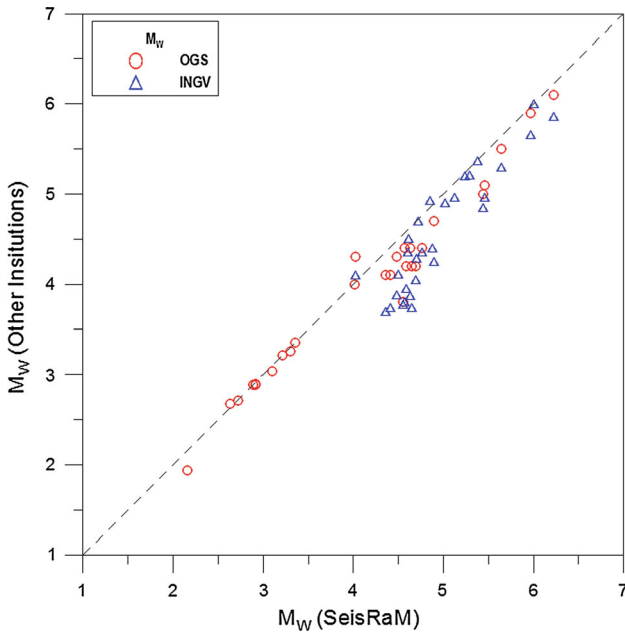


Fig. 5 Comparison between the moment magnitude estimated by INGV (blue triangles) and OGS (red circles) and the moment magnitude estimated by the procedure described in this paper (SeisRaM)

5 Results

The procedure is currently routinely estimating the moment magnitude of events recorded by both RAN (from 2012) and the Transfrontier network.

In Fig. 6 we report the results obtained for 100 events with $M_L > 3$ recorded by the RAN network from January to June 2012. For this test we also compared the data reviewed or not by human to understand how a good location influences the estimates. The Fig. 6a reports the results using automatically located data in which a scatter is obvious. The results improve (Fig. 6b), when locations reviewed by human are used leading to the conclusions that a good location plays a fairly important role in the source parameter estimations. The mean standard deviation associated to these moment magnitude estimates is 0.2.

We tried to establish the quality of our results by plotting also the corner frequency versus moment magnitude (Fig. 7). The obtained results seem quite good: the trend is the hypothetical expected one and the red curves denote constant stress drop (0.1, 1 and 10 bars). The dataset (100 analyzed earthquakes) contains essentially the events that occurred in the Emilia (2012) earthquake area (pre main event and aftershocks). The Emilia (2012) sequence is characterized by quite low stress drops and rather high energy at low frequencies (Malagnini et al. 2012; Castro et al. 2013). Our low values of stress drop estimates are, therefore, in agreement with these observations. The high amount of energy at low frequencies could be also the reason behind the higher values of M_W with respect to those of M_L (Fig. 6b).

Figure 8 shows the results of the procedure applied to the 2012 period of the Transfrontier network where all earthquakes are automatically located and analyzed. In this case a discrepancy is evident, mainly at low magnitudes, very probably due to errors in the automatic location or maybe in the not updates metadata. The standard deviation associated to these estimates range between 0.3 and 0.5.

6 Discussion

The procedure for estimating M_W was devised in order to have a common magnitude for events occurring in the southeastern Alps that are recorded by the Transfrontier network. Up to now, this procedure was used mainly for the events located in this area. For the wider SE Alps area various studies on the attenuation model have been done and all of them have observed a frequency-dependent Q (Console and Rovelli 1981; Castro et al. 1996; Malagnini et al. 2002). We have adopted the frequency-dependent attenuation parameter proposed by Console and Rovelli (1981) for the Friuli Venezia Giulia region. At first we used the procedure only with locally recorded events and direct waves without taking into account a complex geometrical spreading. The initial conditions, epicentral distance and frequency range, were similar to those proposed by Console and Rovelli (1981). Since about a year, the procedure is being routinely used also at DPC in Rome, estimating the source parameter values for all events recorded by DPC in Italy. This suggests that it would be better to perform a more accurate attenuation analysis and dependent on the studied area: using a geometrical spreading as a function of distance, together with a frequency-dependent Q coefficient.

It should be noted that our method for estimating M_W does not take into account the site effects influence on the spectral amplitudes. Even if we have high-quality data coming mostly from instruments located either in boreholes or on rock, in some cases, mainly concerning small events, there is a strong influence of site effects on the spectral amplitude (Fig. 9). This is reflected on the magnitude and on the corner frequency (Brune 1970, 1971):

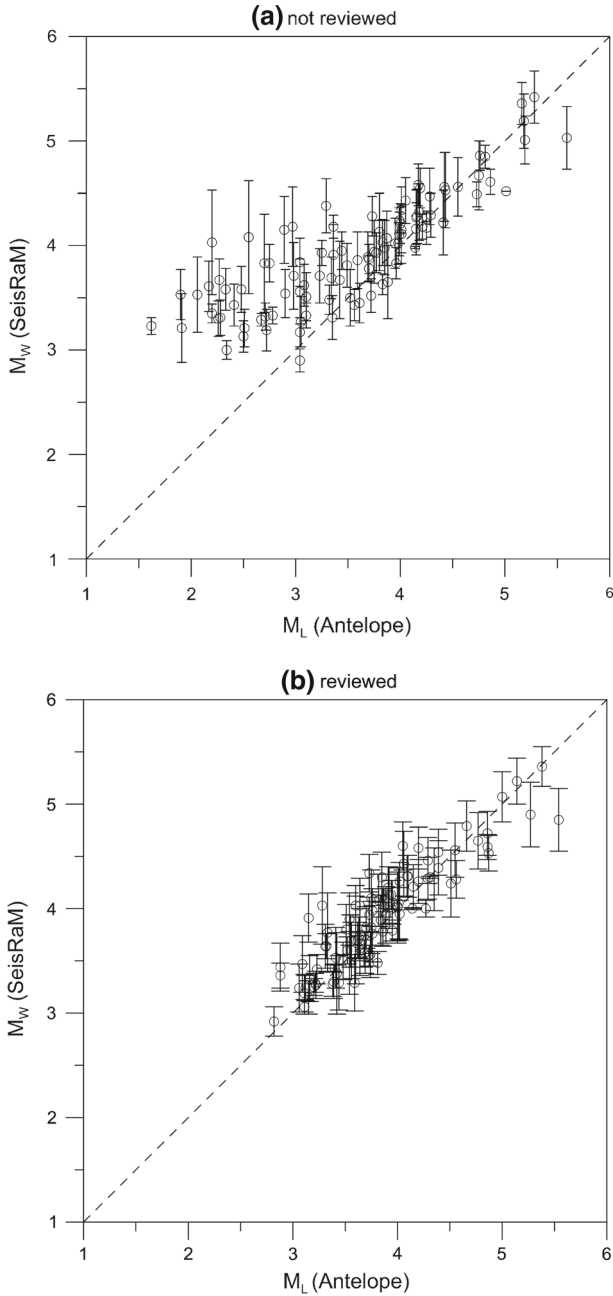


Fig. 6 Comparison between the local magnitude as estimated by Antelope and the moment magnitude estimated by the procedure described in this paper (SeisRaM) for events recorded by the RAN network. **a** Automatic event location. **b** Event location reviewed by human. The *error bars* denote the SD of the data points

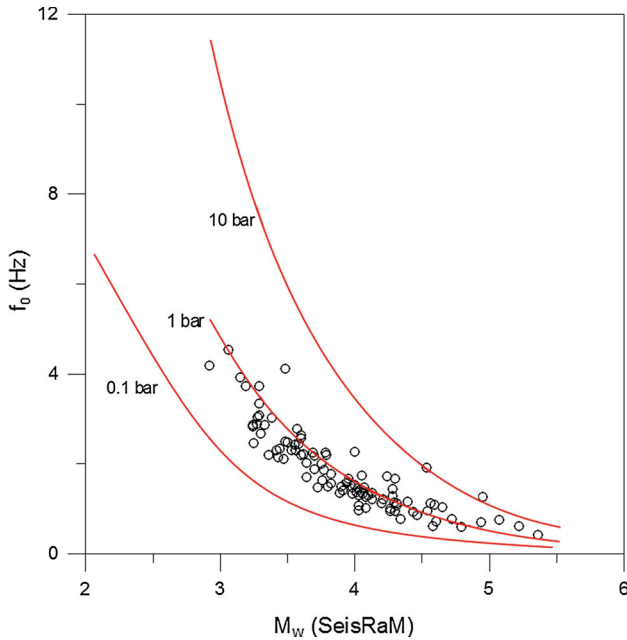


Fig. 7 Corner frequency versus moment magnitude for the events recorded by the RAN. The curves are theoretical ones that denote three different values of stress drop. The locations of these events were reviewed by human

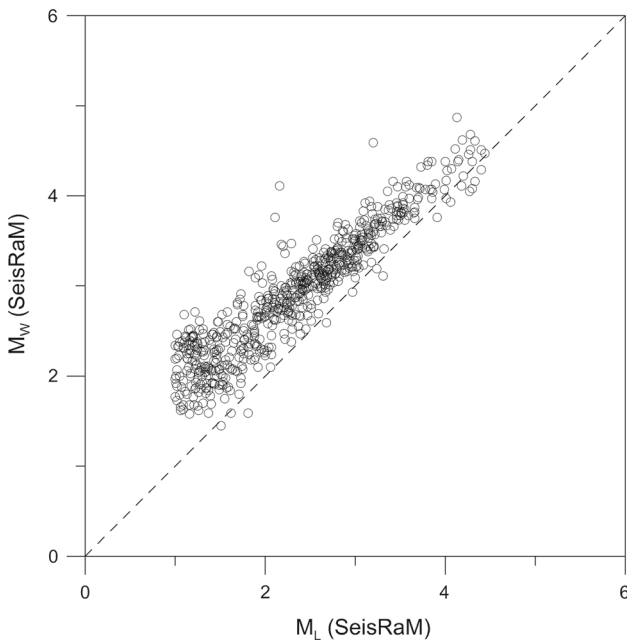


Fig. 8 Comparison between M_L estimates by Antelope software and M_W values, obtained from our procedure, for $1 \leq M_L \leq 5$ events that occurred in the SE Alps in 2012

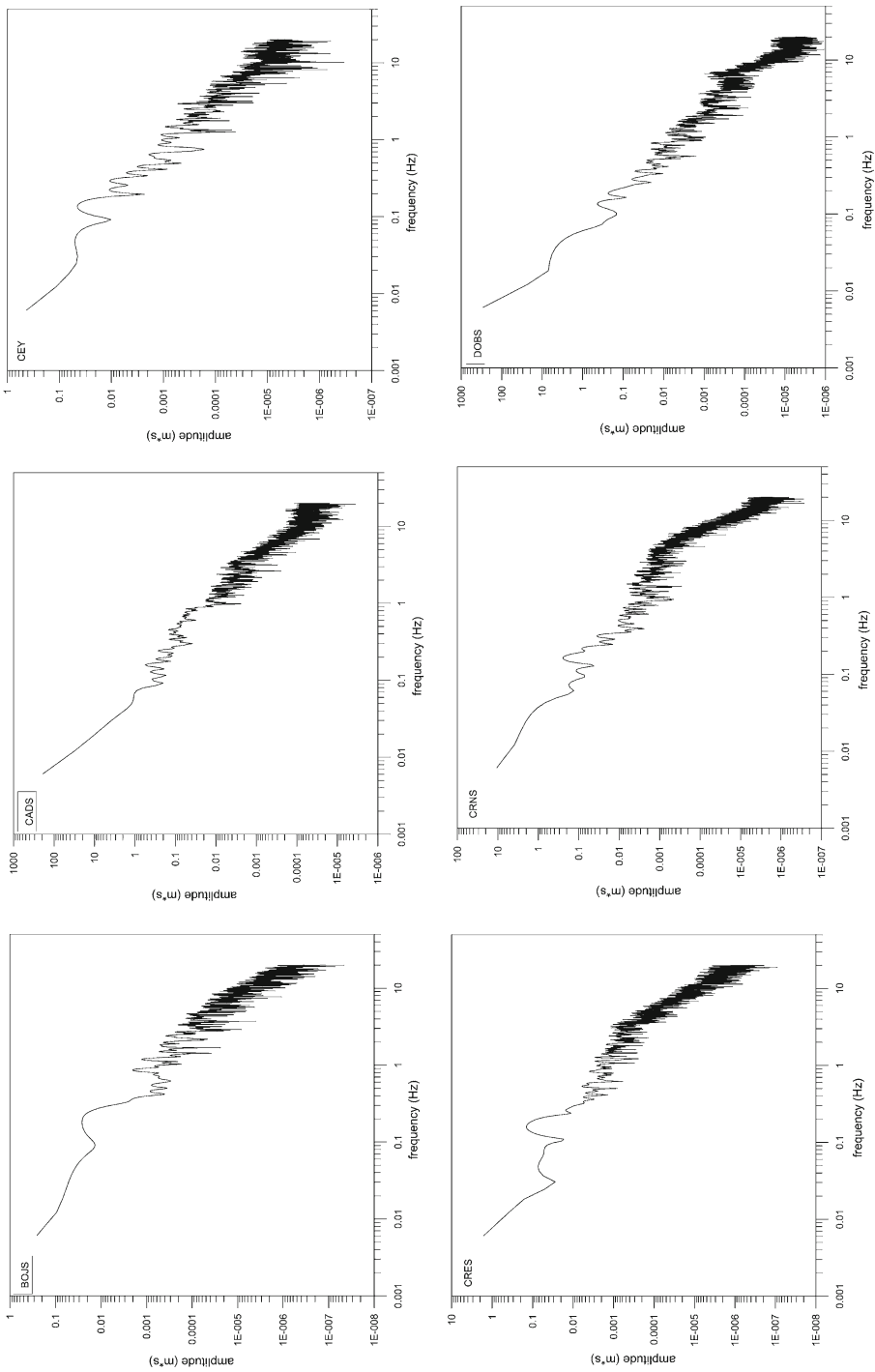


Fig. 9 Example of source spectra (displacement) of a small event ($M_L = 1.2$) recorded by the Transfrontier network at different stations

$$f_c = \frac{2.34}{2\pi} \frac{v_S}{r}$$

with v_S the S-wave speed and r the equivalent radius of the fault. In particular, the f_c estimates are much more questionable. Generally, the obtained source spectra show a good agreement with the [Brune \(1970, 1971\)](#) model leading to an accurate corner frequency value. In some cases, especially for events with low magnitude ($M_W < 3$), the spectrum shows the presence of distorting effects such as bumps and spurious frequencies, often confined in the frequency band close to the corner frequency (Fig. 9), that leads to erroneous corner frequency estimates. In fact, sometimes it seems that the source spectrum has two corner frequencies, requiring the use of a different source model. The relative error associated to the corner frequency is, therefore, rather high and ranges between 10% and 50%. The problem of currently taking into account both site effects and attenuation effects is not an easy one and could be perhaps dealt with by stacking spectra of different stations. A more detailed analysis of the effects associated with the propagation such as geometrical spreading, scattering, intrinsic attenuation and site effect, would probably also improve the results. The determination of a stable corner frequency remains therefore an open question that requires further work.

7 Conclusions

Using a procedure implemented by the SeisRaM group of the Department of Mathematics and Geosciences (DMG) of the University of Trieste, we estimate the seismic moment in near real-time, as well as the moment magnitude and the corner frequency of the events recorded by the Transfrontier and by the RAN networks. Assuming the [Brune \(1970, 1971\)](#) source model and using spectral analysis we estimate the seismic source parameters following [Andrews \(1986\)](#). The main advantage of this method is that it is based on a physical model and is independent from empirical relationships.

We use the recent events that occurred in Italy and Slovenia such as Emilia (2012), Macerata (2009), L'Aquila (2009), Parma (2008), Bovec (2004) and Carnia (2002) earthquakes to validate this procedure. In particular, for the Emilia sequence we compared our obtained moment magnitude estimates with those obtained by other agencies and authors. For events with $M_W > 5$ there is a good agreement among the various estimations. The discrepancy for some (especially low) M_W values can be associated to the different initial assumptions of the methods used (e.g. velocity model, epicentral distances, frequency range). We have also used some events that occurred in the SE Alps area for which we have independent moment magnitude estimates.

We have analyzed about 843 events in the magnitude range $1.0 < M_W < 6.3$. A stable moment magnitude estimate is available within a few minutes from the first earthquake detection. The results show that the procedure is reliable and robust for events with $M_W > 3$ recorded by RAN stations and a good correspondence between M_W and M_L is noted.

In the case the procedure is applied to events recorded by the Transfrontier stations with automatic location and analysis, the results are less acceptable for small-magnitude events. A possible reason could be due to site effects not yet taken into account by the procedure. The not perfect correspondence between M_W and M_L for magnitude below 3 is a result that needs further investigations. Furthermore a more detailed analysis of the attenuation and spreading of waves will most probably lead to improved estimates of seismic source parameters.

In conclusion, our moment magnitude estimates are in reasonable agreement with both the local magnitude and the moment magnitude values calculated by most other authors.

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