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# The Wood–Anderson of Trieste (Northeast Italy): One of the Last Operating Torsion Seismometers


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*Online Material:* Earthquake catalog.

## INTRODUCTION

The Wood–Anderson (WA) torsion seismograph, used by Richter (1935) for the definition of the local magnitude ($M_L$) of an earthquake, has been abandoned over time due to the cumbersome nature of its use. With the progress of technology, modern digital broadband (BB) instruments have replaced older instruments such as the WA, and the equivalent $M_L$, obtained from simulated WA seismograms after convolution of the recorded BB data with a proper transfer function (Bormann, 2002a,b), has replaced the WA $M_L$.

Despite the paucity of WA instruments today, the $M_L$ in its original form remains relevant for continuity with old earthquake catalogs and as a long-standing reference for all other magnitude scales up to approximately $M_L$ 6.5. For larger earthquakes, the $M_L$ scale progressively underestimates the actual energy release and $M_L$ is said to saturate (Kanamori, 1983). Even so, $M_L$ is a good predictor of structural damage caused by earthquakes because many buildings have resonant periods close to that of the WA seismograph (0.8 s).

In Trieste, located in northeastern Italy, there is one of the few stations equipped with an original pair of WA instruments that are still operating. The two horizontal WA seismometers (Lehner-Griffith TS-220) were installed in September 1971 and have been managed since then by the Osservatorio Geofisico Sperimentale, presently the Istituto Nazionale di Oceanografia e di Geofisica Sperimentale (OGS). The Trieste station was part of the Worldwide Standardized Seismographic Station Network (WWSSN) with the code TRI-117, and it dates its operation back to 29 July 1963. At that time, three Benioff seismometers were employed as short-period seismographs, and three Ewing-Press seismometers were used for teleseismic detection. The WWSSN seismometers were installed at the bottom (161 m above sea level) of Grotta Gigante, a giant cave of the Trieste karst, 12 km away from the city center. The WA seismometers were placed over a plinth in a darkroom at the surface (336 m above sea level, latitude 45.709° N, longitude 13.764° E). The daily processing of the photographic paper was quite expensive and very time consuming. This aspect also contributed to the abandonment of the Trieste WA recordings in April 1992.

In 2002, the WA instruments were recovered and upgraded by replacing the recording on photographic paper with an electronic device. From 17 December 2002 to 31 December 2013, the refurbished WA seismometer recorded 1252 events, with a break between May 2005 and March 2010 due to the restoration of the building where it was operated.

At present, the Trieste station concurrently acquires data from (1) the upgraded digitized WA seismometer, (2) the Güralp 40-T BB instrument placed at the top of the cave (since 2004), and (3) the BB Streckeisen STS-1 seismometers installed at the bottom of the cave (since 1995). The STS-1 instrument corresponds to station TRI of the MedNet network (Mazza *et al.*, 1998).

In this paper, after describing the upgrade of the WA seismograph and verifying its static magnification ($G_s$), we reevaluate old estimates of $M_L$ and compile a new catalog of Trieste WA $M_L$ values updated to 2013. Finally, we compare the Trieste WA $M_L$ values with moment magnitudes.

## MODERNIZATION OF THE WOOD–ANDERSON OF TRIESTE

A tungsten wire, tightly stretched between two suspension lugs, composed the original WA instrument (Anderson and Wood, 1925). At the middle of the wire, there was a copper cylinder to which a small moving planar mirror was fastened (Fig. 1a).

To obtain the damping for the torsional vibrations, the copper cylinder was suspended in a magnetic field generated by a permanent magnet. When the system was energized, the cylinder moved and generated Foucault currents proportional to the velocity of the cylinder such that the resulting magnetic field contrasted with that of the permanent magnet, damping the movement of the cylinder.

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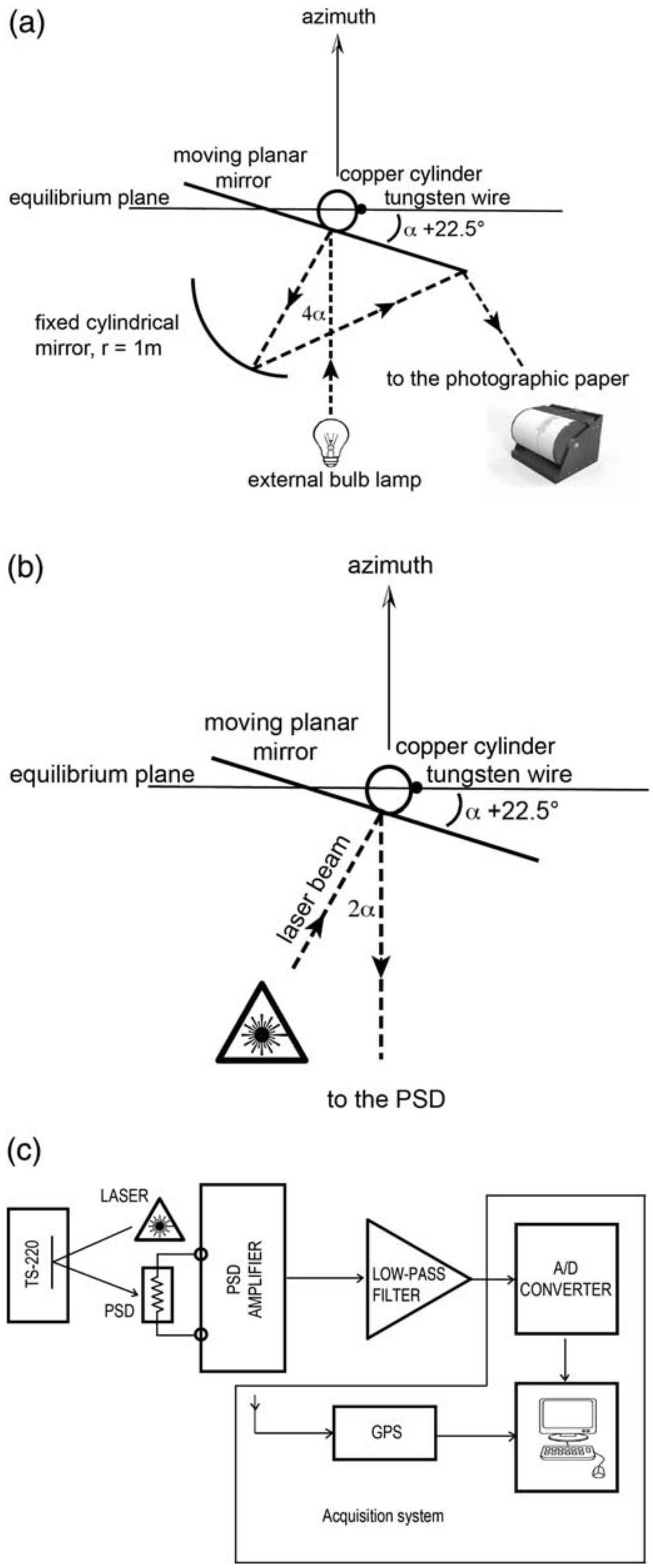



▲ **Figure 1.** (a) Original Wood–Anderson layout, (b) refurbished Wood–Anderson layout, and (c) recording and acquisition system layout. (GPS, Global Positioning System; PSD, position-sensing detector.)

A light beam coming from an external bulb lamp (Fig. 1a) was projected to the planar mirror mounted on the seismometer wire, from which it was reflected into a cylindrical mirror and then back again to the planar mirror. The beam was then focused to a point on photosensitive paper placed over a rotating drum located 1 m from the sensor (Fig. 1a). This arrangement secured a great static magnification in a limited space and made the effective length of the optical lever arm equal to four times the physical distance from the instrument to the recording system.

## The New Optical Leverage

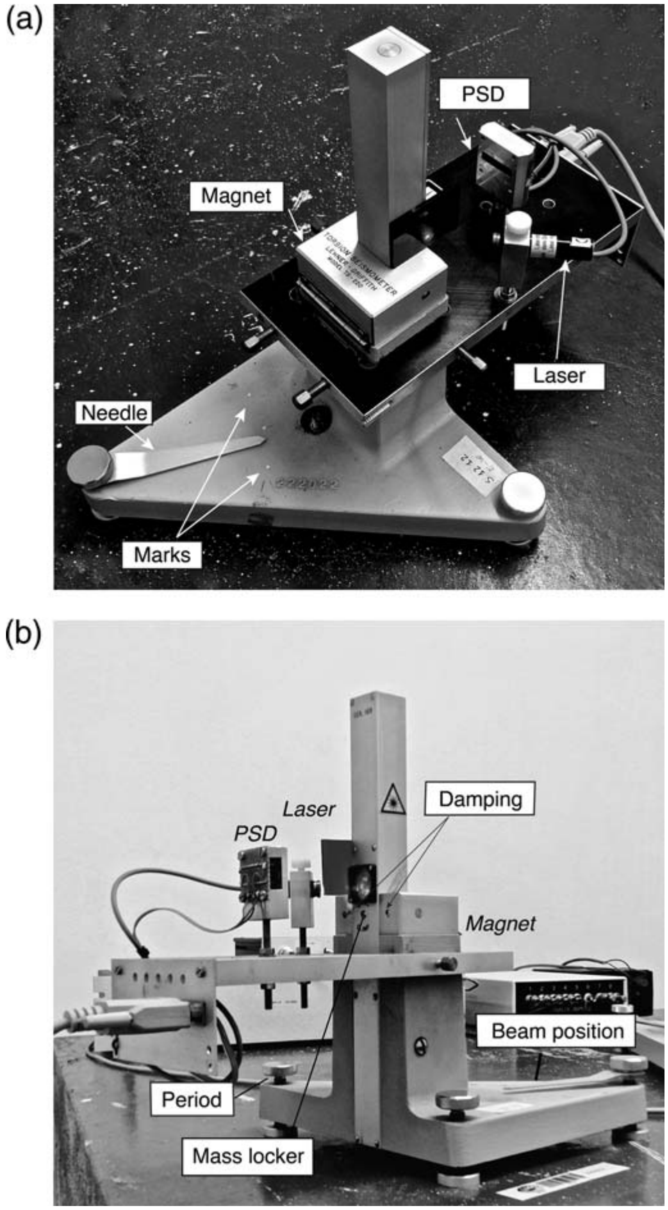
The WA seismograph of Trieste was modernized in 2002, mainly by changing the recording system from analog to digital format (Gentile, 2002). The WA seismograph was upgraded by replacing (1) the bulb lamp, used as the light source in the original arrangement (Fig. 1a), with a laser device (Fig. 1b) and (2) the photosensitive paper with a position-sensing detector (PSD). The PSD is an optoelectronic device that is sensitive to visible red light; it converts an incident light spot into a tension directly proportional to the barycenter of the spotlight. The departure from linearity is ∼0.1% when the light beam spot lies inside 80% of the sensor surface. In the new arrangement (Fig. 1b), the cylindrical mirror and the two lenses placed in the optical window were removed from the TS-220 WA seismometer, leaving a small side window. The laser beam enters from the side window, hits the moving mirror, and then is reflected back to the PSD (Fig. 1b). In this new configuration, the optical arm is twice the distance between the mirror and the sensor. The analog signal from the PSD goes to an amplifier (Fig. 1c) and, after filtering with a 40 Hz antialiasing filter, is acquired through a 16-bit digital converter with a sampling rate of 100 samples/s.

## Calibration

Following the refurbishment, in 2002 the TS-220 instruments were disassembled, cleaned, and completely recalibrated according to the maintenance manual.

Some adjustment screws located on the base of the instrument housing (Fig. 2a,b) allowed the instrument to tilt, thus affecting the gravity component acting on the mass. Changing the tilt enabled us to adjust the seismometer's natural period. The value of the damping was recalibrated to obtain a value that is as close as possible to the theoretical damping ($h = 0.8$). The period and the damping values obtained after the 2002 calibration of the TS-220 are reported in Table 1.

The trace amplitude recorded by the new digital acquisition system of the TS-220 is given in counts. To compute the factor to convert counts into millimeters, we manually moved a needle between two marks located on the base of the instrument case and read the corresponding counts (Fig. 2a). We repeated this operation several times to obtain a mean value of counts corresponding to the needle movements. In the original WA, moving the needle between the two marks corresponded to a 100 mm displacement on the photographic paper. Therefore, the conversion factor is computed as the ratio




▲ **Figure 2.** (a) Top view and (b) side view of the Wood–Anderson seismometer currently operating in Trieste. The screws for tuning the beam position, the period and the damping factor, and for blocking the mass of the instrument are shown in the side view.

between 100 mm and the mean value of the counts, and its error is the standard deviation of the distribution of the peaks in counts.

### Static Magnification

Anderson and Wood (1925) determined a magnification value of 2800 for their instrument, but in doing so they neglected the deformation of the taut tungsten wire from a straight line due to the catenary effect. In fact, the deformation of the wire could be sufficient to reduce the magnification by a factor of 0.3 due to the increase in the polar moment of inertia of the copper cylinder. For this reason, when Uhrhammer and Collins (1990) and Uhrhammer *et al.* (1996) carefully computed the static magnification ($G_s$) of the instrument at Berkeley, California, they found $G_s$ to be equal to $2080 \pm 60$. Using 2800 instead of 2080 in the BB WA simulations leads to a magnitude difference of $+0.129$ (e.g., Uhrhammer *et al.*, 2011).

To verify the $G_s$ value of the Trieste instrument, we adopted two different approaches. The first approach involves a direct action on the instrument. According to Anderson and Wood (1925), $G_s$ is determined by

$$G_s = \frac{A}{a} = \frac{L}{l} = \frac{4\pi^2 A}{gbT_0^2}, \tag{1}$$

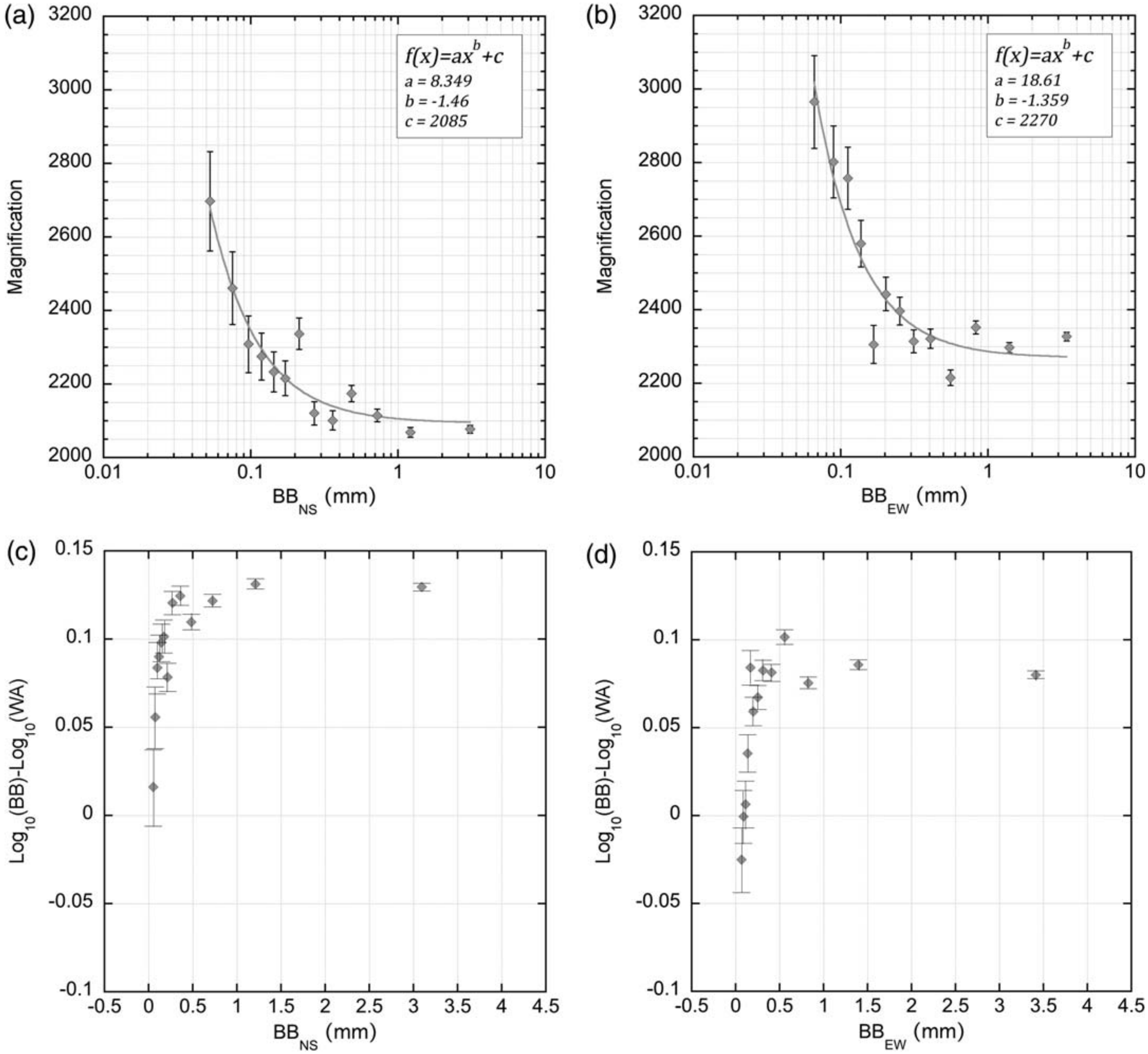
in which $A$ is the seismogram trace amplitude, $a$ is the amplitude of the ground-motion component normal to the equilibrium plane, $l$ is the mass swinging center distance from the rotation axis, $L$ is the optical lever length, $g$ is the gravitational acceleration, $b$ is the instrument tilt angle (in radians), and $T_0$ is the period of oscillation of the instrument (ideally 0.8 s). Tilting the instrument by a known angle $b$ and measuring the output voltage from the PSD, which is proportional to $A$, equation (1) allowed us to calculate $G_s$ (Table 1). The error associated with the estimate of $G_s$ is evaluated considering that the amplifier error is 1% on the linearity of the response, and the accuracy of the voltmeter is equal to 0.05 V.

The second method to check $G_s$ is based upon a comparison of the maximum peak of the seismograms recorded by the WA and Güralp 40-T BB instruments that are placed side by side (see Introduction). Hereinafter we use "amplitude" to mean the trace amplitude in millimeters that we would have obtained with the original recording system. We first sorted the values of the WA amplitudes in ascending order; using a sliding window of 50 WA samples, we calculated the $G_s$ values as the weighted average of the ratios between the WA and BB peak amplitudes (Fig. 3a and 3b, corresponding to north–south and east–west components, respectively). In this test, the BB waveforms were corrected for instrument response to simulate the WA but using a static magnification equal to 1. The weights are given by the inverse of the uncertainty

**Table 1**
**Results of the Wood–Anderson (WA) Calibration after the New Assembly and Parameters Used for the WA $G_s$ Computation**

| Component | Period (s) | Damping | *O* (V) | *A* (mm) | $G_s$ |
|---|---|---|---|---|---|
| North–south | 0.792 | 0.787 | 2.00 ± 0.07 | 45.8 ± 1.6 | 2092 ± 73 |
| East–west | 0.796 | 0.818 | 2.31 ± 0.07 | 52.9 ± 1.8 | 2339 ± 82 |

*O*, PSD controller output (V); *A*, equivalent trace amplitude on paper (mm); $G_s$, static magnification.




▲ **Figure 3.** WA static magnification ($G_s$) versus amplitude of the broadband (BB) seismograms recorded on the (a) north–south (NS) and (b) east–west (EW) components. The error in magnitude estimation due to setting $G_s$ equal to 2800 is plotted as a function of the amplitude of the waveforms recorded on the (c) N-S and (d) E-W components.

on each peak, which is a function of the conversion factor error of the system and the 1% error of the amplifier. We observe that the trends of the values obtained for the north–south component (Fig. 3a) and for the east–west component (Fig. 3b), respectively, are slightly different. This is probably because the east–west component was repaired by the OGS technical staff after a partial detachment of the moving mirror, which could have affected somehow the accuracy of the orientation of the mirror with respect to the wire. However, we observe that the $G_s$ values on both components decrease with increasing amplitude (Fig. 3a,b): for amplitudes $> 1$ mm they tend asymptotically to 2080, which is the value calculated by Uhrhammer and Collins (1990), suggesting the results of Uhrhammer and Collins are only valid for larger amplitudes. For an amplitude value of 0.05 mm, $G_s$ is close to 2800. The value obtained for large amplitudes is very similar to that obtained by the first method (see Table 1). Figure 3c,d show the difference between the equivalent $M_L$ estimated by a BB record with $G_s$ equal to 2800 and the $M_L$ estimated by the WA seismograph. Fixing $G_s$ equal to 2800, we compute an error in magnitude that, depending on the recorded peak amplitudes, spans from 0 for smaller amplitudes to 0.13 for amplitudes $> 0.05$ mm.

## THE NEW CATALOG OF WA $M_L$

Finetti and Morelli (1972) were the first to compute $M_L$ from the WA of Trieste following the Richter (1935) definition:

$$M_L = \log A - \log A_0, \tag{2}$$

in which $A$ is the peak trace amplitude (mm) and $A_0$ depends only on the epicentral distance of the station. As a correction factor for the epicentral distance, they used the values of $-\log A_0$ determined by Richter (1935, 1958) for shallow earthquakes in southern California with epicentral distances below 600 km. Gasperini (2002) determined the $-\log A_0$ attenuation function for all of Italy, showing that it was not significantly different from that estimated by Richter (1935) for southern California, at least in the far field. In the near field, at distances $< 200$ km, the inversion by Gasperini (2002) was not well constrained due to the data scarcity. Bragato and Tento (2005) calibrated a local magnitude scale for northeastern Italy in the hypocentral distance range of 10–250 km, providing a new attenuation function that is in good agreement with that reported by Gasperini (2002) in the range 180–250 km and more generally with that reported by Richter (1935). Bragato and Tento (2005) claimed that no station correction is needed for the TRI station at the bottom of the cave because, according to the station residuals, it behaves as a bedrock-sited station. On the basis of the above pieces of evidence and for consistency with past studies, we employed the Richter (1958) attenuation curve with the extension by Finetti and Morelli (1972) out to 1000 km in revaluating the WA $M_L$.

Since 1971 the Trieste WA $M_L$ was regularly listed in seismological bulletins issued monthly up to 1981, twice a month from 1982 to 1983, and fortnightly from 1984 to 13 September 1989, when the publication of the WA $M_L$ in the bulletins stopped. In 2002, after the upgrade of the WA, we started to compile a Trieste WA earthquake catalog for internal use. For each event, the epicentral distance (estimated whenever possible by the time lag between wave arrivals) and a suite of magnitudes (including $M_L$) were reported. The hypocenter location was reported only for a small fraction of events.

In the catalog compiled in this study, we listed only the earthquakes to which it was possible to associate a hypocentral location retrieved from Italian and international seismological bulletins. The bulletins issued yearly by the Seismological Research Centre of OGS (*The Friuli Venezia Giulia Seismometric Network Bulletin*; OGS, no date) since 1977 were our first reference. These bulletins contain information on the earthquakes that have occurred in northestern Italy since 1977—one year after the 1976 $M_s$ 6.5 Friuli earthquake (e.g., Aoudia *et al.*, 2000; Carulli and Slejko, 2005)—when OGS began installing a seismographic network that today covers a large part of northeastern Italy (Bragato *et al.*, 2011). Over time, the earthquake locations have undergone a number of revisions (e.g., Renner, 1995) to guarantee the best quality solutions. For earthquakes that occurred outside the area monitored by OGS, we took into account the catalog of the Istituto Nazionale di Geofisica e Vulcanologia (ISIDe Working Group, 2010) and the online bulletins of the European-Mediterranean Seismological Centre (2010; *Euro-Med Bulletin*) and of the International Seismological Centre (ISC, 2013; *ISC Bulletin*), which also included contributions from other agencies such as the International Seismological Survey of the Republic of Croatia, the Environmental Agency of the Republic of Slovenia, and the U.S. Geological Survey National Earthquake Information Center. On the basis of how the $M_L$ was computed at the time, we can identify two different periods of data analysis methods as detailed below.

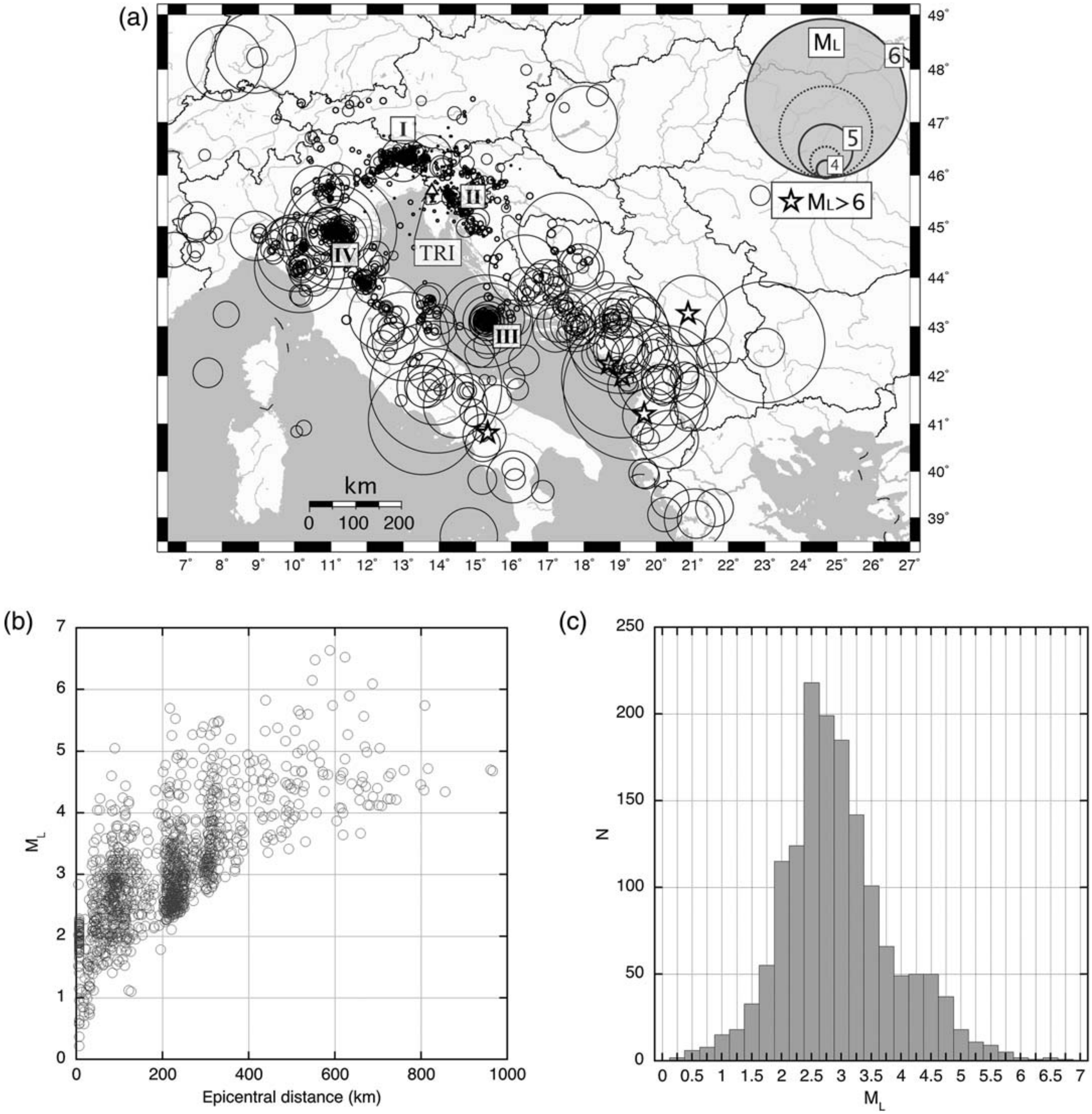
In the first period (1971–1992), the $M_L$ was computed by combining the horizontal components vectorially. However, following the Richter (1958) definition, when using recordings of both horizontal components, it is preferable to compute the magnitude independently from each component and then to average the two determinations, because the maximum motion does not represent the same wave on the two components, and the maxima even may occur at different times on the two components.

To recompute the $M_L$ following the Richter recommendations, we recovered, the OGS archive traces of 83 earthquakes recorded on photographic sheets. Four seismologists (three of whom were part of the seismological staff from the 1970s to the 1990s) reread the peak amplitudes of the two components of the WA waveforms and computed their own $M_L$ values, which were subsequently averaged to retrieve the final $M_L$ for each event. The standard deviation of the residuals between the values of the magnitude obtained by the seismologists and the mean $M_L$ is equal to 0.1. The recomputed $M_L$ values ($M_{LOK}$) were correlated with the $M_L$ originally computed ($M_{LKO}$) through orthogonal regression (Castellaro *et al.*, 2006; Gasperini *et al.*, 2013). The variance ratio is assumed to be 1, because the estimated uncertainty for both $M_{LOK}$ and $M_{LKO}$ was set to 0.1:

$$M_{LKO} = 0.97(\pm 0.01)M_{LOK} + 0.29(\pm 0.05) \tag{3}$$

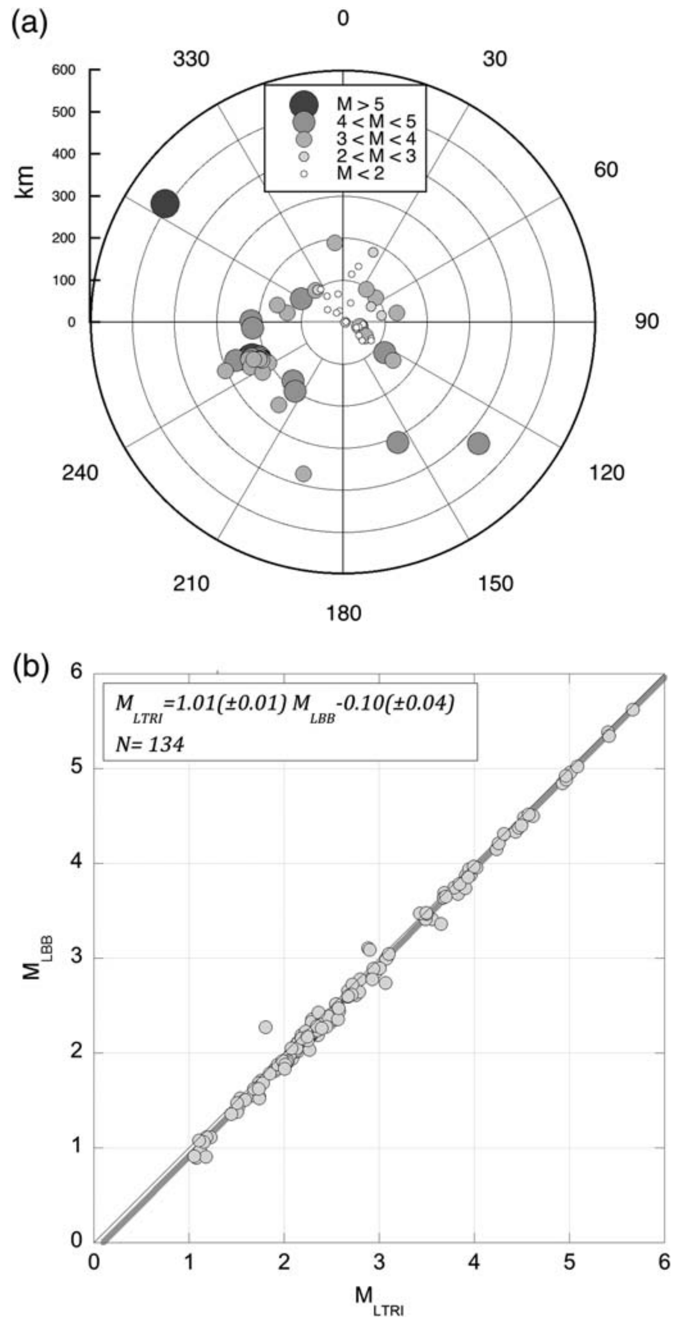
To verify if, from a statistical point of view, there is a linear relationship between $M_{LOK}$ and $M_{LKO}$ we applied the Student's $t$-test (see Davis, 1973). In particular, we tested if the slope can be considered equivalent to 1 (the null hypothesis) or if it is significantly different from 1 (the alternative hypothesis). We found that, in this case, the null hypothesis cannot be rejected at the 0.05 level ($t = 1.49$; the cumulative probability that $t_{teor} \leq 1.49$ via the two-tailed test is 0.86); in other words, we can accept a linear relation between $M_{LOK}$ and $M_{LKO}$, and we compute the difference between the two data sets as the mean difference between $M_{LKO}$ and $M_{LOK}$, which is 0.2. This value is in agreement with Gasperini (2002), who speculated that the original $M_L$ computed as the vectorial sum of the two horizontal components ($M_{LKO}$) overestimated the magnitude by ~0.2 units with respect to the magnitude computed following the Richter recommendation ($M_{LOK}$).

Applying equation (3), we recomputed the magnitudes of the first period without rereading all of the WA seismograms. We re-evaluated the $M_L$ of 370 seismic events and included them in the new catalog.




▲ **Figure 4.** (a) Epicenters of the earthquakes (circles proportional to $M_L$) recorded by the Trieste WA (TRI; white triangle) and investigated in this study. The zones of clustered events are I, the Friuli region; II, Dinaric region; III, Adriatic Sea; and IV, Emilia region. $M_L$ is shown as (b) a function of epicentral distance and (c) a histogram of $M_L$ with bins sized at 0.25 magnitude units.

In the second period (2002–2005 and 2010–2013), $M_L$ was estimated using data recorded by the modernized WA seismometer ($M_{Ld}$). There are 1152 earthquakes for which a location was found in the catalogs: for 956 of them, it was possible to also calculate the equivalent WA magnitude ($M_{LBB}$) by using the Güralp 40-T BB seismometer that, since 22 October 2004, was placed very close to the WA (see Introduction). In addition, for 134 events recorded during the period 2010–2013, the equivalent $M_L$ was estimated both by the BB instruments placed on the surface ($M_{LBB}$) and at the bottom ($M_{LTRI}$) of the cave (see Introduction). To compute the equivalent $M_L$, we first deconvolved the BB instrument transfer function to obtain a ground displacement record, and then we convolved the signal with the WA transfer function.




▲ **Figure 5.** $M_{\mathrm{LTRI}}$–$M_{\mathrm{LBB}}$ data set: (a) azimuth versus distance of considered earthquakes with respect to the Trieste WA seismometer. Circle sizes are proportional to the earthquake magnitudes, and the maximum distance is 600 km. (b) Orthogonal regression fit (thick gray line). $N$ is the number of data.

From 26 May 2005 to 5 March 2010, the WA and BB instruments were temporarily moved from their historical site due to the restoration of the building in which they were housed. The recordings of that period were not included in this study, because the temporary location was not on hard rock.

Over the two periods, we have compiled a database of 1522 earthquakes that occurred between 1977 and 2013 with $0.2 < M_{\mathrm{L}} < 6.6$. The location map (Fig. 4a) highlights four main clusters, as confirmed by the distribution of the earthquakes versus the epicentral distances (Fig. 4b). The events of the first period are mainly located in Friuli (I in Fig. 4a) and northwestern Dinarides (II in Fig. 4a), with nearly half of the epicentral distances between 60 and 100 km. The cluster visible in the central Adriatic Sea (III in Fig. 4a) mainly refers to a seismic sequence that occurred in March 2003 (during the second period of digitized WA data), whereas 25% of the events of the second period, with epicentral distances in the range between 200 and 240 km, belong to the 2012 Emilia sequence (IV in Fig. 4a) following the 20 May 2012 ($M_{\mathrm{w}}$ 6.1) and 29 May 2012 ($M_{\mathrm{w}}$ 5.8) events (e.g., Saraò and Peruzza, 2012). Regarding the $M_{\mathrm{L}}$ distribution (Fig. 4c), most of the earthquakes have magnitude in the range 2.5–3.5, and there are 40 events with $M_{\mathrm{L}} \geq 5$.

## Comparison of $M_{\mathrm{Ld}}$ with Equivalent $M_{\mathrm{L}}$ from BB Records

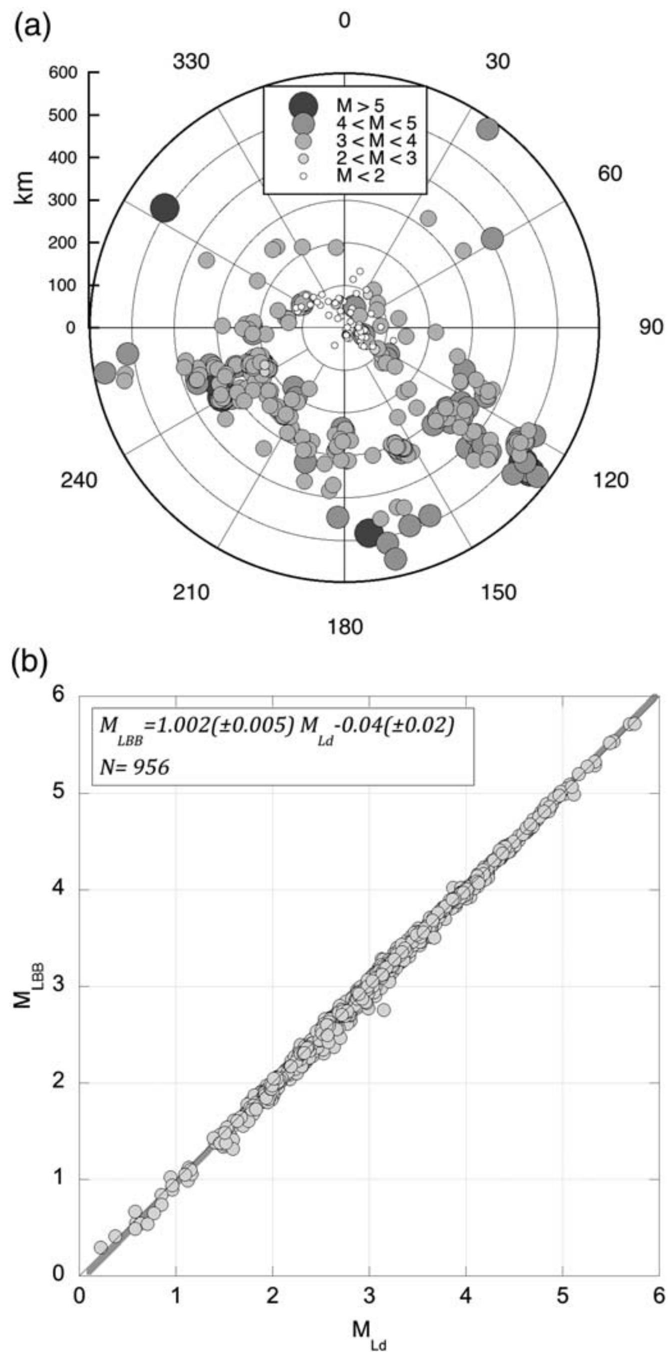
We compared the $M_{\mathrm{L}}$ with the equivalent $M_{\mathrm{L}}$ recovered from the records of the STS-1 installed at the bottom of the cave ($M_{\mathrm{LRTI}}$) and that of the Güralp 40-T ($M_{\mathrm{LBB}}$). Both of the BB are placed on the limestone rock of the Karst plateau but with a difference in elevation of 175 m. Because the STS-1 ($M_{\mathrm{LTRI}}$) is installed in a bedrock site (Bragato and Tento, 2005), we can document the effect of the different instrument elevations on the magnitude computation. To this end, we first compared $M_{\mathrm{LTRI}}$ and $M_{\mathrm{LBB}}$, in the range of 1.2–5.5, calculated for the 134 events that occurred during the period 2010–2013 (Fig. 5a). Epicentral distances $<250$ km are well represented, whereas the azimuthal distribution shows a gap of data in the eastern sector for distances $>150$ km (Fig. 5a).

For $M_{\mathrm{LTRI}}$, we fixed the uncertainty to 0.1 as estimated by Gentili *et al.* (2011). Because both magnitudes are calculated in the same manner, we assumed the same uncertainty for $M_{\mathrm{LBB}}$. The orthogonal fit supplies the following results (Fig. 5b):

$$M_{\mathrm{LTRI}} = 1.01(\pm 0.01)M_{\mathrm{LBB}} - 0.10(\pm 0.04). \tag{4}$$

The Student $t$-test confirms that the null hypothesis (slope = 1) cannot be rejected at the 0.05 level ($t = 1.39$; the cumulative probability that $t_{\mathrm{teor}} \leq 1.39$ is 0.83 via the two-tailed test). Therefore, if we assume the slope is 1, we can say that the bias between $M_{\mathrm{LBB}}$ and $M_{\mathrm{LTRI}}$, obtained as the mean difference, is 0.07. Such a bias can be due to the different elevations of the two seismometers, because one is located at the bottom and one on the top of the cave.

For the comparison with $M_{\mathrm{Ld}}$, we chose $M_{\mathrm{LBB}}$, for which we have a large data set available. We considered 956 events with $0.2 < M_{\mathrm{Ld}} < 5.8$ (Fig. 6a) located within 1000 km from the Trieste station. To compute the $M_{\mathrm{LBB}}$, we used a variable $G_{\mathrm{s}}$ based on our findings described earlier (see the Static Magnification section and Fig. 2). The average uncertainty on $M_{\mathrm{Ld}}$ was estimated considering an error of 1% in the value of the recorded amplitudes due to the amplifier, an error of 2% in the voltmeter accuracy during the instrument calibration, and a bias of 0.02 mm, which is the standard deviation of the dis-




▲ **Figure 6.** $M_{\mathrm{LBB}}$–$M_{\mathrm{Ld}}$ data set: (a) azimuth versus distance of considered earthquakes with respect to the Trieste WA seismometer. Circle sizes are proportional to the earthquake magnitudes, and the maximum distance is 600 km. (b) Orthogonal regression fit (thick gray line). *N* is the number of data.

tribution of amplitudes used for the estimate of the conversion factor. The obtained equation (Fig. 6b) is:

$$M_{\mathrm{LBB}} = 1.002(\pm 0.005)M_{\mathrm{Ld}} - 0.04(\pm 0.02). \qquad (5)$$

The Student's *t*-test, ($t = 0.86$; the cumulative probability that $t_{\mathrm{teor}} \leq 0.86$ is 0.61) says that the null hypothesis cannot be rejected at the 0.05 level and that $M_{\mathrm{LBB}}$ and $M_{\mathrm{Ld}}$ can be really considered as equivalent, thus validating our procedure for computing the equivalent $M_{\mathrm{L}}$.

### Wood–Anderson versus Moment Magnitude $M_{\mathrm{w}}$

Both $M_{\mathrm{L}}$ and seismic moment $M_0$ or, equivalently, moment magnitude $M_{\mathrm{w}}$ are in principle, measures of basic properties of the earthquake source over the entire range for which $M_{\mathrm{L}}$ can be determined (e.g., Hanks and Kanamori, 1979; Deichmann, 2006). In practice deviations of $M_{\mathrm{L}}$ with respect to $M_{\mathrm{w}}$ are very often observed, and Deichmann (2006) comprehensively discussed the reasons for such discrepancies as related to source and rupture complexities, to radiation pattern, to instruments or to path effects. He also showed that some specific strategies are needed to quantify such effects and properly interpret the systematic deviations of $M_{\mathrm{L}}$ relative to $M_{\mathrm{w}}$ (Deichmann, 2006).

To investigate the relationship between $M_{\mathrm{L}}$ and $M_{\mathrm{w}}$ in our case, we compared the WA $M_{\mathrm{L}}$ estimates ($M_{\mathrm{L}}$ hereafter) with the $M_{\mathrm{w}}$ values available in the literature; the Regional Centroid Moment Tensor (RCMT) catalog (Pondrelli *et al.*, 2006), the Global Centroid Moment Tensor (Global CMT) catalog (Dziewonski *et al.*, 1981; Ekström *et al.*, 2012), and the northeast Italy catalog (Saraò, 2008; Saraò and Centro Ricerche Sismologiche Staff, 2013).

For 183 events with azimuthal distribution $100° < \mathrm{Az} < 260°$ and in the magnitude range $3.5 < M_{\mathrm{w}} < 6.9$, we have both $M_{\mathrm{L}}$ and $M_{\mathrm{w}}$ (Fig. 7a). Four events, identified as outliers according to Chauvenet's criterion (e.g., Taylor, 1997) applied to the population of the residuals between the two magnitude values, were discarded.

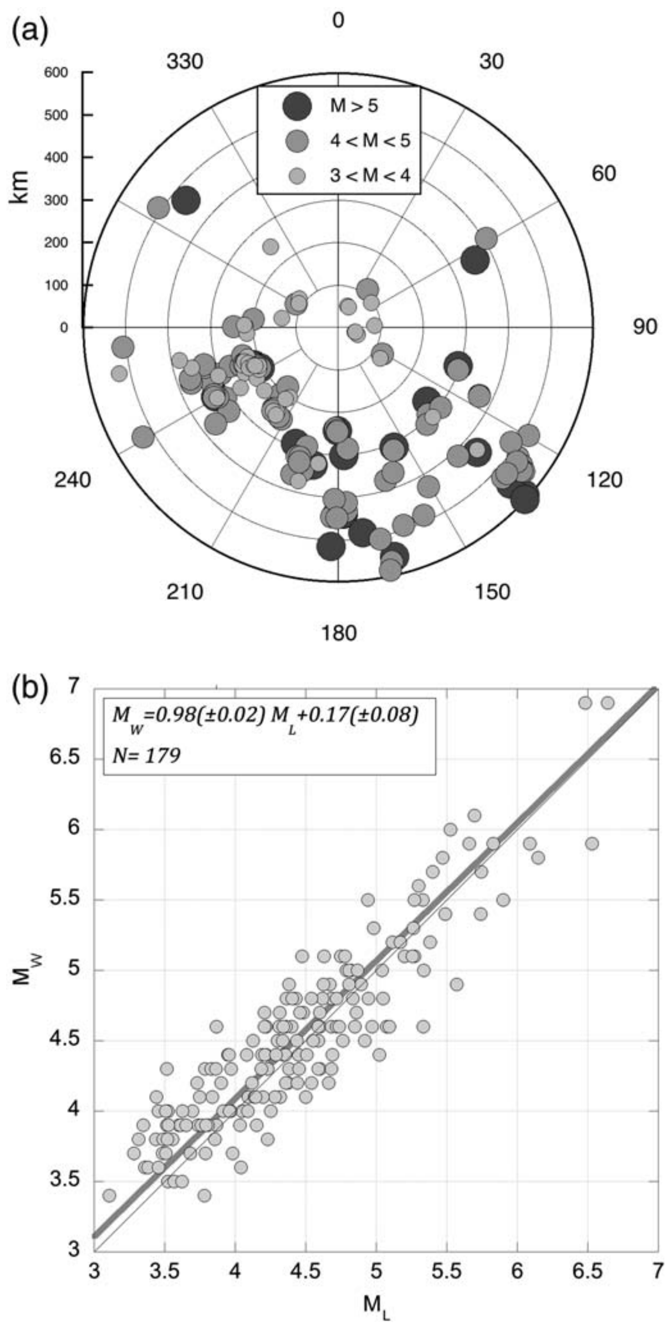
The $M_{\mathrm{w}}$ uncertainty was obtained as the standard deviation of the residuals between the values reported for the same event in the three moment tensor databases mentioned previously. Forty-five events have multiple $M_{\mathrm{w}}$ reports, and we obtained an estimate of the uncertainty equal to 0.07. The orthogonal regression (Fig. 7b) between $M_{\mathrm{L}}$ and $M_{\mathrm{w}}$ gives

$$M_{\mathrm{w}} = 0.98(\pm 0.02)M_{\mathrm{L}} + 0.17(\pm 0.08). \qquad (6)$$

The Student *t*-test, applied to verify the null hypothesis that the slope is 1, provides an absolute value of $t = 2.27$; the cumulative probability that $t_{\mathrm{teor}} \leq 2.27$ using the two-tailed test is 0.98; the null hypothesis can be rejected at the 0.05 level; it is necessary to go down to the 0.02 level to not reject the hypothesis. Gasperini *et al.* (2013) found that $M_{\mathrm{L}}$ scales 1:1 with $M_{\mathrm{w}}$ with a general underestimation of ~0.2 magnitude units in the Italian territory.

## CONCLUSIONS

Two horizontal WA seismometers, installed in 1971, are still operating at the Trieste station (in northeast Italy) after having been restored and modernized in 2002 through the replacement of the recording on photographic paper by an electronic device. The original $M_{\mathrm{L}}$ values related to the Trieste WA were




▲ **Figure 7.** $M_L$–$M_w$ data set: (a) azimuth versus distance of considered earthquakes with respect to the Trieste WA seismometer. Circle sizes are proportional to the earthquake magnitudes, and the maximum distance is 600 km. (b) Orthogonal regression fit (thick gray line). $N$ is the number of data.

published in bulletins up to 1989. We generated a new catalog from digital data after 2002 by taking the locations in different national and international catalogs. After the analysis described in this paper, we carefully revised the magnitudes of the old data to remove the bias of ~0.2 units introduced by past errors in the old bulletins. We compiled a new catalog of 1522 WA $M_L$ values for the time window 1977–2013 for events with magnitude $0.2 < M_L < 6.6$. The new catalog can be downloaded from the Centro Richerche Sismologiche (CRS)–OGS website, and it is also annexed to this article as an Ⓔ electronic supplement.

Other issues that were addressed in our study show that the following:

1. The proper static magnification $G_s$ of the WA depends on the recorded wave amplitudes and approximately follows a power law, ranging from 2800 for amplitudes of 0.05 mm and reaching an asymptotic value of 2080 for amplitudes $> 1$ mm. In this paper, we assumed a variable $G_s$ to obtain the proper equivalent $M_L$ using the data from the broadband instruments.
2. The $M_L$ estimates computed by the simulated WA seismograms recorded at the top of Grotta Gigante have a constant bias with respect to the broadband instruments located at the bottom of the cave.
3. The relationship found between the $M_L$ and the $M_w$ shows a general underestimation of ~0.2 magnitude units for $M_L$.

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