

Evidence for Non-Self-Similarity of Microearthquakes Recorded at a Taiwan Borehole Seismometer Array

Journal:	<i>Geophysical Journal International</i>
Manuscript ID	GJI-S-15-0748.R2
Manuscript Type:	Research Paper
Date Submitted by the Author:	n/a
Complete List of Authors:	Lin, Yen-Yu; National Central University, Department of Earth Sciences; Academia Sinica, Institute of Earth Sciences Ma, Kuo-Fong; National Central University, Department of Earth Sciences; Academia Sinica, Institute of Earth Sciences Kanamori, Hiroo; California Institute of Technology, Division of Geological and Planetary Sciences Song, Teh-Ru; University College London, Department of Earth Sciences Lapusta, Nadia; California Institute of Technology, Division of Geological and Planetary Sciences; California Institute of Technology, Division of Engineering and Applied Science Tsai, Victor; California Institute of Technology, Division of Geological and Planetary Sciences
Keywords:	Earthquake source observations < SEISMOLOGY, Seismic attenuation < SEISMOLOGY, Earthquake dynamics < SEISMOLOGY

Evidence for Non-Self-Similarity of Microearthquakes Recorded at Taiwan Borehole Seismometer Array

1
2
3
4 1
5
6

7 2
8
9

10 3
11
12

13 4
14
15
16

17 5 Evidence for Non-Self-Similarity of Microearthquakes

18 6 Recorded at a Taiwan Borehole Seismometer Array

19
20
21 7 Yen-Yu Lin^{1,2*}, Kuo-Fong Ma^{1,2}, Hiroo Kanamori³, Teh-Ru Alex Song⁴, Nadia

22
23
24 8 Lapusta^{3,5}, and Victor C. Tsai³

25
26
27
28 9 ¹Department of Earth Sciences, National Central University, Taiwan

29
30 10 ²Institute of Earth Sciences, Academia Sinica, Taiwan

31
32 11 ³Division of Geological and Planetary Sciences, California Institute of Technology, USA

33
34 12 ⁴Department of Earth Sciences, University College London, London, UK

35
36 13 ⁵Division of Engineering and Applied Science, California Institute of Technology, USA

37
38
39 14
40
41 15
42
43 16

44
45
46
47
48 17 Accepted date

49
50 18 Received date

51
52
53 Corresponding author: Yen-Yu Lin

54 Email: yulinm22@gmail.com

55
56
57 19 Tel: +886-2-27839910#609

58
59
60 20

1 Summary

2 We investigate the relationship between seismic moment M_0 and source duration
3 t_w of microearthquakes by using high-quality seismic data recorded with a vertical
4 borehole array installed in central Taiwan. We apply a waveform cross-correlation
5 method to the three-component records and identify several event clusters with high
6 waveform similarity, with event magnitudes ranging from 0.3 to 2.0. Three
7 clusters—Clusters A, B, and C—contain 11, 8, and 6 events with similar waveforms,
8 respectively. To determine how M_0 scales with t_w , we remove path effects by using a
9 path-averaged Q . The results indicate a nearly constant t_w for events within each
10 cluster, regardless of M_0 , with mean values of t_w being 0.058, 0.056, and 0.034 s for
11 Clusters A, B, and C, respectively. Constant t_w , independent of M_0 , violates the
12 commonly used scaling relation $t_w \propto M_0^{1/3}$. This constant duration may arise either
13 because all events in a cluster are hosted on the same isolated seismogenic patch, or
14 because the events are driven by external factors of constant duration, such as fluid
15 injections into the fault zone. It may also be related to the earthquake nucleation size.

16
17
18 **Keywords:** Earthquake source observations; Seismic attenuation; Earthquake
19 dynamics

1. Introduction

If earthquakes are self-similar, then the source duration t_w (or, equivalently, the corner frequency) should scale with $M_0^{1/3}$, where M_0 is the seismic moment (e.g., Aki 1967). However, several studies have found that some earthquakes have similar source durations regardless of their size. For example, from a study of small earthquakes in Parkfield, California, Harrington & Brodsky (2009) found different seismic moments that correspond to the same minimum source dimension, and Bouchon et al. (2011) made a similar observation for repeating earthquakes with stress drop variation in the earthquake sequence preceding the 1999 Izmit earthquake in Turkey. Lengliné et al. (2014) discovered that microearthquakes induced by a fluid injection test exhibited a large variability in stress drop, and they suggested that the variations may result from fluid pressure at a localized interface that reduce the normal stress. These observations suggest non-self-similar seismic source behavior.

If a minimum rupture size is required for an earthquake to nucleate, as suggested, for example, by rate-and-state friction laws (e.g., Rice & Ruina 1983; Rice 1993; Rubin & Ampuero 2005; Lapusta & Liu 2009), the conventional relation $t_w \propto M_0^{1/3}$, which characterises earthquake self-similarity, may break down at a certain magnitude. For example, events could have sizes dictated by the nucleation length but have different stress drops and hence different moments. However, the minimum rupture

Evidence for Non-Self-Similarity of Microearthquakes Recorded at Taiwan Borehole Seismometer Array

1 size of natural earthquakes has not been directly determined.

2 Although the source scaling relationship $t_w \propto M_0^{1/3}$ has been observed for a
3 wide range of earthquakes (e.g., Allmann & Shearer 2009; Duputel et al. 2013), it has
4 also been a subject of debate, especially for small earthquakes (Abercrombie 1995;
5 Ide et al. 2004; Stork & Ito 2004; Oye et al. 2005; Venkataraman et al. 2006; Yamada
6 et al. 2007). Conventionally, the source duration of an event is determined from the
7 spectral corner frequency by considering a frequency-dependent (Stork & Ito 2004;
8 Oye et al. 2005) or frequency-independent Q -model to remove path effects. It has
9 been suggested that incomplete removal of path effects and an inappropriate
10 instrumental bandwidth affect observed scaling relations (e.g., Ide et al. 2003). To
11 reduce the influence of the path effect on source duration estimates, some studies have
12 deconvolved target event records relative to a small event record (Mori et al. 2003;
13 Ide et al. 2003, 2004; Oye et al. 2005; Venkataraman et al. 2006; Mayeda &
14 Malagnini 2009; Lin et al. 2012). This method is generally called the empirical
15 Green's function (EGF) method.

16 Lin et al. (2012) investigated the seismic activity and scaling relationships of
17 microearthquakes recorded with a vertical borehole seismometer array in Taiwan (Fig.
18 1a). They determined the source parameters, such as the corner frequency and seismic
19 moment, by using a frequency-domain spectral fitting method, Brune's

Evidence for Non-Self-Similarity of Microearthquakes Recorded at Taiwan Borehole Seismometer Array

1
2
3
4 1 omega-squared source model (Brune 1970), and a frequency-independent Q -model.
5
6
7 2 Source dimensions were estimated from corner frequencies. They found that in the
8
9
10 3 magnitude range from M_w 0.0 to 2.0, source dimensions were essentially constant and
11
12
13 4 stress drop increased with magnitude. They also analyzed 14 events from several
14
15
16 5 seismic clusters by using the EGF method to remove path effects. The results
17
18
19 6 supported the conclusion obtained using the spectral fitting method. However, some
20
21
22 7 ambiguity in the determined microearthquake scaling relations still remained because
23
24
25 8 of the difficulty in removing path effects completely and the limited frequency
26
27
28 9 bandwidth of the instruments.
29
30
31

32 10 In the present study, we extend the study of Lin et al. (2012) by applying a
33
34
35 11 waveform cross-correlation method to three-component records. (In the study of Lin
36
37
38 12 et al. (2012), only the vertical component was used.) The consideration of all three
39
40
41 13 components enables us to use S-wave information to define the similarity of
42
43
44 14 waveforms accurately, thus allowing us to include more events in clusters with similar
45
46
47 15 waveforms. In particular, we identify three clusters—Clusters A, B, and
48
49
50 16 C—containing 11, 8, and 6 events with similar waveforms, respectively. The
51
52
53 17 consideration of all three components results in a considerably larger database.
54
55
56 18 Moreover, we use several methods to assess the path effects for determining upper
57
58
59 19 and lower bounds on source duration without considering the omega-squared source
60

Evidence for Non-Self-Similarity of Microearthquakes Recorded at Taiwan Borehole Seismometer Array

1 model whereas Lin et al. (2012) used the omega-squared source model as a reference
2 spectrum. Although the omega-squared source model has been established as a useful
3 reference model for small earthquakes, we do not apply it here because we do not
4 want our conclusion to be influenced by the use of a particular source model. We
5 focus on Clusters A, B, and C and examine the scaling problem for small earthquakes
6 by considering a set of well-constrained source parameters.

7 We find that source duration is nearly constant within each cluster, regardless of
8 event magnitude, with mean values of the source duration being 0.058, 0.056, and
9 0.034 s for Clusters A, B, and C, respectively. The nearly constant durations,
10 independent of M_0 , violates the commonly used scaling relation $t_w \propto M_0^{1/3}$.

12 2. Instrumentation

13 The borehole data used in this study were recorded by a 7-level
14 three-component vertical borehole seismic array (TCDPBHS) installed in Hole-A of
15 the Taiwan Chelungpu-Fault Drilling Project (TCDP) in July 2006 (Ma et al. 2012;
16 Lin et al. 2012; Lin 2014). This array covers a depth range from 946 to 1274 m at
17 intervals of 50–60 m (from the top to the bottom, the seismometers are labelled BHS1
18 to BHS7; Fig. 1b). Hole-A is a 2-km-deep hole that crosses the main fault of the 1999
19 M_w 7.6 Chi-Chi earthquake (Ma et al. 2006) at a depth of 1111 m. The sensors are

Evidence for Non-Self-Similarity of Microearthquakes Recorded at Taiwan Borehole Seismometer Array

1 velocity-type short-period seismometers with a natural frequency of approximately
2 4.5 Hz, a damping of approximately 29%, and a sensitivity of 1.6 V/(cm/s). Fig. 2(a)
3 shows the average instrument response curve for the 21 sensors at the 7 stations; the
4 curve indicates that the instruments have the capability to record seismic signals
5 higher than 4 Hz. The recording gain is 100. The instrument response was removed
6 from each record, and corrections for the Galperin angle and orientation of the
7 three-component sensors were performed systematically before waveform analysis
8 (Lin et al. 2012; Lin 2014). The sampling rate was set at 1000 samples/s for detecting
9 microearthquakes with high frequency signals up to the end of 2007.

11 3. Data

12 To identify seismic clusters, we employ an improved version of the detection
13 method used by Lin et al. (2012). We search for event clusters by cross-correlating
14 over 3278 local microearthquakes that were manually selected by Lin et al. (2012);
15 the difference in the arrival time between the S wave and the P wave is less than or
16 equal to 2.0 s for the microearthquakes. The microearthquakes occurred between
17 November 2006 and December 2007. To compare waveforms of the three-component
18 BHS4 records, we use a 6-s time window from 1 s before to 5 s after the P-wave
19 arrival time. Compared with the study of Lin et al. (2012), which considered P-wave

Evidence for Non-Self-Similarity of Microearthquakes Recorded at Taiwan Borehole Seismometer Array

1 similarity with only vertical component data, we perform a more comprehensive
2 comparison: we compare both P and S waves with vertical and both horizontal
3 components. For cross-correlation analyses, we use a band-pass filter (10–50 Hz) to
4 capture the dominant frequency band of microearthquakes ($M \leq 2.0$). We group the
5 events into a cluster if the cross-correlation coefficient CC of their three-component
6 waveforms is greater than 0.8. We identify a total of 130 seismic clusters, with each
7 cluster having 2 to 11 events.

8 We estimate the magnitudes of these events from maximum S-wave vectorial
9 amplitudes (A in cm/s) by using the empirical relationship $M_{we} = 0.67 \log A + 3.48$
10 (Fig. 2b). This relationship was obtained using the 242 located events of the 3278
11 events identified by Lin et al. (2012), where the empirical moment magnitude M_{we}
12 was estimated using spectral analysis and the S-wave amplitudes of the 242 located
13 events (Lin et al. 2012). The two-sigma variation of M_{we} is approximately ± 0.3 .

14 Among the clusters, seven clusters comprise more than four events (Fig. 2c), and
15 three of them contain events with a magnitude range greater than 1. We label these
16 three clusters A, B, and C. The other clusters are labelled D to G. We focus on
17 Clusters A, B, and C because they have a large magnitude range of 0.27–1.97,
18 0.49–1.62, and 0.35–1.32, respectively (Table 1, Fig. 2c).

19 To locate these clusters, we stack records of the events in each cluster to

Evidence for Non-Self-Similarity of Microearthquakes Recorded at Taiwan Borehole Seismometer Array

1
2
3
4 1 improve the signal-to-noise (S/N) ratio and then use the P-wave polarization angles
5
6
7 2 determined from particle motions and the difference in the arrival time between the S
8
9
10 3 wave and the P wave. This method was also used to locate microearthquakes using
11
12
13 4 records of the vertical borehole array (Oye & Roth 2003; Lin et al. 2012). We assume
14
15
16 5 a laterally homogeneous velocity structure estimated from a 3D velocity model for
17
18
19 6 this area of Taiwan (Kim et al. 2005). Since the ray paths to the borehole stations
20
21
22 7 avoid very-low-velocity shallow structures, we do not expect large velocity variations
23
24
25 8 over the study area, which extends over a few kilometres (Lin et al. 2012). Fig. 1(a)
26
27
28 9 shows the distribution of Clusters A to G together with the 242 microearthquakes
29
30
31 10 identified by Lin et al. (2012). These event clusters have similar locations to the
32
33
34 11 background seismicity in the depth range of approximately 10–15 km and they are
35
36
37 12 within the deformation zone of a fold-and-thrust system along the decollement (Lin et
38
39
40 13 al. 2012).

44 14 Focal mechanisms of these event clusters are difficult to determine because of
45
46
47 15 the lack of useful recordings from surface stations. Only one event in Cluster A, with
48
49
50 16 M_{we} 1.51, was well recorded at surface stations (Fig. 1a). Fig. 1(c) shows the
51
52
53 17 first-motion focal mechanism determined using the method described by Reasenber
54
55
56 18 & Oppenheimer (1985). The focal mechanism has two nodal planes with a strike of
57
58
59 19 350/225, a dip of 35/68, and a rake of $-140/-62$; the slip vector suggests oblique
60

1 motion.

2

3 **4. Similar P-wave durations of events in each cluster**

4 We can visually observe that the P-wave duration within each event cluster is

5 essentially constant. Fig. 3 shows waveform comparisons between the smallest (M_{we}

6 = 0.27) and the largest ($M_{we} = 1.97$) events in Cluster A at stations BHS1 and BHS4.

7 Both P and S waves show high similarity at both stations, regardless of the event

8 magnitude. We use the original records after instrument and orientation corrections,

9 and a notch filter is applied over the frequency band 58–62 Hz to remove strong

10 electronic noise. Fig. A1 shows examples of the P-wave recordings of events in

11 Clusters A, B, and C over the stations BHS1-BHS7. Since the records at BHS4 are of

12 higher quality (Lin et al. 2012) than those at other borehole stations, we use the

13 waveforms at BHS4 in the following analysis. We focus on P-wave duration because

14 the waveforms of P waves are simpler than those of S waves.

15 The vertical velocity records at BHS4 show distinct constant P durations in

16 Clusters A, B, and C (Figs. 4a–4c). We measure P-wave duration (P_{dur}^{vel}) by identifying

17 zero crossings of the first cycle of the P phase, as illustrated in Fig. 4(a). The

18 parameter P_{dur}^{vel} of Clusters A, B, and C is 0.086, 0.081, and 0.053 s, respectively, and

19 the dominant frequency of the P waves is approximately 10–20 Hz. Fig. 4(e) shows

Evidence for Non-Self-Similarity of Microearthquakes Recorded at Taiwan Borehole Seismometer Array

1 the superposition of event waveforms from Clusters A, B, and C after amplitude
 2 normalization. The superposed waveforms exhibit a consistent initial downward pulse
 3 for all events in each cluster, regardless of event magnitude. The constant P_{dur}^{vel}
 4 observed in these event clusters is in striking contrast with the variable P_{dur}^{vel} for other
 5 events in a similar magnitude range, shown in Fig. 4(d). Fig. 5 compares P_{dur}^{vel} of
 6 Clusters A, B, and C with that of other microearthquakes with similar magnitudes.
 7 The constant P_{dur}^{vel} in Clusters A, B, and C is evident.

8 We determine the difference in the arrival time between the S wave and the P
 9 wave (Δt_{sp}) for events in each cluster by comparing the waveforms of P waves in the
 10 Z component and those of S waves in the N component. The first event in each cluster
 11 is used as a template. We window a 0.4-s record containing P and S waves from the
 12 original seismograms, and determine the P and S arrival times by cross-correlating the
 13 windowed records of the template event with those of other events. With P and S
 14 arrival times thus determined, we compute Δt_{sp} .

15 In general, difference in Δt_{sp} between the events within a cluster is less than
 16 0.02 s, indicating that the maximum distance between events in a cluster is
 17 approximately 160 m. In a medium in which the P-wave speed is 5700 m/s, the
 18 wavelength of a P wave with a dominant frequency of 10 to 20 Hz is 285 to 570 m.
 19 Thus, the difference in distance is smaller than the wavelength of the P wave, and

Evidence for Non-Self-Similarity of Microearthquakes Recorded at Taiwan Borehole Seismometer Array

1 therefore, we consider the events within a cluster to be from approximately the same
2 location and to have similar path effects.

4 **5. Source duration estimates**

5 As shown in Fig. 4, the pulse width of the observed P wave is essentially
6 constant regardless of event magnitude. However, this does not necessarily mean that
7 the width of the source pulse is constant. To investigate this question, we first
8 consider the following two end-member models. 1) In the first end-member model,
9 we assume that the observed waveform represents the source-time function (STF).
10 This model can be used to obtain the longest estimate of the STF. 2) In the second
11 end-member model, we assume that the STF of the smallest event is a delta function
12 $\delta(t)$ and that the observed waveform is due to path effects. This model has the
13 potential to yield the shortest estimate of the STF, at least within the signal resolution.
14 Then, we obtain the most plausible durations by correcting the waveforms for
15 attenuation determined for the region in prior studies. We illustrate our analysis for
16 the events in Cluster A. The same analysis is performed for the events in the other
17 clusters.

18 For the first end-member model, we assume that the path effect is small and that
19 the observed waveforms represent the STFs. The waveforms shown in Fig. 4 are

Evidence for Non-Self-Similarity of Microearthquakes Recorded at Taiwan Borehole Seismometer Array

1 velocity waveforms. Since the source waveform is the moment-rate function (as we
2 shall show later), which corresponds to the displacement waveform, we convert the
3 velocity waveforms in Fig. 4 to displacement waveforms and define their pulse width
4 as follows.

5 First, we integrate the velocity waveforms in Fig. 4, and apply a zero-phase
6 band-pass filter with a bandwidth of 5 to 50 Hz. Fig. 6 shows an example of an
7 estimated STF, which typically has a triangular shape. The peak amplitude is denoted
8 by PA, and the two minima on either side of the PA are called the left bottom
9 amplitude (LBA) and right bottom amplitude (RBA). We define the minimum
10 amplitude of the STF as the average of the LBA and RBA. The full amplitude of the
11 STF can be defined as the difference between the PA and the minimum amplitude.
12 The half-amplitude points are shown by dots. The source pulse width t_w is twice the
13 time interval between the two dots. Results for this first end-member are shown in Fig.
14 7. The P-wave pulse width P_{dur}^{dis} thus defined is essentially constant for all the records
15 in each cluster and represents t_w , and is 0.070, 0.065, and 0.043 s for Clusters A, B,
16 and C, respectively.

17 Using the second end-member model, we deconvolve the observed waveforms
18 with that of the smallest event. This method is generally referred to as the EGF
19 method. As shown later, the S/N ratio of the spectra of these events deteriorates at

Evidence for Non-Self-Similarity of Microearthquakes Recorded at Taiwan Borehole Seismometer Array

1 very high frequencies, greater than 50 Hz. Therefore, we filter the deconvolved
2 records using a band-pass filter with a pass band of 5 to 50 Hz. Fig. 8(a) shows the
3 filtered deconvolved signals for Cluster A. We define the pulse width using the
4 method illustrated in Fig. 6. The top trace labelled 'impulse' is the filtered delta
5 function, which is assumed as the STF for the smallest event. With our definition of
6 the pulse width (illustrated in Fig. 6), the width of this pulse is 0.020 s. The
7 deconvolved source pulses for larger events are slightly broader, and they have a
8 width ranging from 0.024 to 0.030 s for the M_{we} range from 0.38 to 1.97 (Fig. 8 and
9 Table 1). In practice, the resolution may be reduced because of the band-pass filtering
10 (5 to 50 Hz) of the deconvolved signal, as illustrated in Appendix B for a triangular
11 STF.

12 The results of the EGF method are consistent with our hypothesis of the
13 near-constant duration of the STFs for the events of different magnitude in each
14 cluster. By assuming that the smallest event is an impulse, or has the duration of 0.020
15 s when filtered between 5 and 50 Hz, we get similar durations for the other events in
16 the cluster of 0.024-0.030 s. These durations are almost constant in contrast to the
17 behavior expected of the commonly assumed scaling of $t_w \propto M_0^{1/3}$, as discussed
18 further in the next section.

19 Given the similarity of the time-domain signals, the broadening of the source

Evidence for Non-Self-Similarity of Microearthquakes Recorded at Taiwan Borehole Seismometer Array

1 pulse widths with respect to the smallest event is somewhat unexpected, but as shown
2 later, the spectral fall-off of larger events is steeper than that of the smallest event over
3 the frequency band 20-50 Hz. This difference in the high-frequency spectral fall-off
4 causes the broadening of the source pulse.

5 The source pulse widths estimated with the two end-member models bound the
6 real source pulse width. Note that the resolution of deconvolution is naturally limited
7 by the S/N ratio and the pass band of the filter. Although the spectral amplitude of the
8 smallest event appears to be higher than the noise spectral amplitude at frequencies
9 lower than 50 Hz, as shown in Fig. 9, we cannot completely rule out that the spectrum
10 of the smallest event is influenced by noise at lower frequencies and, consequently,
11 that the pulse widths of the moment-rate functions obtained by deconvolution are
12 affected. In this case, the lower bound becomes somewhat uncertain. In particular, we
13 cannot rule out the case in which the larger events have no broadening of their STFs
14 and hence also appear impulsive, which would mean that all sources have durations
15 below the data resolution of 0.02 s. However, such a case would require quite large
16 and unusual attenuation, and hence it is unlikely, as shown in the discussion section 6.

17 A plausible real situation is that the path effect is approximately accounted for by
18 attenuation. Although we do not know the exact value of Q_p (Q for P waves) for the
19 path involved, using the Q_p structure of the fault zone estimated from the data

Evidence for Non-Self-Similarity of Microearthquakes Recorded at Taiwan Borehole Seismometer Array

1 obtained by the TCDPBHS (Wang et al. 2012) and 3D Q tomography results for
 2 Taiwan (Wang et al. 2010), the average Q_p value estimated for the path from the
 3 source to the location of BHS4 (at a depth of approximately 1100 m) is approximately
 4 202, which yields a t_p^* (P-wave travel time divided by the path-averaged Q_p) of
 5 0.0117 s (Fig. 10a). Although this value may not strictly apply to our source station
 6 geometry, we use this value to estimate the width of the source pulse. The effect of
 7 attenuation on the waveform can be represented by a Futterman function (Futterman
 8 1962), which is the impulse response of a dissipative medium. As shown in Fig. 10(b),
 9 the effective width of the Futterman function for $t_p^* = 0.0117$ s is smaller than the
 10 width of the observed displacement pulses shown in Fig. 7. In this case, the difference
 11 between the width of the observed pulse and the effective width of the Futterman
 12 function is the approximate width of the source pulse, and we can estimate the width
 13 of the source pulses by deconvolving the observed waveforms with the Futterman
 14 function. As shown in Fig. 11(a), the source pulse width is approximately 0.054 to
 15 0.064 s (table 1) and essentially constant. A smaller or larger t_p^* (i.e., a higher or
 16 lower Q_p , respectively) results in broader or narrower source pulses, respectively.
 17 Although we do not know the exact value of t_p^* for the path, the source pulse widths
 18 obtained are reasonable, as they are between those of the two end-member models.

19 **Estimation of Q_p from P- and S-wave pulse width**

Evidence for Non-Self-Similarity of Microearthquakes Recorded at Taiwan Borehole Seismometer Array

1 Since the effect of attenuation is critical to our conclusion, we make an
 2 additional test to examine the effect of $Q_P(t_p^*)$ by examining the pulse width
 3 difference between P and S waves. Since the observed SH waves are complex,
 4 probably because of the structure near the borehole, we assume that the first pulse of
 5 the SH wave on the T -component carries the path effect, and estimate t_p^* by
 6 comparing its pulse width to that of the P wave on the vertical component.

7 Assuming that the waveform change is solely due to attenuation, we can write

$$8 \quad O_P(t) = S_P(t) * F(t; t_p^*), \quad (1)$$

$$9 \quad O_S(t) = S_S(t) * F(t; t_s^*), \quad (2)$$

10 where $O_P(t)$, $S_P(t)$, and $F(t; t_p^*)$ are the observed waveform, the source time
 11 function, and the Futterman function for the P wave, respectively. The corresponding
 12 functions for the S wave are given with a subscript ‘‘S’’. Using (1) and (2), we can
 13 write the observed S-wave spectrum $\hat{O}_S(f)$ as

$$14 \quad \hat{O}_S(f) = \hat{S}_S(f) \hat{F}(f; t_s^*) = \hat{O}_P(f) \left[\frac{\hat{S}_S(f)}{\hat{S}_P(f)} \right] \left[\frac{\hat{F}(f; t_s^*)}{\hat{F}(f; t_p^*)} \right], \quad (3)$$

15 where

$$16 \quad \hat{F}(f; t_p^*) = \exp(-\pi f t_p^*) \exp(2i f t_p^* \ln |f / f_H|), \quad (4)$$

$$17 \quad \hat{F}(f; t_s^*) = \exp(-\pi f t_s^*) \exp(2i f t_s^* \ln |f / f_H|), \quad (5)$$

18 which are derived from an absorption-band model (e.g., Kanamori & Anderson 1977;
 19 Kanamori & Rivera 2015). In (4) and (5), f_H is the upper bound on the frequency of

Evidence for Non-Self-Similarity of Microearthquakes Recorded at Taiwan Borehole Seismometer Array

1 the absorption band and i is the imaginary unit. Although we do not know the exact
 2 value of f_H , as long as it is sufficiently higher than the frequency band we are
 3 concerned, it is not relevant to the waveform analysis.

4 First, we assume that the source spectrum is the same for the P and S waves

5 except for a constant factor, i.e., $\left[\frac{\hat{S}_S(f)}{\hat{S}_P(f)} \right] = c$ where c is a constant. Then,

6 $\left[\frac{\hat{F}(f; t_S^*)}{\hat{F}(f; t_P^*)} \right]$ can be written as

$$7 \left[\frac{\hat{F}(f; t_S^*)}{\hat{F}(f; t_P^*)} \right] = \exp[-\pi f t_P^* (t_S^* / t_P^* - 1)] \exp[2i f t_P^* (t_S^* / t_P^* - 1) \ln |f / f_H|]. \quad (6)$$

8 Relation (6) suggests that the form of $\left[\frac{\hat{F}(f; t_S^*)}{\hat{F}(f; t_P^*)} \right]$ is the same as that of $\hat{F}(f; t_P^*)$ in

9 equation (4), with t_P^* replaced by $t_P^* (t_S^* / t_P^* - 1)$. In our analysis, we use $t_S^* / t_P^* = 4$,

10 $V_P/V_S = 1.73$, and $Q_P/Q_S = 2.25$ (Stein & Wysession 2003). Thus, equation (3) leads

11 to

$$12 O_S(t) = c O_P(t) * F(t; 3t_P^*). \quad (7)$$

13 We estimate t_P^* by the following procedure: compute $O_S(t)$ by convolving the

14 observed P-wave pulse and the Futterman function with a given t_P^* , ranging from

15 0.010 to 0.030 s with 0.005 s interval, and compare the pulse width of the computed S

16 pulse with that of the observed SH pulse. The comparison is illustrated in Fig. 12. The

17 results indicate that reasonable values of t_P^* are 0.015~0.020, 0.020, and 0.010 for

Evidence for Non-Self-Similarity of Microearthquakes Recorded at Taiwan Borehole Seismometer Array

1 Clusters A, B, and C, respectively, as shown in Fig. 12. The corresponding Q_p values
 2 are 112~150, 131, and 238 for Clusters A, B, and C, respectively.

3 In the above consideration, we assumed that $\left[\frac{\hat{S}_S(f)}{\hat{S}_P(f)} \right] = c$ which assumes that
 4 the corner frequency of the P wave, f_{cP} , is the same as that of the S wave, f_{cS} , but the
 5 observed ratio, f_{cP} / f_{cS} is commonly around 1.5, probably because of the finiteness
 6 of the source (e.g., Madariaga 1976). Then the P-wave source pulse should be shorter
 7 than S-wave pulse and the difference between the observed pulse widths of the S and
 8 P waves is partly due to the difference in corner frequency, and the effect of
 9 attenuation should be smaller than that estimated with the assumption $f_{cP} = f_{cS}$. Thus,
 10 t_p^* estimated above is the upper bound, and the actual t_p^* can be smaller, although
 11 the difference between them can be small.

13 6. Discussion

14 Scaling of the STF

15 The source duration–seismic moment relations obtained using the three models,
 16 with different assumptions for path effects, consistently show that the source
 17 durations are essentially constant with respect to the seismic moment (Fig. 13), and
 18 hence they do not obey the self-similar empirical relation $t_w \propto M_0^{1/3}$ (e.g., Duputel et

Evidence for Non-Self-Similarity of Microearthquakes Recorded at Taiwan Borehole Seismometer Array

1 al. 2013) shown by dashed curves in Fig. 13. The mean values of the source duration
2 estimated with the correction for Q (dotted lines) are 0.058, 0.056, and 0.034 s for
3 Clusters A, B, and C, respectively. Since these values are based on the attenuation
4 values estimated for the region in prior studies, we use these values as representative
5 solutions in this study. Notably, the scaling relation of Duputel et al. (2013), which
6 was derived using large earthquakes over a magnitude range $M_w = 6.5$ to 9.2, yields a
7 source duration of approximately 0.05 s, which is approximately equal to the source
8 duration of the largest earthquake ($M_w = 1.97$) considered in the current study.

9 With the EGF method, the displacement record of the smallest event (i.e., the
10 EGF) is thought to represent the upper bound of the path effect. Thus, deconvolution
11 of the other events with the smallest event yields the smallest source moment-rate
12 function within the resolution limits determined by the noise level (50 Hz). For
13 example, as shown in Fig. 8(a), the pulse width of the largest event of Cluster A
14 measured after applying the 5–50 Hz band-pass filter is 0.030 s. As shown in Fig. A2,
15 this means that the source duration could be approximately 0.024 s without the filter,
16 if the STF is triangular. The source durations for all other events are approximately
17 the same, with the average being 0.017 s. Such lower bounds for the source duration
18 imply that the stress drop (i.e., stress parameter) cannot be substantially higher than 2
19 MPa (Fig. 13).

Evidence for Non-Self-Similarity of Microearthquakes Recorded at Taiwan Borehole Seismometer Array

1 Based on the EGF analysis alone, we cannot rule out that the source pulse is
 2 much narrower, since the obtained lower bounds of the pulse widths are on the
 3 boundary of the data resolution. Hence we cannot rule out the possibility of the
 4 commonly assumed scaling $t_w \propto M_0^{1/3}$ over a small range of t_w shorter than 0.02 s.
 5 However, such a case is unlikely as it would require much larger attenuation than
 6 determined in the region, as explained next.

7 Suppose that the sources have the scaling $t_w \propto M_0^{1/3}$, with a constant (and high)
 8 stress drop, e.g., 20 MPa. Then, from the scaling relation shown in Fig. 13, t_w ranges
 9 from 0.002 to 0.012 s for M_w in the range of 0.3 to 2.0. For such sources, we compute
 10 the displacement P-pulses at our station for a suite of t_p^* from 0.010 to 0.045 s and
 11 measure the width of the displacement P-pulses thus computed, using the method
 12 illustrated in Fig. 6. Comparing the computed P-pulse widths for various t_p^* to the
 13 observed P-pulse widths shown in Fig. 7, we find that t_p^* of 0.035, 0.030, and 0.015
 14 s is required to match the observations for Clusters A, B, and C, respectively (Fig.
 15 14a). These values of t_p^* are too large compared both with the value inferred from
 16 Wang et al. (2010) and Wang et al. (2012) ($t_p^* = 0.0117$ s) and with the upper bounds
 17 estimated in section 5 from the pulse-width difference between P and S waves ($t_p^* =$
 18 0.015~0.020, 0.020, 0.010 s for Clusters A, B, and C, respectively), even if we allow
 19 for the possibility of uncertainties in the t_p^* measurements. The discrepancy is

Evidence for Non-Self-Similarity of Microearthquakes Recorded at Taiwan Borehole Seismometer Array

1 especially substantial for Clusters A and B. The results for Cluster C can be
 2 reconciled with the constant-stress-drop scaling for $t_p^* = 0.015$ s, which is too large
 3 but perhaps still realistic. However, even this higher-than-observed value of $t_p^* =$
 4 0.015 s is insufficient to explain the results for Clusters A and B, for any value of
 5 (constant) stress drop, as illustrated in Fig. 14(b).

6 We also compare the P-wave pulse width P_{dur}^{vel} between the events within the
 7 clusters and the nearby areas within less than 2 km. As shown in Fig. 15, some of
 8 these nearby events have shorter P_{dur}^{vel} compared with the events within the clusters.
 9 Since these events should experience similar attenuation and path effects in general,
 10 their shorter P_{dur}^{vel} suggests that the P-wave pulse width indeed contains information
 11 about the source, and it is not merely dominated by the attenuation.

12 **Potential physical explanations for the constant duration**

13 What can cause the observed constant source duration? There can be several
 14 explanations, the plausibility of which needs to be further studied, e.g. using
 15 numerical models.

16 One possibility is that all events in each cluster rupture the same seismogenic
 17 patch of a certain size. We can interpret our duration results in terms of the source
 18 dimension of an assumed circular fault with rupture velocity $V_{rup} = 0.75V_s$ where V_s
 19 is the shear-wave velocity. This assumption is commonly used in estimating the

Evidence for Non-Self-Similarity of Microearthquakes Recorded at Taiwan Borehole Seismometer Array

1 source dimension from spectral corner frequency (Brune 1970). We assume that all
 2 events in the present study occurred on the same fault plane (350, 35, -140), which is
 3 shown in Fig. 1(c). Then, the angle between the normal to the fault plane and the
 4 outgoing ray is given by $\theta = 152^\circ$; V_{rup} is 2505 m/s and the P-wave velocity in the
 5 source region is given by $V_p = 5700$ m/s. Thus, the dimension $2r$ can be estimated
 6 from the relation presented by Lanza et al. (1999): $r = \frac{t_w V_{rup}}{1 + V_{rup} \sin \theta / V_p}$, and the
 7 range of rupture lengths $2r$ of the events in Clusters A, B, and C are 100–260 m,
 8 100–310 m, and 80–160 m, respectively (Table 1). For the specific model where we
 9 assume a Q -structure (i.e., Q -correction method), the source dimensions are 242, 232,
 10 and 144 m, respectively.

11 In this model, all events in each cluster, regardless of the magnitude, are hosted
 12 on a patch of the corresponding size, with events of different magnitude
 13 corresponding to different stress drops, e.g. ranging from 0.0007 to 0.32 MPa in
 14 Cluster A. The difference in stress drops can be caused by time-dependent variations
 15 in prestress and other patch properties, e.g. due to variations in pore pressure; during
 16 rupture propagation, such differences could be strongly coupled to differences in the
 17 rupture mode, amount of slip, and amount of co-seismic (dynamic) weakening.
 18 Numerical studies can help identify plausible situations that can lead to the implied
 19 variations in stress drop over the same seismogenic patch. The assumption that all

Evidence for Non-Self-Similarity of Microearthquakes Recorded at Taiwan Borehole Seismometer Array

1 events for each cluster are contained within the same seismogenic patch suggests that
2 the patch is surrounded by aseismic fault regions capable of containing seismic slip, at
3 least for such small events; this can be accomplished by sufficient rate-strengthening
4 properties of the fault outside the patch. Since the source dimensions here are
5 estimated using a dynamic rupture speed, this implies that the corresponding source
6 size is due to dynamic rupture propagation, and hence may be unrelated to the
7 nucleation which is a quasi-static process; in particular, the nucleation size can be
8 much smaller than the estimated source dimensions.

9 Alternatively, the constant duration may be related to the nucleation size. The
10 identified clusters contain some of the smallest earthquakes observed in the region,
11 implying that at least those smallest events may be controlled by the nucleation
12 processes, perhaps arresting right after nucleation due to unfavorable conditions
13 outside the nucleation patch. Since the duration of all events is the same, one can
14 hypothesize that all events in each cluster are controlled by the nucleation. In this
15 model, estimating the source dimension using the traditional assumption on constant
16 rupture speeds may not be relevant, since rupture speeds right after nucleation and
17 right before arrest are likely to be much smaller than the (average) estimates of $0.75V_s$.
18 It remains to be seen whether a physical model can be created in which the nucleation
19 size remains constant – implying a constancy in a number of fault properties,

Evidence for Non-Self-Similarity of Microearthquakes Recorded at Taiwan Borehole Seismometer Array

1 including the effective normal stress – while events spontaneously arrest shortly after
2 nucleation, with significant variations in their stress drop.

3 Another possibility is that the identified seismic events are caused and driven
4 by rapid external factors, e.g. bursts of fluid into the fault zone due to episodic
5 hydrofracturing. In that case, the duration is controlled by the physics of the external
6 factor. Pore pressure variations due to fluid injection into the fault zone from the
7 surrounding rocks have been hypothesized to play a significant role in triggering
8 episodic slow slip in subduction zones (Liu & Rice, 2005; 2007). In those models, the
9 fluid injection and pore pressure variation is a quasi-static process, but perhaps such
10 processes can be dynamic on the small temporal and spatial scales corresponding to
11 the observed events.

12 The study of Lengliné et al. (2014) showed that microearthquakes during a water
13 circulation test in a geothermal reservoir had very similar properties to the events
14 studied here. In particular, the study identified clusters of microearthquakes with
15 constant duration but substantially different magnitudes, similar to our Clusters A–C,
16 and hypothesized that the large differences in magnitudes, linked by Lengliné et al.
17 (2014) to differences in slip and hence stress drop, are due to pore pressure effects.
18 Their discussion focused on aseismic pore pressure processes, in which fluids trigger
19 aseismic fault slip which, in turn, triggers microearthquakes on asperities. The

Evidence for Non-Self-Similarity of Microearthquakes Recorded at Taiwan Borehole Seismometer Array

1 injection-induced variations of pore pressure may indeed cause or contribute to the
2 inferred variations in stress drops. However, by itself, the model with fluid-induced
3 aseismic slip may not explain the constancy of the event durations; for that, an
4 additional assumption of isolated seismogenic patches seems to be needed, as already
5 discussed. An alternative fluid-related possibility is that the slip is driven by a rapid
6 local fluid injection into the fault zone. The associated rapid increase in pore pressure
7 would cause a rapid drop in fault strength and induce rapid (seismic) slip. Once the
8 local fluid injection episode is exhausted (or significantly slows), the slip would cease
9 to be seismic, provided that the scale of the process is below the local nucleation size
10 or the fault has stable (e.g., velocity-strengthening) friction properties. In this
11 mechanism, the duration is controlled not by the fault properties but rather the fluid
12 movements.

14 Variation of spectral shape at high frequency

15 Although the smallest and largest events within each cluster have approximately
16 the same duration, their spectral shapes are different at very high frequencies, above
17 20 Hz. Fig. 9(a) shows the multi-taper displacement spectra (Thomson 1982) of
18 Z-component P waves for the smallest and largest events in Clusters A, B, and C. The
19 noise levels for the smallest and largest events in each cluster are also shown. Fig. 9(b)

Evidence for Non-Self-Similarity of Microearthquakes Recorded at Taiwan Borehole Seismometer Array

1 shows the spectral ratio of the largest event to the smallest one. The spectral ratio
2 shows an appreciable drop in the frequency band 20–50 Hz for all clusters, indicating
3 that the smallest earthquake is relatively richer in high-frequency energy compared
4 with the largest earthquake. Since these events are within the same cluster, they
5 should have the same path effect (attenuation), and we can attribute the different
6 frequency content to the source effects.

7 Differences in the spectral shape between the largest and smallest earthquakes at
8 high frequencies for each cluster should be investigated, e.g. through numerical
9 modelling of the earthquake source. One hypothesis is that the difference can result
10 from heterogeneity or roughness of the source patch, as schematically illustrated in
11 Fig. 16. When rupture is stopped or delayed at a strong asperity within the source
12 patch (illustrated by the irregular circles in Fig. 16), the high-frequency radiation is
13 enhanced (e.g. Lapusta & Liu 2009). Smaller events would correspond to a smaller
14 slip for the same source size and hence may be incapable of rupturing stronger
15 asperities, as illustrated in Fig. 16(a); this produces multiple arrest fronts that radiate
16 high frequencies. In contrast, larger events potentially rupture some or all such
17 asperities, resulting in smoother rupture and hence less high-frequency content. This
18 hypothetical model is consistent with smaller stress drops for smaller events and the
19 constant source duration. The constant duration would be the result of the same

Evidence for Non-Self-Similarity of Microearthquakes Recorded at Taiwan Borehole Seismometer Array

1 overall source size. The smaller stress drops for smaller events would be the result of
2 smaller slip over the patch. When the rupture cannot break the stronger asperities, the
3 slip next to the asperities is suppressed, and the average slip per seismogenic patch
4 (the circle in Fig. 16) would be smaller. Since the stress drop scales with the ratio of
5 slip to the patch size, smaller average slip implies smaller stress drop. Every time slip
6 goes around an asperity, its shear stress increases due to the slip mismatch, bringing
7 the stress on the asperity closer to its (higher) strength. Even though such a scenario
8 creates heterogeneous stress on the fault before a larger event, the patch also has
9 heterogeneous strength, so that mismatch between stress and strength is actually
10 decreased, potentially creating smoother slip during larger events and hence not as
11 much high-frequency seismic signals.

7. Conclusions

14 Using high-quality seismic data from a vertical borehole array installed in central
15 Taiwan, we identified several seismic clusters with high waveform similarity and
16 magnitudes ranging from 0.3–2.0. To investigate the source scaling, we removed the
17 path effect by using an EGF method and a path-averaged Q -correction method. In
18 both analyses, the results indicate a nearly constant source duration for events with
19 $M_w \leq 2.0$ within the clusters, violating the commonly used scaling relation $t_w \propto M_0^{1/3}$.

Evidence for Non-Self-Similarity of Microearthquakes Recorded at Taiwan Borehole Seismometer Array

1
2
3
4 1 If we assume a constant rupture speed within each cluster, this constant duration
5
6
7 2 corresponds to a characteristic length, which may correspond to the size of an isolated
8
9
10 3 seismogenic patch that hosts the cluster events. In this model, events of different slip
11
12
13 4 and hence stress drop would be created by heterogeneity within the patch. The
14
15
16 5 constant duration may also be related to the nucleation processes or outside triggering
17
18
19 6 factors. Further observational and numerical studies are needed to test and distinguish
20
21
22 7 between these potential explanations.
23
24
25 8

1 Acknowledgments

2 We thank Dr. Volker Oye at NORSAR, Norway, who provided Microseismic
3 Monitoring (MIMO) for seismic clusters location. We appreciate the helpful
4 comments from the Associated Editor, two referees, and Prof. Luis Rivera that helped
5 us to improve this manuscript. Support from the Graduate Students Study Abroad
6 Program 101-2917-I-008-001- of the National Science Council (NSC) contributed to
7 this international collaboration. This research was supported by the Taiwan
8 Earthquake Research Center (TEC) funded through the Ministry of Science and
9 Technology (MoST) with grant number MOST-xxxxxxxxxxx. The TEC contribution
10 number for this article is xxxxx.

Evidence for Non-Self-Similarity of Microearthquakes Recorded at Taiwan Borehole Seismometer Array

1 **References**

- 2 Abercrombie, R. E., 1995. Earthquake source scaling relationships from -1 to 5 M_L
3 using seismograms recorded at 2.5-km depth, *J. Geophys. Res.*, 100(12),
4 24015-24036.
- 5 Aki, K., 1967. Scaling law of seismic spectrum, *J. Geophys. Res.*, 72(4), 1217-1231.
- 6 Allmann, B. P. & Shearer, P. M., 2009. Global variations of stress drop for moderate
7 to large earthquakes, *J. Geophys. Res.*, 114, B01310, doi:10.1029/2008JB005821.
- 8 Bouchon, M., Karabulut, H., Aktar, M., Ozalaybey, S., Schmittbuhl, J., & Bouin, M.,
9 2011, Extended nucleation of the 1999 M_w 7.6 Izmit earthquake, *Science*, 331,
10 877-880.
- 11 Brune, J., 1970. Tectonic stress and the spectra of seismic shear waves from
12 earthquakes, *J. Geophys. Res.*, 75(26), 4997-5009.
- 13 Duputel, Z., Tsai, V., Rivera, L. & Kanamori, H., 2013. Using centroid time-delays to
14 characterize source durations and identify earthquakes with unique characteristics,
15 *Earth Planet Sci. Lett.*, 375, 92-100.
- 16 Futterman, W., 1962. Dispersive body waves, *J. Geophys. Res.*, 67(13), 5279-5291.
- 17 Harrington, R. & Brodsky, E., 2009. Source duration scales with magnitude
18 differently for earthquakes on the San Andreas fault and on secondary faults in
19 Parkfield, California, *Bull. Seismol. Soc. Am.*, 99, 2323–2334.
- 20 Ide, S., Beroza, G. C., Prejean, S. G. & Ellsworth, W. L., 2003. Apparent break in
21 earthquake scaling due to path and site effects on deep borehole recordings, *J.*
22 *Geophys. Res.*, 108(5), 2271-2284.

Evidence for Non-Self-Similarity of Microearthquakes Recorded at Taiwan Borehole Seismometer Array

- 1
2
3
4 1 Ide, S., Matsubara, M. & Obara, K., 2004. Exploitation of high-sampling Hi-net data
5
6 2 to study seismic energy scaling: The aftershocks of the 2000 western Tottori, Japan,
7
8 3 earthquake, *Earth Planets Space*, 56, 859-871.
9
10
11 4 Kanamori, H. & Anderson, D., 1977. Importance of Physical dispersion in surface
12
13 5 wave and free oscillation problems: review, *Rev. Geophys. Space Phys.*, 15(1),
14
15 6 105-112.
16
17
18 7 Kanamori, H. & Rivera, L., 2015. Near-vertical multiple ScS phases and vertically
19
20 8 averaged mantle properties, in Foulger, G.R., Lustrino, M., and King, S.D., eds.,
21
22 9 *The Interdisciplinary Earth: A Volume in Honor of Don L. Anderson*: Geological
23
24 10 Society of America Special Paper 514 and American Geophysical Union Special
25
26 11 Publication 71, 9–31, doi:10.1130/2015.2514(02).
27
28
29 12 Kim, K., Chiu, J., Pujol, J., Chen, K., Huang, B., Yeh, Y. & Shen, P., 2005.
30
31 13 Three-dimensional V_P and V_S structural models associated with the active
32
33 14 subduction and collision tectonics in the Taiwan region, *Geophys. J. Int.*, 162,
34
35 15 204-220.
36
37
38 16 Lanza, V., Spallarossa, D., Cattaneo, M., Bindi, D. & Augliera, P., 1999. Source
39
40 17 parameters of small events using constrained deconvolution with empirical Green's
41
42 18 functions, *Geophys. J. Int.*, 137, 651-662.
43
44
45
46 19 Lapusta, N. & Liu, Y., 2009. Three-dimensional boundary integral modeling of
47
48 20 spontaneous earthquake sequences and aseismic slip, *J. Geophys. Res.*, 114,
49
50 21 B09303, doi:10.1029/2008JB005934.
51
52
53
54
55 22 Lengliné, O., Lamourette, L., Vivin, L., Cuenot, N. & Schmittbuhl, J., 2014.
56
57 23 Fluid-induced earthquakes with variable stress drop, *J. Geophys. Res.*, 119,
58
59 24 8900-8913, doi: 10.1002/2014JB011282.
60

Evidence for Non-Self-Similarity of Microearthquakes Recorded at Taiwan Borehole Seismometer Array

- 1
2
3
4 1 Lin, Y. Y., Ma, K. F. & Oye, V., 2012. Observation and scaling of microearthquakes
5
6
7 2 from the Taiwan Chelungpu-fault borehole seismometers, *Geophys. J. Int.*, 190,
8
9
10 3 665-676.
11
12
13 4 Lin, Y. Y., 2014. *TCDP Borehole Seismometers Array – Microearthquake*
14
15 5 *Observations and Seismic Source Characteristics Investigation*, PhD thesis,
16
17 6 National Central University, Zhongli, Taiwan, in Chinese.
18
19
20 7 Liu, Y. & Rice, J. R., 2005. Aseismic slip transients emerge spontaneously in 3D rate
21
22 8 and state modeling of subduction earthquake sequences, *J. Geophys. Res. - Solid*
23
24 9 *Earth*, 110, B08307, doi:10.1029/2004JB003424.
25
26
27 10 Liu, Y. & Rice, J. R., 2007. Spontaneous and triggered aseismic deformation
28
29 11 transients in a subduction fault model, *J. Geophys. Res. - Solid Earth*, 112, B09404,
30
31 12 doi: 10.1029/2007JB004930.
32
33
34 13 Ma, K. F., Tanaka, H., Song, S. R., Wang, C. Y., Hung, J. H., Tsai, Y. B., Mori, J.,
35
36 14 Song, Y. F., Yeh, E. C., Sone, H., Kuo, L. W. & Wu, H. Y., 2006. Slip zone and
37
38 15 energetics of a large earthquake from the Taiwan Chelungpu-fault Drilling Project,
39
40 16 *Nature*, 444, 473-476.
41
42
43 17 Ma, K. F., Lin, Y. Y., Lee, S. J., Mori, J. & Brodsky, E. E., 2012. Isotropic events
44
45 18 observed with a borehole array in the Chelungpu fault zone, Taiwan, *Science*, 337,
46
47 19 459-463, doi: 10.1126/science.1222119.
48
49
50 20 Madariaga, R., 1976. Dynamics of an expanding circular fault, *Bull. Seismol. Soc.*
51
52 21 *Am.*, 66, 639-666.
53
54
55 22 Mayeda, K. & Malagnini, L., 2009. Apparent stress and corner frequency variations in
56
57 23 the 1999 Taiwan (Chi-Chi) sequence: Evidence for a step-wise increase at $M_w \sim$
58
59 24 5.5, *Geophys. Res. Lett.*, 36, L10308.
60

Evidence for Non-Self-Similarity of Microearthquakes Recorded at Taiwan Borehole Seismometer Array

- 1
2
3
4 1 Mori, J., Abercrombie, R. & Kanamori, H., 2003. Stress drops and radiated energies
5
6 2 of aftershocks of the 1994 Northridge, California, earthquake, *J. Geophys. Res.*,
7
8 3 108(11), 2545-2557.
9
10 4 Oye, V. & Roth, M., 2003. Automated seismic event location for hydrocarbon
11
12 5 reservoirs, *Computers & Geosciences*, 29, 851-863.
13
14 6 Oye, V., Bungum, H. & Roth, M., 2005. Source parameters and scaling relations for
15
16 7 mining-related seismicity within the Pyhasalmi Ore Mine, Finland, *Bull. Seism.*
17
18 8 *Soc. Am.*, 95(3), 1011-1026.
19
20 9 Reasenber, P. A. & Oppenheimer, D., 1985. FPFIT, FPLOT, and FPPAGE: Fortran
21
22 10 computer programs for calculating and displaying earthquake fault-plane solutions,
23
24 11 *U.S. Geol. Surv. Open-File Rep.* 85-739.
25
26 12 Rice, J., 1993. Spatio-temporal complexity of slip on a fault, *J. Geophys. Res.*, 98,
27
28 13 9885-9907, doi:10.1029/93JB00191.
29
30 14 Rice, J. & Ruina, A., 1983. Stability of steady frictional slipping, *J. Appl. Mech.*, 50,
31
32 15 343-349.
33
34 16 Rubin, A. & Ampuero, J., 2005. Earthquake nucleation on (aging) rate and state faults,
35
36 17 *J. Geophys. Res.*, 110(B1), B11312, doi:10.1029/2005JB003686.
37
38 18 Stein, S. & Wysession, M., 2003. *An Introduction to Seismology, Earthquakes, and*
39
40 19 *Earth Structure*. pp. 192, Oxford: Blackwell Publishing Ltd.
41
42 20 Stork, A. L. & Ito, H., 2004. Source Parameter Scaling for Small Earthquakes
43
44 21 Observed at the Western Nagano 800-m-Deep Borehole, Central Japan, *Bull. Seism.*
45
46 22 *Soc. Am.*, 94(5), 1781-1794.
47
48 23 Thomson, D. J., 1982. Spectrum estimation and harmonic analysis, *Proc. IEEE*, 70,
49
50 24 1055-1096.
51
52
53
54
55
56
57
58
59
60

Evidence for Non-Self-Similarity of Microearthquakes Recorded at Taiwan Borehole Seismometer Array

- 1 Venkataraman, A., Beroza, G. C., Ide, S., Imanishi, K., Ito, H. & Iio, Y., 2006.
2 Measurements of spectral similarity for microearthquakes in western Nagano,
3 Japan, *J. Geophys. Res.*, 111, B03303.
- 4 Wang, Y. J., Ma, K. F., Mouthereau, F. & Eberhart-Phillips, D., 2010. Three
5 dimensional Q_p - Q_s -Tomography beneath Taiwan Orogenic Belt: Comparison to
6 the Tectonic and the Thermal Structure, *Geophys. J. Int.*, 180, 891–910.
- 7 Wang, Y. J., Lin, Y. Y., Ma, K. F. & Lee, M. C., 2012. Fault zone Q values derived
8 from Taiwan Chelungpu Fault borehole seismometers (TCDPBHS),
9 *Tectonophysics*, 578, 76-86, doi:10.1016/j.tecto.2011.12.027.
- 10 Yamada, T., Mori, J., Ide, S., Abercrombie, R. E., Kawakata, H., Nakatani, M., Iio, Y.
11 & Ogasawara, H., 2007. Stress drops and radiated seismic energies of
12 microearthquakes in a South African gold mine, *J. Geophys. Res.*, 112, B03305.
13

1 Figure Captions

2 **Fig. 1** (a) Distribution of the clusters (squares) and background earthquakes (circles).

3 The color and size of the circles indicate the duration P_{dur}^{vel} and magnitude of
 4 the events, respectively. The waveforms of some of the background events are
 5 shown in Fig. 4(d). The red triangle and red lines indicate the surface locations
 6 of the Taiwan Chelungpu-Fault Drilling Project borehole seismometer array
 7 (TCDPBHS) and Chelungpu fault, respectively. The grey inverted triangles
 8 show the distribution of the surface seismic stations. Ellipses in the lower
 9 figure show the microearthquake detection capability of the TCDPBHS. (b)
 10 The layout of the TCDPBHS (modified from Lin et al. 2012). The seven
 11 stations are located over the depth range from 946 m to 1274 m at 50 to 60 m
 12 depth intervals (right section, green rectangles). Station no. 4 (BHS4) was
 13 installed very close to the Chelungpu main fault (right section, purple line)
 14 inside the Chinshui shale. (c) The focal mechanism of an event in Cluster A
 15 determined from the P-wave first-motion solution obtained using the method
 16 of Reasenberg & Oppenheimer (1985). The traces represent P-wave
 17 waveforms from the stations. The crosses and circles indicate the upward and
 18 downward P-wave first motion, respectively. The best pairs of nodal planes
 19 are 350, 35, -140 and 225, 68, -62, where each set of three values denotes the
 20 strike, dip, and rake, respectively.

Evidence for Non-Self-Similarity of Microearthquakes Recorded at Taiwan Borehole Seismometer Array

Fig. 2 (a) Instrument response of the TCDPBHS. (b) Regression analysis of the maximum S-wave amplitude of the horizontal vector sum A (cm/s) and M_w . The circles show 242 microearthquakes observed by the TCDPBHS (Lin et al. 2012). The regression line $M_{we} = 0.67\log A + 3.48$ is indicated by a solid line. Magnitudes estimated from this empirical relation are called empirical M_w (M_{we}). We ignore the hypocenter distance term in the regression because all microearthquakes occurred at approximately the same hypocenter distances of 10–15 km in the study area. The dotted lines indicate the 95% confidence interval (2σ) of the regression estimates. The variation in M_{we} as estimated from the relation is ± 0.3 . (c) Distribution of event sizes for Clusters A–G. The circles indicate the magnitudes of the events. The number of events in each cluster is indicated at the top of each bar.

Fig. 3 Comparison of the Z - and T -component waveforms for the smallest (M_{we} 0.27) and largest (M_{we} 1.97) events in Cluster A. The waveforms recorded at BHS1 (upper section) and BHS4 (lower section) are shown; the P-wave and S-wave observations are reflected in the vertical (Z) and transverse (T) components, respectively. The smallest event shows higher-frequency content, but the P-wave source duration is very similar to that of the largest event. S waves for the smaller and larger events are almost identical.

Evidence for Non-Self-Similarity of Microearthquakes Recorded at Taiwan Borehole Seismometer Array

Fig. 4 Z-component P-wave velocity seismograms recorded at BHS4 for (a) Cluster A, (b) Cluster B, and (c) Cluster C. Red bars on the first trace in (a) indicate the P-wave duration (P_{dur}^{vel}) of the event. The maximum amplitude and magnitude M_{we} are mentioned above each trace. (d) Seismograms of regular events. (e) Comparison of the normalized P waves for Clusters A (left), B (middle), and C (right), respectively, in a smaller time window (0.3 s). The black and red lines show the records for the largest and smallest events, respectively. The thinner blue lines show seismograms of the other events in each cluster.

Fig. 5 P-wave duration (P_{dur}^{vel}) versus seismic moment for regular events (black circles) and for events in Clusters A, B, and C. The circles show 242 observations (Lin et al. 2012). The red, blue, and green horizontal lines indicate P_{dur}^{vel} of Clusters A, B, and C, respectively, and the yellow circles refer to the regular events shown in Fig. 4(d).

Fig. 6 Definition of the displacement source pulse width. The pulse width of the displacement record is defined as t_w . For more details, see the text.

Fig. 7 Waveforms (0.8-s resolution) of the Z-component (P wave) displacement seismograms recorded at BHS4 for (a) Cluster A, (b) Cluster B, and, (c) Cluster C. The zero-phase band-pass filter with a pass band of 5 to 50 Hz is

Evidence for Non-Self-Similarity of Microearthquakes Recorded at Taiwan Borehole Seismometer Array

1 applied. The black dots indicate half of the P-wave duration (P_{dur}^{dis}) of the
 2 events. The average of P_{dur}^{dis} for each cluster is shown above the traces.

3 **Fig. 8** Source-time functions (STFs) determined using the empirical Green's function
 4 (EGF) method for (a) Cluster A, (b) Cluster B, and (c) Cluster C. The
 5 magnitudes are indicated above each trace. The black dots indicate the
 6 half-pulse width of the STF of the events. The waveforms without dots are not
 7 used for pulse width determination because of a low signal-to-noise ratio.

8 **Fig. 9** (a) Displacement spectra of the observed Z-component (P waves) for the
 9 smallest (red curve) and largest (blue curve) events in Clusters A (left), B
 10 (middle), and C (right). The multi-taper spectral method (Thomson 1982) was
 11 used for calculating spectra. The red and blue dash-dot lines indicate the noise
 12 level of the smallest and largest events in the clusters, respectively. (b)
 13 Spectral ratios between the largest and smallest events for the three clusters.
 14 Appreciable drops in the ratio can be seen over the frequency band from 20 to
 15 50 Hz for signal-to-noise ratios greater than 3.0.

16 **Fig. 10** (a) Q -model ($Q_p = 202$, $t_p^* = 0.0117$) used in this study. (b) The Futterman
 17 function determined for the Q -structure shown in Fig. 10(a).

18 **Fig. 11** STFs determined using the Q -correction method and a zero-phase band-pass
 19 filter (5–50 Hz) for (a) Cluster A, (b) Cluster B, and (c) Cluster C. The black

Evidence for Non-Self-Similarity of Microearthquakes Recorded at Taiwan Borehole Seismometer Array

1 dots and grey triangles on each trace indicate the half-pulse width and the left
 2 and right bottom amplitudes presented in Fig. 6.

3 **Fig. 12** Comparison of the first pulse of the SH wave, $O_S(t)$, on the T -component
 4 (red lines) and $O_P(t) * F(t; 3t_p^*)$, convolution of the P wave on the vertical
 5 component and $F(t; 3t_p^*)$ with a given t_p^* (black lines), for (a) Cluster A, (b)
 6 Cluster B, and (c) Cluster C.

7 **Fig. 13** Source durations of events in Clusters A, B, and C estimated using the three
 8 different methods. The red, green, and blue symbols denote the results for
 9 Clusters A, B, and C, respectively. The circles indicate the source durations (t_w)
 10 estimated using the Q -correction method. The dotted lines show the mean
 11 values for each cluster, and the triangles show the source durations (t_w)
 12 estimated using the EGF method, with the average indicated by a solid black
 13 line. The red, green, and blue solid lines represent the upper bounds of the
 14 STF duration for Clusters A, B, and C, respectively. The dashed black curves
 15 show the self-similarity relation ($t_w \propto M_0^{1/3}$) for constant stress drops of 0.2, 2,
 16 and 20 MPa. The curve labelled as $\Delta\sigma = 0.2$ MPa is close to that presented by
 17 Duputel et al. (2013). The right vertical axis gives the source dimension.

18 **Fig. 14** (a) Comparison between the computed P-pulse widths of an event with a
 19 stress drop of 20 MPa and observations. The triangles and lines represent the
 20 computed P-pulse widths for a suite of t_p^* from 0.010 to 0.045 s. Red, green,

Evidence for Non-Self-Similarity of Microearthquakes Recorded at Taiwan Borehole Seismometer Array

1 and blue circles indicate the observed pulse widths of the P waves for Clusters
 2
 3
 4
 5
 6
 7
 8
 9
 10
 11
 12
 13
 14
 15
 16
 17
 18
 19
 20
 21
 22
 23
 24
 25
 26
 27
 28
 29
 30
 31
 32
 33
 34
 35
 36
 37
 38
 39
 40
 41
 42
 43
 44
 45
 46
 47
 48
 49
 50
 51
 52
 53
 54
 55
 56
 57
 58
 59
 60

Fig. 15 Events near the cluster with a shorter duration P_{dur}^{vel} compared to an event in
 (a) Cluster A, (b) Cluster B, and (c) Cluster C. Locations of the nearby events
 (circles) and the cluster (asterisk) are shown in the upper section. The color
 and size of the circles indicate the duration P_{dur}^{vel} and magnitude of the events,
 respectively. The crosses are the events in the background. The Z-component
 velocity waveforms for the event inside the cluster (top trace in the red box)
 and the nearby events are shown in the lower section. The duration P_{dur}^{vel} is
 measured between T1 and T2 markers. The red dashed line is drawn at the T2
 marker of the cluster event as a reference. The estimated P_{dur}^{vel} is indicated on
 each trace.

Fig. 16 Schematic of (a) small and (b) large earthquakes, motivated by the spectral
 ratios shown in Fig. 9. The white irregular circles represent stronger asperities
 that delay or stop rupture for small events, exciting high-frequency energy

Evidence for Non-Self-Similarity of Microearthquakes Recorded at Taiwan Borehole Seismometer Array

1 (illustrated in the right column). Schematic of slip distributions are shown on
 2 the left, and the associated STFs are shown on the right.

3 **Fig. A1** Records at the seven TCDPBHS stations, BHS1-BHS7, of the unfiltered
 4 Z-component (P wave) velocity seismograms of the largest event of (a) Cluster
 5 A, (b) Cluster B, and (c) Cluster C. The records are similar for all the stations,
 6 as expected given the distance of 10-15 km to the sources and the spatial
 7 extent of the array of ~300 m. (d) Comparison of the P wave records for
 8 Clusters A (left), B (middle), and C (right).

9 **Fig. A2** Effect of the band-pass filter on the pulse width measurement of the STF. (a)
 10 Triangular STF with t_w ranging from 0.002 to 0.030 s. (b) Waveforms filtered
 11 by a zero-phase band-pass filter with a pass band of 5 to 50 Hz. The pulse
 12 widths are measured using the method illustrated in Fig. 6 and indicated on the
 13 left of the waveforms. (c) Comparison of the given t_w and estimated t_w . (d)
 14 The difference between the estimated and the given t_w as a function of the
 15 estimated t_w .

Table Legends

17 **Table 1.** Seismic clusters used in the present study: results of the EGF and the
 18 Q -correction methods
 19

Evidence for Non-Self-Similarity of Microearthquakes Recorded at Taiwan Borehole Seismometer Array

Table 1: Seismic clusters used in the present study: results of the EGF and the Q -correction methods

Cluster	Number	Date	Hour	Min	Second	M_{we}	CC	Δt_{sp} (s)	P_{dur}^{dis} (s)	EGF		Q -correction	
										t_w (s)	Dimension $2r$ (m)	t_w (s)	Dimension $2r$ (m)
A	1	20061126	18	02	173.313	0.38	1.00	1.650	0.068	0.025	103.8	0.054	224.3
	2	20061129	07	37	2223.476	1.51	0.92	1.652	0.071	0.029	120.4	0.062	257.9
	3	20061129	07	37	2240.536	1.02	0.82	1.660	0.072	0.025	103.8	0.063	262.9
	4	20061129	07	37	2254.785	0.70	0.82	1.657	0.068	0.024	99.7	0.059	245.0
	5	20070129	09	34	2051.911	0.79	0.87	1.654	0.071	0.024	99.7	0.064	265.8
	6*	20070526	18	48	2892.767	0.27	0.82	1.648	0.074	-	-	0.060	249.2
	7	20070820	13	13	802.829	0.88	0.90	1.648	0.069	0.025	103.8	0.055	228.4
	8	20071018	16	20	1231.643	1.97	0.90	1.647	0.068	0.030	124.6	0.056	230.9
	9	20071018	16	21	1285.838	1.15	0.83	1.641	0.070	0.027	112.1	0.056	232.6
	10	20071020	06	12	751.139	1.41	0.81	1.642	0.072	0.028	116.3	0.060	249.2
	11	20071020	06	33	1996.932	1.51	0.81	1.641	0.068	0.027	112.1	0.054	224.3
B	1	20061124	13	56	3419.909	1.51	1.00	1.800	0.062	-	-	0.058	241.7
	2	20061124	13	59	3566.534	0.50	0.92	1.798	0.065	0.026	108.0	0.056	232.6
	3	20061124	14	01	118.051	1.04	0.93	1.798	0.065	0.025	103.8	0.054	224.3
	4	20061124	14	04	268.517	1.62	0.88	1.805	0.082	0.029	120.4	0.075	311.5
	5	20061124	14	04	275.232	1.52	0.94	1.797	0.069	-	-	0.066	274.1
	6*	20061124	19	40	2433.970	0.49	0.85	1.795	0.063	-	-	0.048	199.4
	7	20070101	09	17	1026.288	0.62	0.81	1.811	0.055	-	-	0.047	195.2
	8	20070205	09	00	47.989	0.57	0.85	1.798	0.055	0.025	103.8	0.042	174.4
C	1	20061114	22	38	2286.017	0.77	1.00	1.600	0.039	0.021	87.2	0.033	137.1
	2	20061114	23	14	883.510	0.69	0.95	1.602	0.040	0.024	99.7	0.034	141.2
	3	20061114	23	19	1143.311	1.32	0.94	1.604	0.049	0.025	103.8	0.037	153.7
	4*	20061115	00	00	11.242	0.35	0.92	1.598	0.038	-	-	0.032	132.9
	5	20061115	00	17	1042.864	0.45	0.93	1.597	0.038	0.021	87.2	0.032	132.9
	6	20070723	05	59	3560.207	0.94	0.84	1.606	0.051	-	-	0.039	160.3

*The smallest event that is used as an empirical Green's function in EGF analysis.

Evidence for Non-Self-Similarity of Microearthquakes Recorded at Taiwan Borehole Seismometer Array

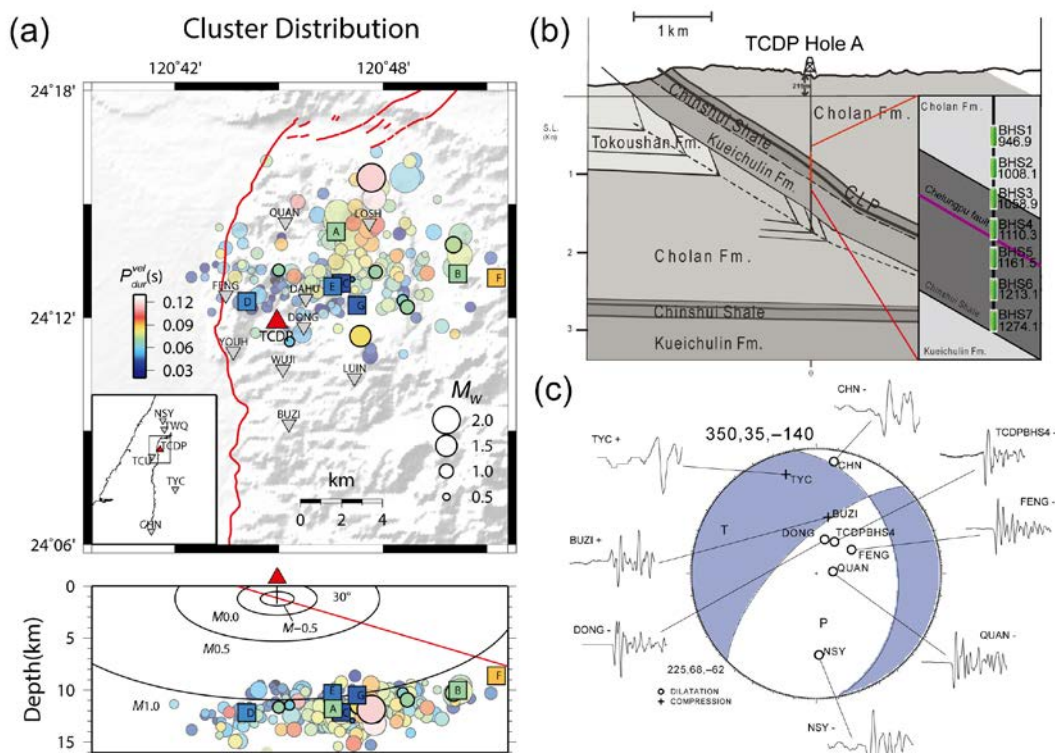


Fig. 1 (a) Distribution of the clusters (squares) and background earthquakes (circles). The color and size of the circles indicate the duration P_{dur}^{vel} and magnitude of the events, respectively. The waveforms of some of the background events are shown in Fig. 4(d). The red triangle and red lines indicate the surface locations of the Taiwan Chelungpu-Fault Drilling Project borehole seismometer array (TCDPBHS) and Chelungpu fault, respectively. The grey inverted triangles show the distribution of the surface seismic stations. Ellipses in the lower figure show the microearthquake detection capability of the TCDPBHS. (b) The layout of the TCDPBHS (modified from Lin et al. 2012). The seven stations are located over the depth range from 946 m to 1274 m at 50 to 60 m depth intervals (right section, green rectangles). Station no. 4 (BHS4) was installed very close to the Chelungpu main fault (right section, purple line) inside the Chinshui shale. (c) The focal mechanism of an event in Cluster A determined from the P-wave first-motion solution obtained using the method of Reasenberg & Oppenheimer (1985). The traces represent P-wave waveforms from the stations. The crosses and circles indicate the upward and downward P-wave first motion, respectively. The best pairs of nodal planes are 350, 35, -140 and 225, 68, -62, where each set of three values denotes the strike, dip, and rake, respectively.

Evidence for Non-Self-Similarity of Microearthquakes Recorded at Taiwan Borehole Seismometer Array

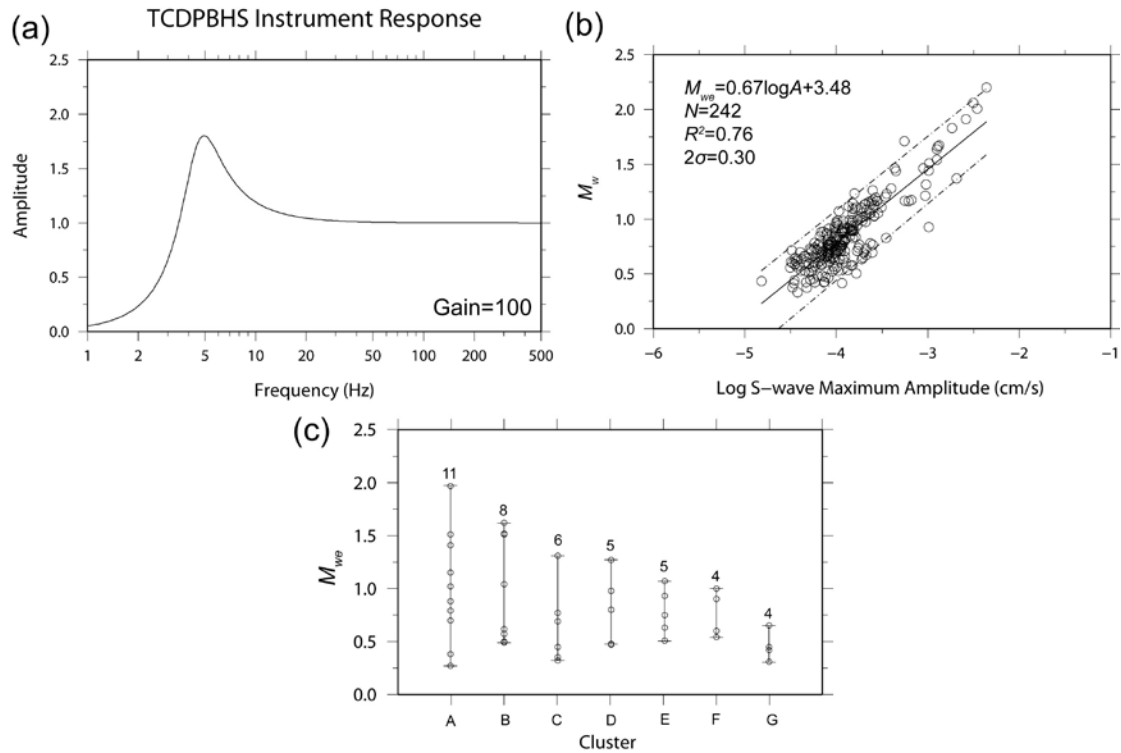
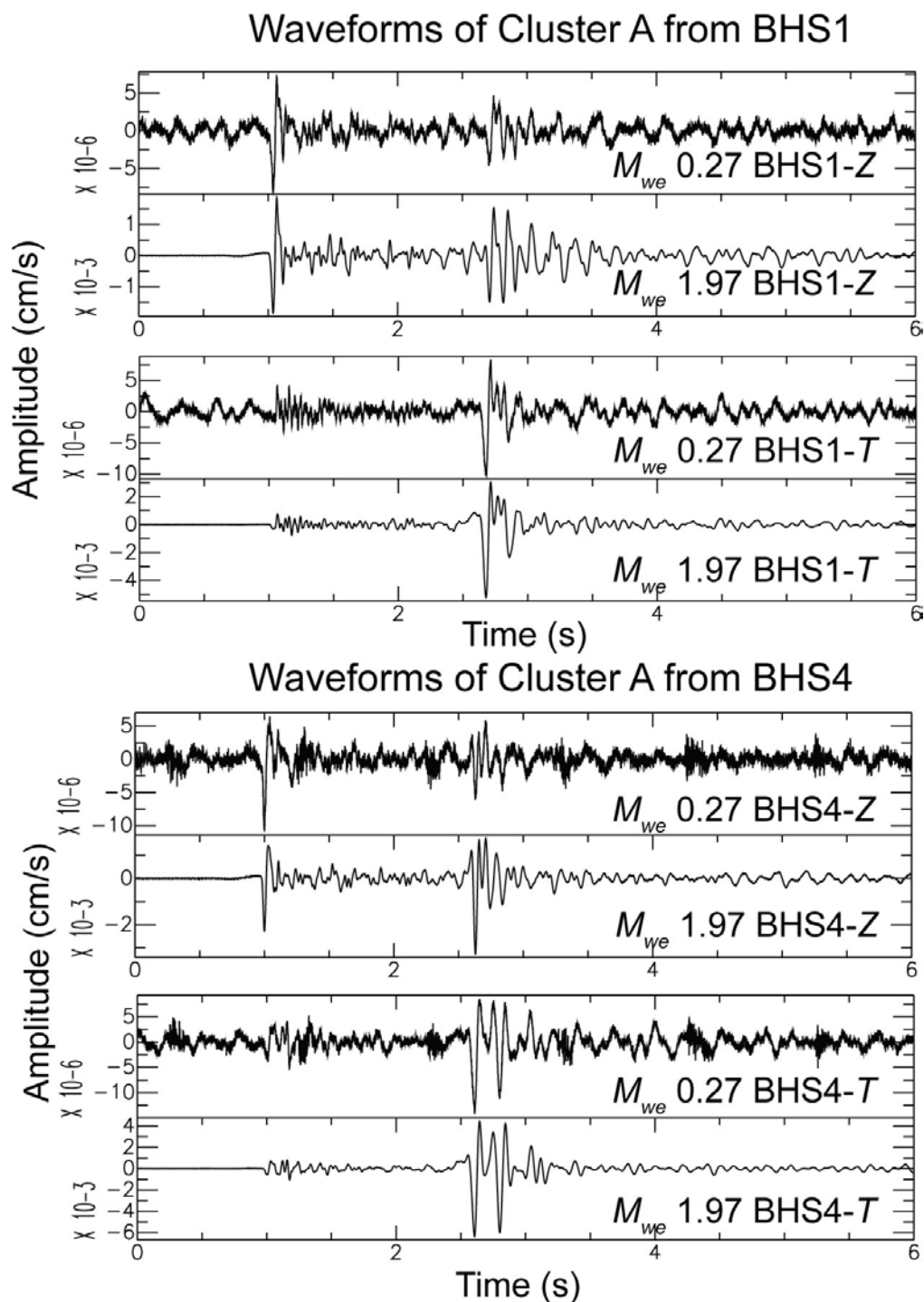


Fig. 2 (a) Instrument response of the TCDPBHS. (b) Regression analysis of the maximum S-wave amplitude of the horizontal vector sum A (cm/s) and M_w . The circles show 242 microearthquakes observed by the TCDPBHS (Lin et al. 2012). The regression line $M_{we} = 0.67 \log A + 3.48$ is indicated by a solid line. Magnitudes estimated from this empirical relation are called empirical M_w (M_{we}). We ignore the hypocenter distance term in the regression because all microearthquakes occurred at approximately the same hypocenter distances of 10–15 km in the study area. The dotted lines indicate the 95% confidence interval (2σ) of the regression estimates. The variation in M_{we} as estimated from the relation is ± 0.3 . (c) Distribution of event sizes for Clusters A–G. The circles indicate the magnitudes of the events. The number of events in each cluster is indicated at the top of each bar.

Evidence for Non-Self-Similarity of Microearthquakes Recorded at Taiwan Borehole Seismometer Array



1

2 **Fig. 3** Comparison of the Z- and T-component waveforms for the smallest (M_{we} 0.27)
 3 and largest (M_{we} 1.97) events in Cluster A. The waveforms recorded at BHS1 (upper
 4 section) and BHS4 (lower section) are shown; the P-wave and S-wave observations
 5 are reflected in the vertical (Z) and transverse (T) components, respectively. The
 6 smallest event shows higher-frequency content, but the P-wave source duration is
 7 very similar to that of the largest event. S waves for the smaller and larger events
 8 are almost identical.

8

59

60

Evidence for Non-Self-Similarity of Microearthquakes Recorded at Taiwan Borehole Seismometer Array

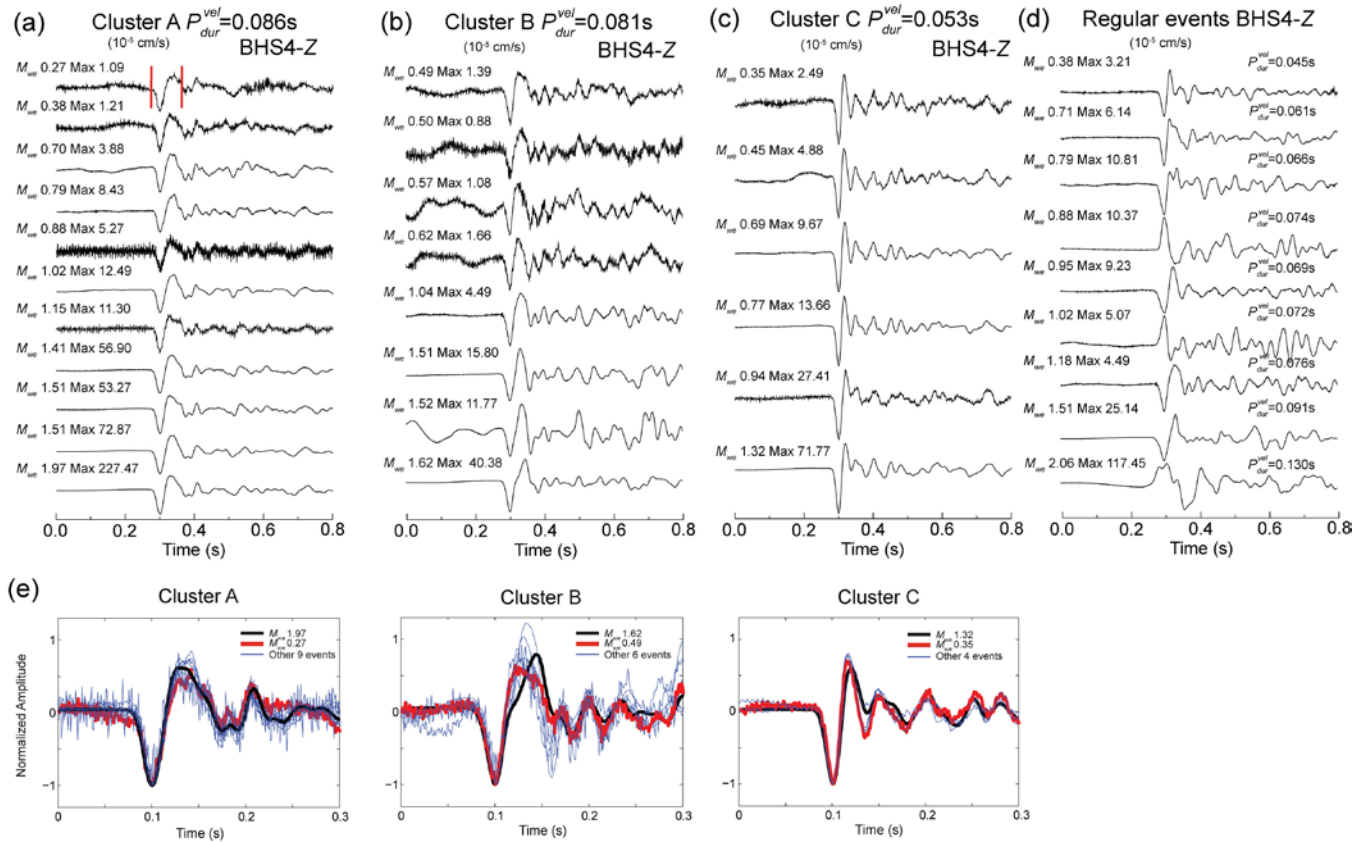


Fig. 4 Z-component P-wave velocity seismograms recorded at BHS4 for (a) Cluster A, (b) Cluster B, and (c) Cluster C. Red bars on the first trace in (a) indicate the P-wave duration (P_{dur}^{vel}) of the event. The maximum amplitude and magnitude M_{we} are mentioned above each trace. (d) Seismograms of regular events. (e) Comparison of the normalized P waves for Clusters A (left), B (middle), and C (right), respectively, in a smaller time window (0.3 s). The black and red lines show the records for the largest and smallest events, respectively. The thinner blue lines show seismograms of the other events in each cluster.

Evidence for Non-Self-Similarity of Microearthquakes Recorded at Taiwan Borehole Seismometer Array

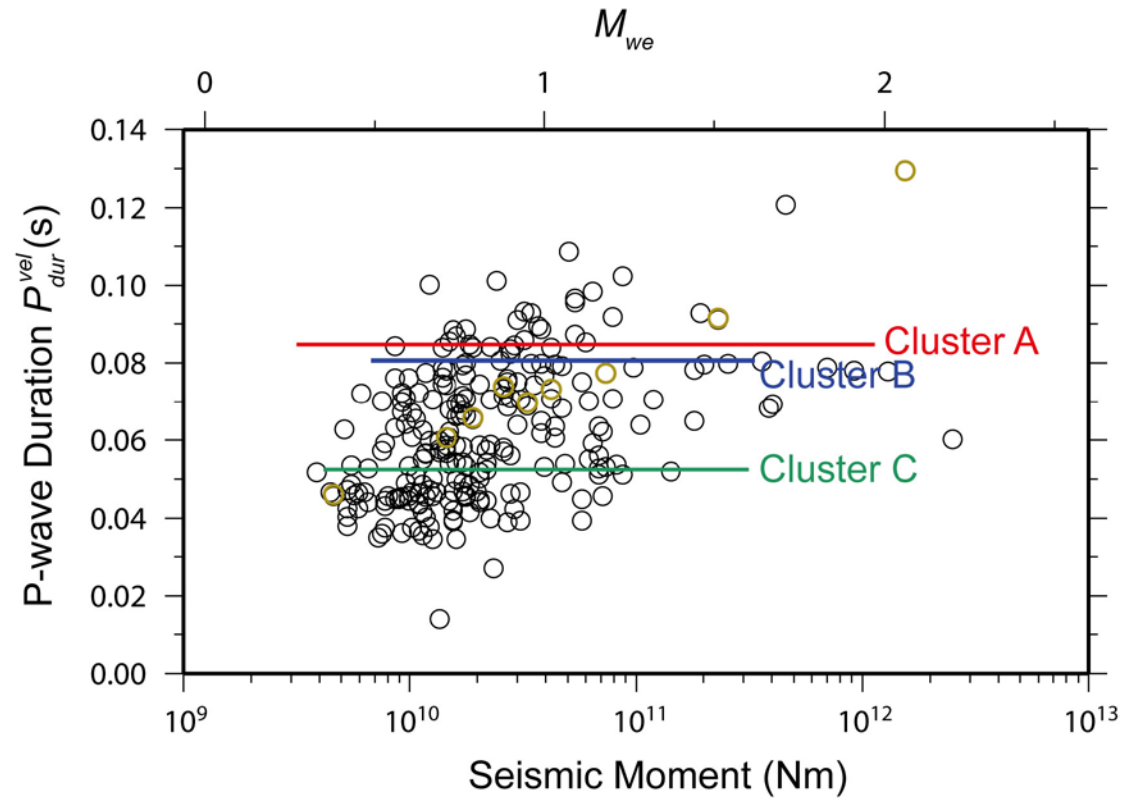
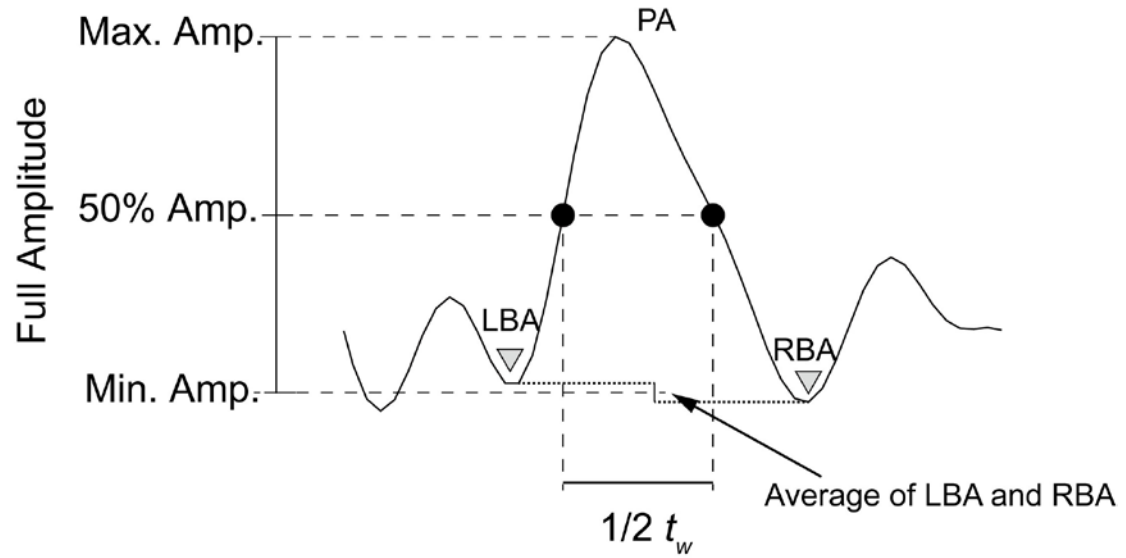


Fig. 5 P-wave duration (P_{dur}^{vel}) versus seismic moment for regular events (black circles) and for events in Clusters A, B, and C. The circles show 242 observations (Lin et al. 2012). The red, blue, and green horizontal lines indicate P_{dur}^{vel} of Clusters A, B, and C, respectively, and the yellow circles refer to the regular events shown in Fig. 4(d).

Definition of Pulse Width



1
2 **Fig. 6** Definition of the displacement source pulse width. The pulse width of the
3 displacement record is defined as t_w . For more details, see the text.

Evidence for Non-Self-Similarity of Microearthquakes Recorded at Taiwan Borehole Seismometer Array

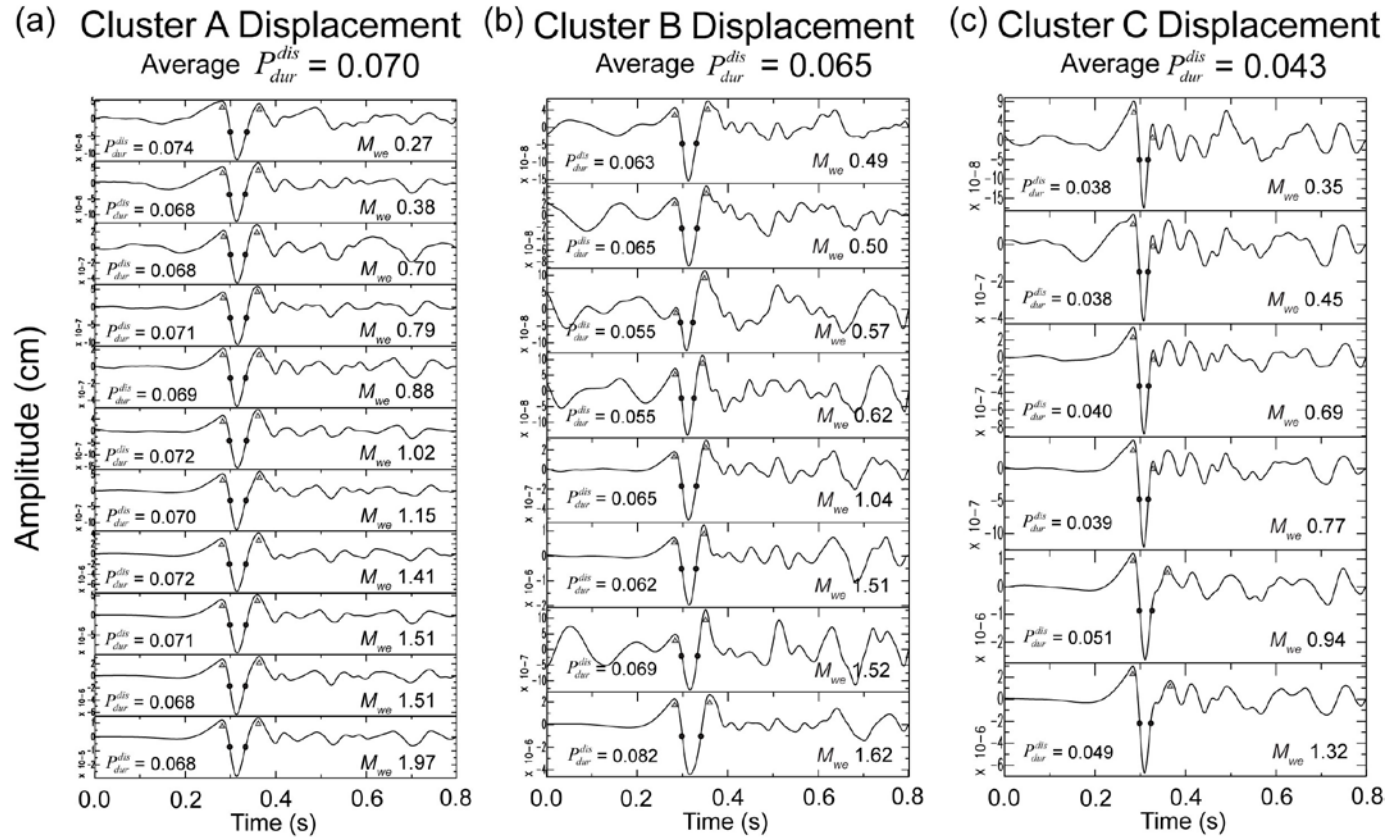


Fig. 7 Waveforms (0.8-s resolution) of the Z-component (P wave) displacement seismograms recorded at BHS4 for (a) Cluster A, (b) Cluster B, and, (c) Cluster C. The zero-phase band-pass filter with a pass band of 5 to 50 Hz is applied. The black dots indicate half of the P-wave duration (P_{dur}^{dis}) of the events. The average of P_{dur}^{dis} for each cluster is shown above the traces.

Evidence for Non-Self-Similarity of Microearthquakes Recorded at Taiwan Borehole Seismometer Array

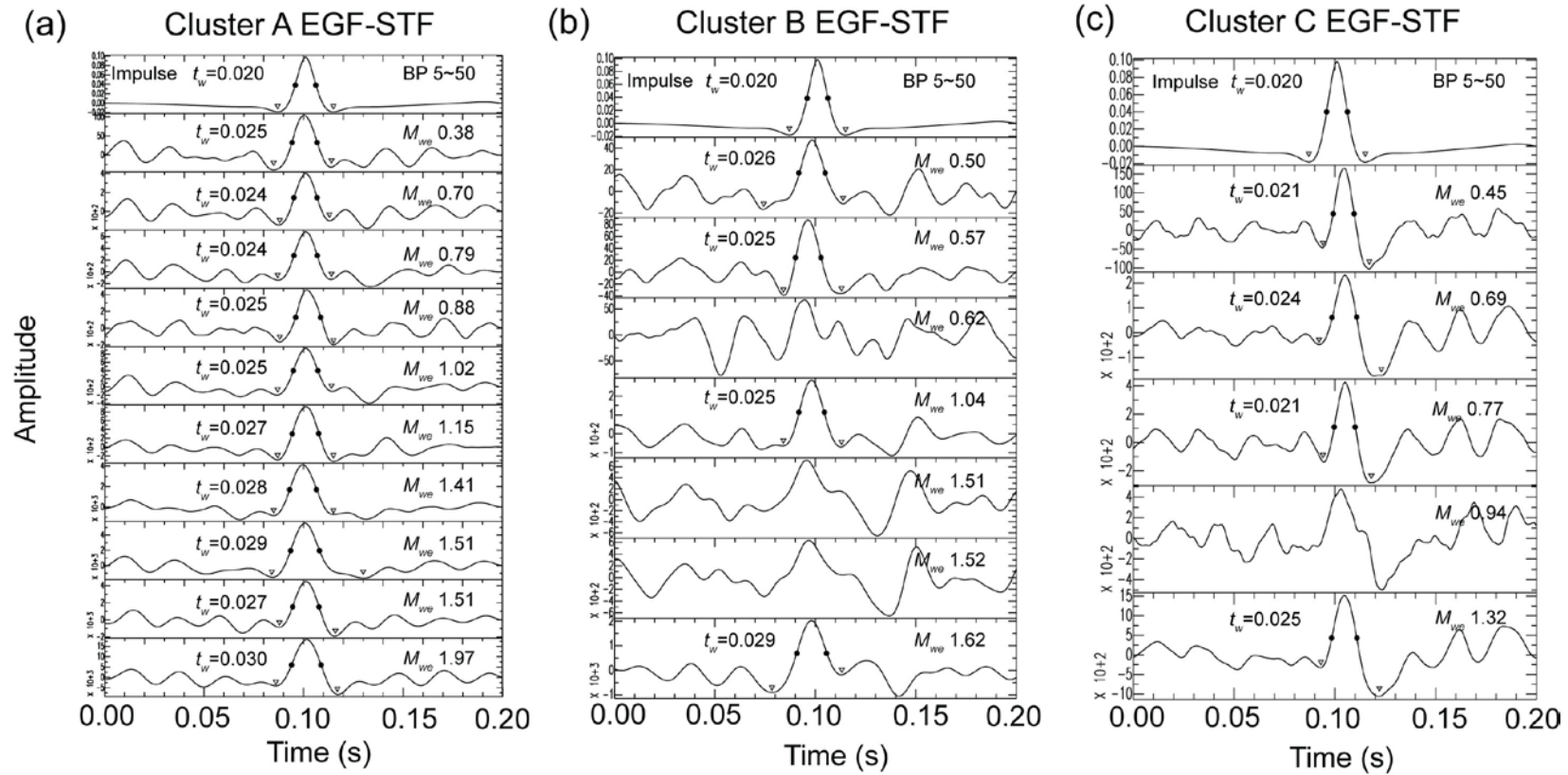


Fig. 8 Source-time functions (STFs) determined using the empirical Green's function (EGF) method for (a) Cluster A, (b) Cluster B, and (c) Cluster C. The magnitudes are indicated above each trace. The black dots indicate the half-pulse width of the STF of the events. The waveforms without dots are not used for pulse width determination because of a low signal-to-noise ratio.

Evidence for Non-Self-Similarity of Microearthquakes Recorded at Taiwan Borehole Seismometer Array

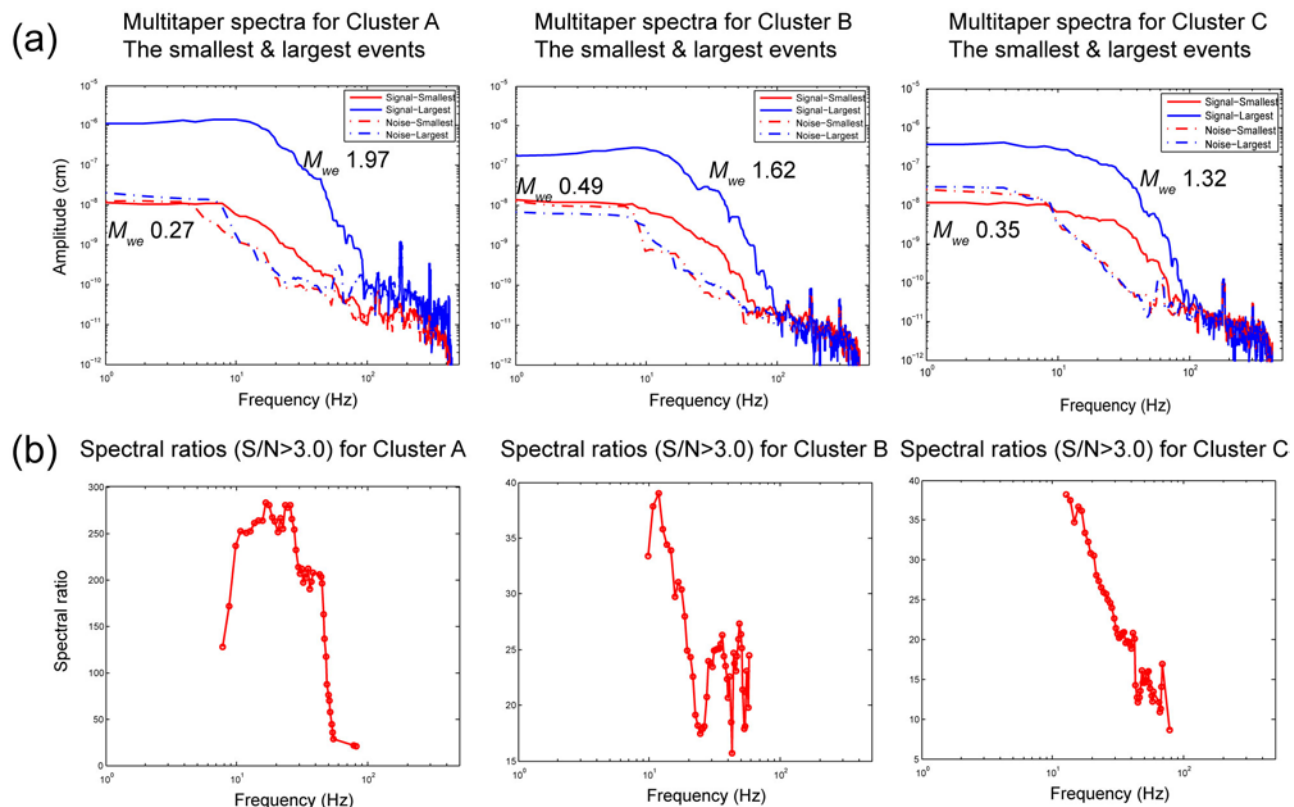
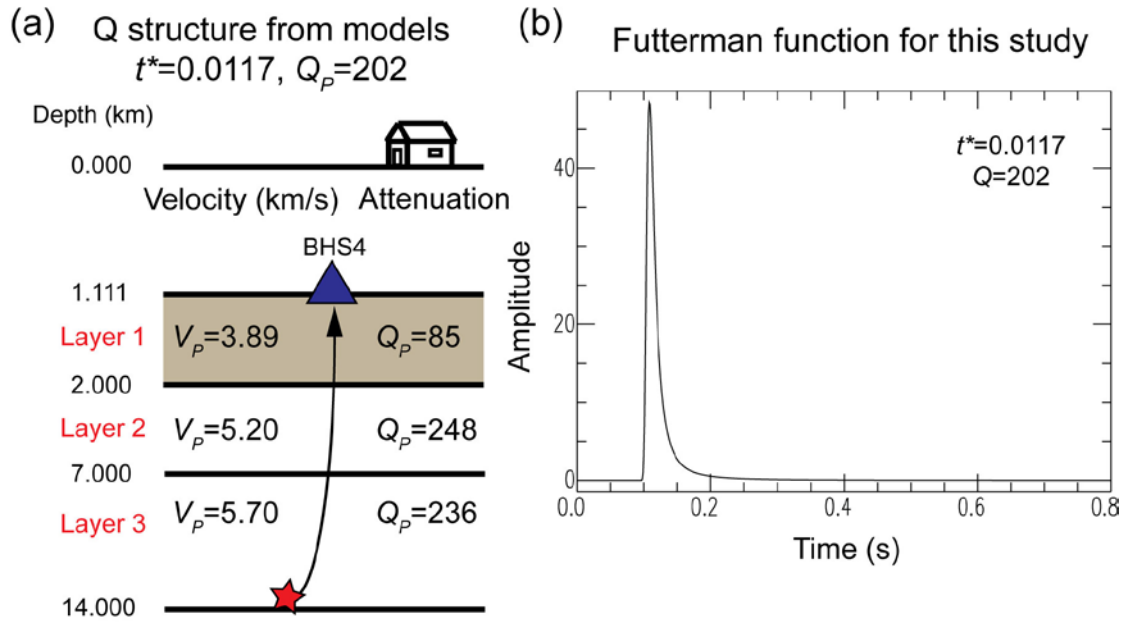


Fig. 9 (a) Displacement spectra of the observed Z-component (P waves) for the smallest (red curve) and largest (blue curve) events in Clusters A (left), B (middle), and C (right). The multi-taper spectral method (Thomson 1982) was used for calculating spectra. The red and blue dash-dot lines indicate the noise level of the smallest and largest events in the clusters, respectively. (b) Spectral ratios between the largest and smallest events for the three clusters. Appreciable drops in the ratio can be seen over the frequency band from 20 to 50 Hz for signal-to-noise ratios greater than 3.0.

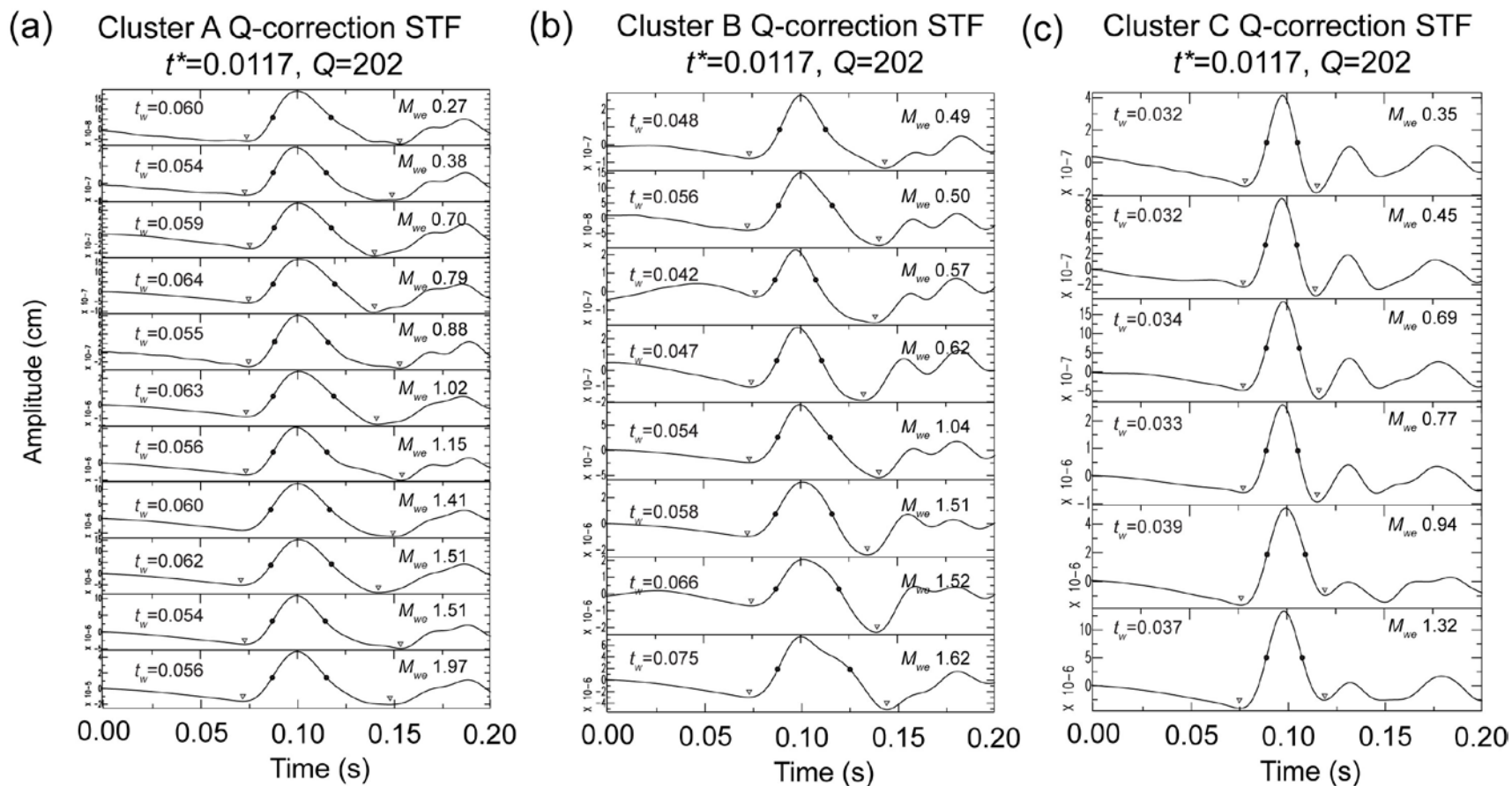
Evidence for Non-Self-Similarity of Microearthquakes Recorded at Taiwan Borehole Seismometer Array



1
2
3
4
5
6
7
8
9
10
11
12
13
14
15
16
17
18
19
20
21
22
23
24
25
26
27
28
29
30
31
32
33
34
35
36
37
38
39
40
41
42
43
44
45
46
47
48
49
50
51
52
53
54
55
56
57
58
59
60

Fig. 10 (a) Q -model ($Q_p = 202, t_p^* = 0.0117$) used in this study. (b) The Futterman function determined for the Q -structure shown in Fig. 10(a).

Evidence for Non-Self-Similarity of Microearthquakes Recorded at Taiwan Borehole Seismometer Array



1
2 **Fig. 11** STFs determined using the Q -correction method and a zero-phase band-pass filter (5–50 Hz) for (a) Cluster A, (b) Cluster B, and (c)
3 Cluster C. The black dots and grey triangles on each trace indicate the half-pulse width and the left and right bottom amplitudes presented in Fig.
4 6.

Evidence for Non-Self-Similarity of Microearthquakes Recorded at Taiwan Borehole Seismometer Array

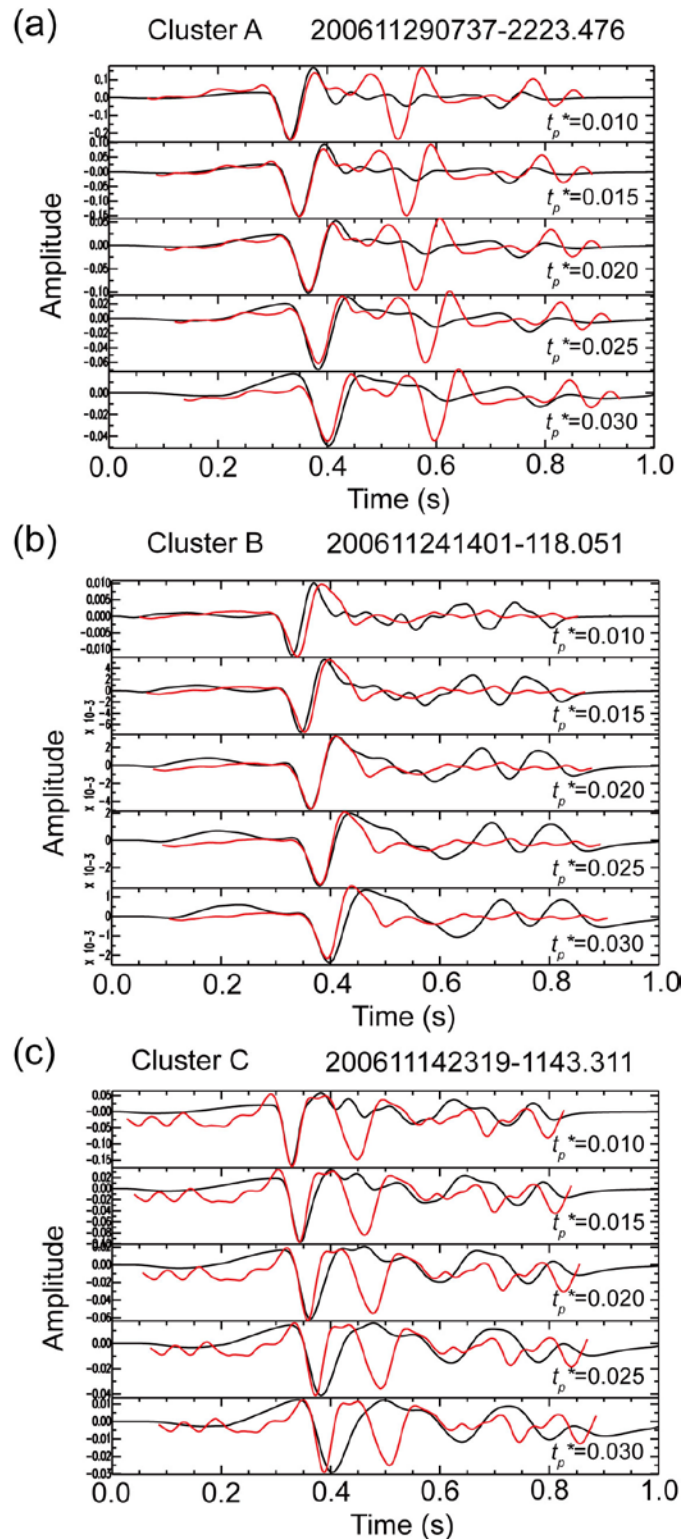


Fig. 12 Comparison of the first pulse of the SH wave, $O_S(t)$, on the T -component (red lines) and $O_P(t) * F(t; 3t_p^*)$, convolution of the P wave on the vertical component and $F(t; 3t_p^*)$ with a given t_p^* (black lines), for (a) Cluster A, (b) Cluster B, and (c) Cluster C.

Evidence for Non-Self-Similarity of Microearthquakes Recorded at Taiwan Borehole Seismometer Array

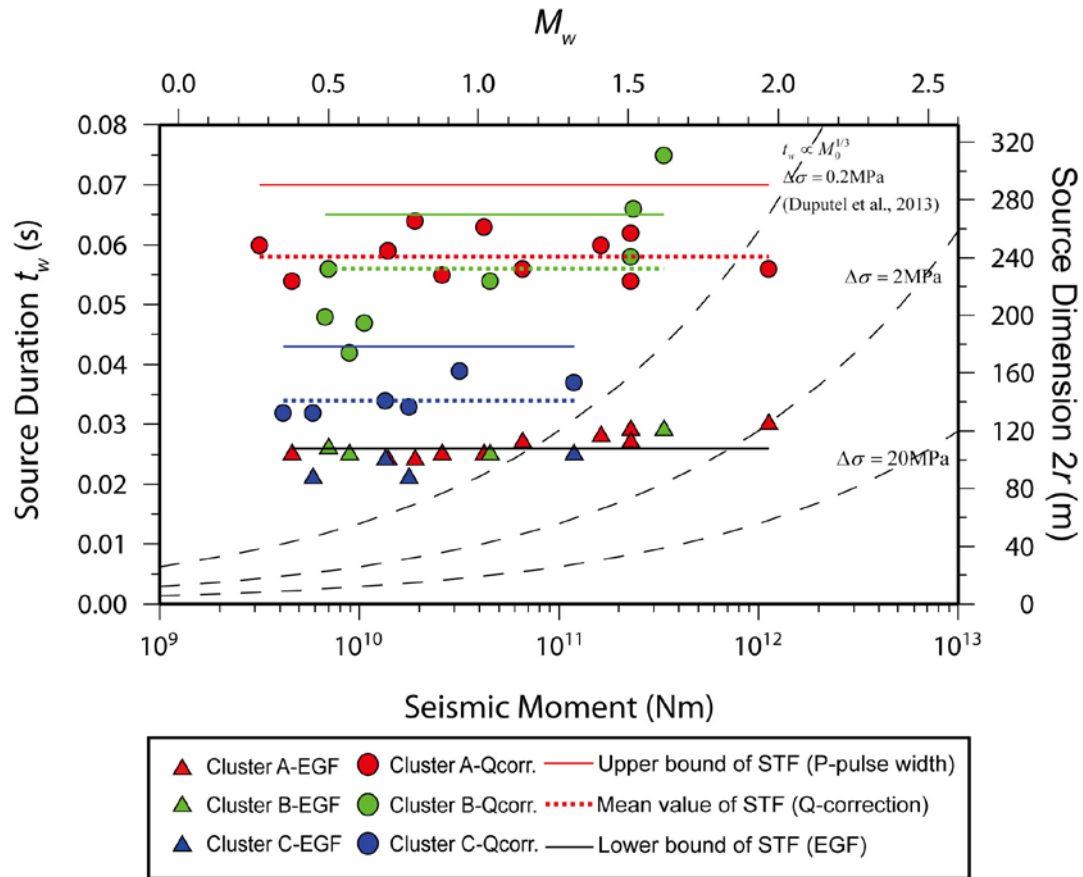
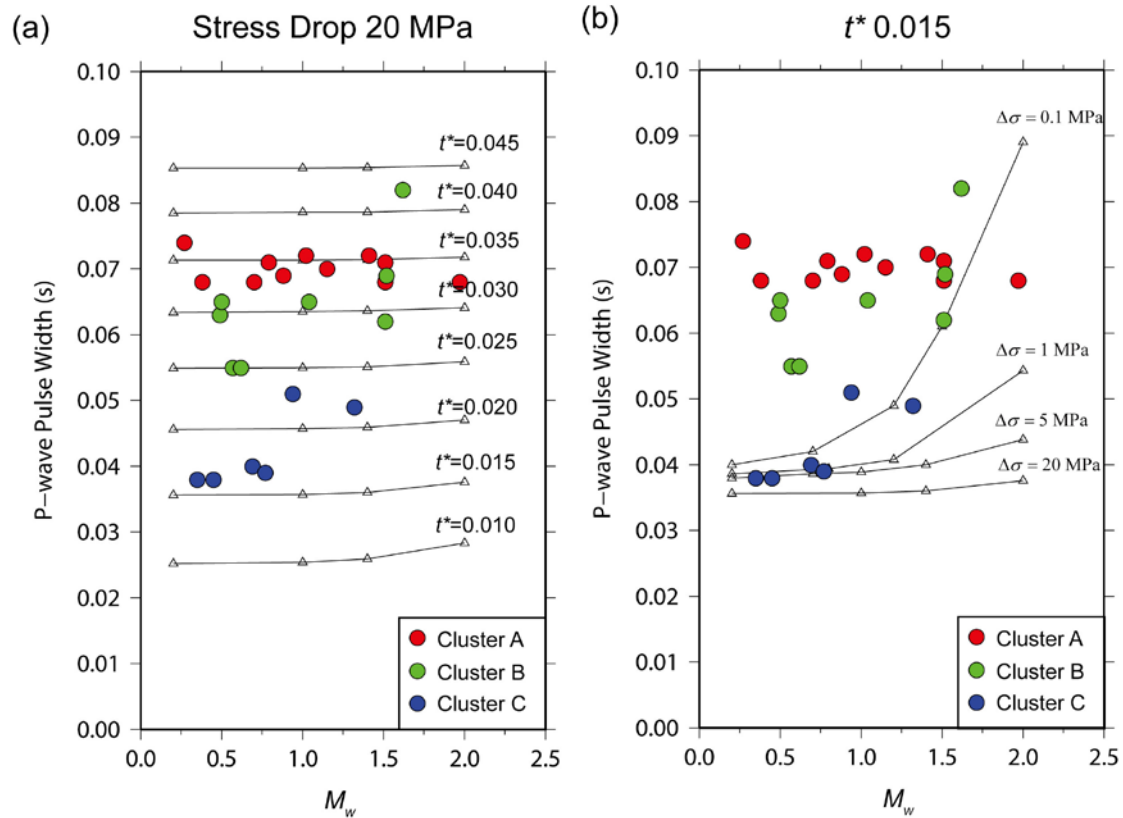


Fig. 13 Source durations of events in Clusters A, B, and C estimated using the three different methods. The red, green, and blue symbols denote the results for Clusters A, B, and C, respectively. The circles indicate the source durations (t_w) estimated using the Q -correction method. The dotted lines show the mean values for each cluster, and the triangles show the source durations (t_w) estimated using the EGF method, with the average indicated by a solid black line. The red, green, and blue solid lines represent the upper bounds of the STF duration for Clusters A, B, and C, respectively. The dashed black curves show the self-similarity relation ($t_w \propto M_0^{1/3}$) for constant stress drops of 0.2, 2, and 20 MPa. The curve labelled as $\Delta\sigma = 0.2$ MPa is close to that presented by Duputel et al. (2013). The right vertical axis gives the source dimension.

Evidence for Non-Self-Similarity of Microearthquakes Recorded at Taiwan Borehole Seismometer Array

1



2

3

4

5

6

7

8

9

10

11

12

13

14

15

16

17

18

19

20

21

22

23

24

25

26

27

28

29

30

31

32

33

34

35

36

37

38

39

40

41

42

43

44

45

46

47

48

49

50

51

52

53

54

55

56

57

58

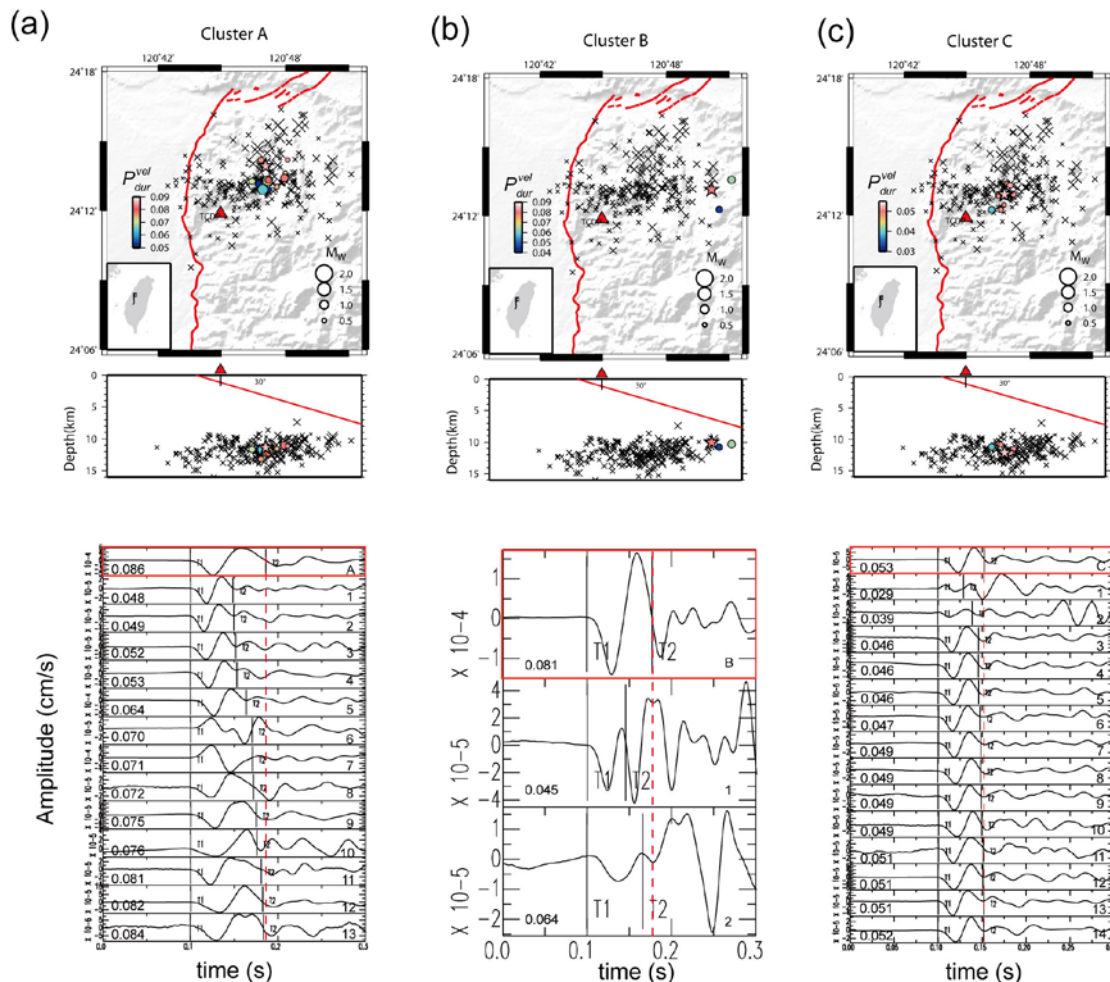
59

60

Fig. 14 (a) Comparison between the computed P-pulse widths of an event with a stress drop of 20 MPa and observations. The triangles and lines represent the computed P-pulse widths for a suite of t_p^* from 0.010 to 0.045 s. Red, green, and blue circles indicate the observed pulse widths of the P waves for Clusters A, B, and C, respectively. (b) Comparison between the computed P-pulse widths of an event with a t_p^* of 0.015 and observations. The triangles and lines represent the computed P-pulse widths for a suite of stress drop from 0.1 to 20 MPa.

Evidence for Non-Self-Similarity of Microearthquakes Recorded at Taiwan Borehole Seismometer Array

1



2

3

Fig. 15 Events near the cluster with a shorter duration P_{dur}^{vel} compared to an event in (a) Cluster A, (b) Cluster B, and (c) Cluster C. Locations of the nearby events (circles) and the cluster (asterisk) are shown in the upper section. The color and size of the circles indicate the duration P_{dur}^{vel} and magnitude of the events, respectively. The crosses are the events in the background. The Z-component velocity waveforms for the event inside the cluster (top trace in the red box) and the nearby events are shown in the lower section. The duration P_{dur}^{vel} is measured between T1 and T2 markers. The red dashed line is drawn at the T2 marker of the cluster event as a reference. The estimated P_{dur}^{vel} is indicated on each trace.

12

13

14

15

16

17

18

19

20

21

22

23

24

25

26

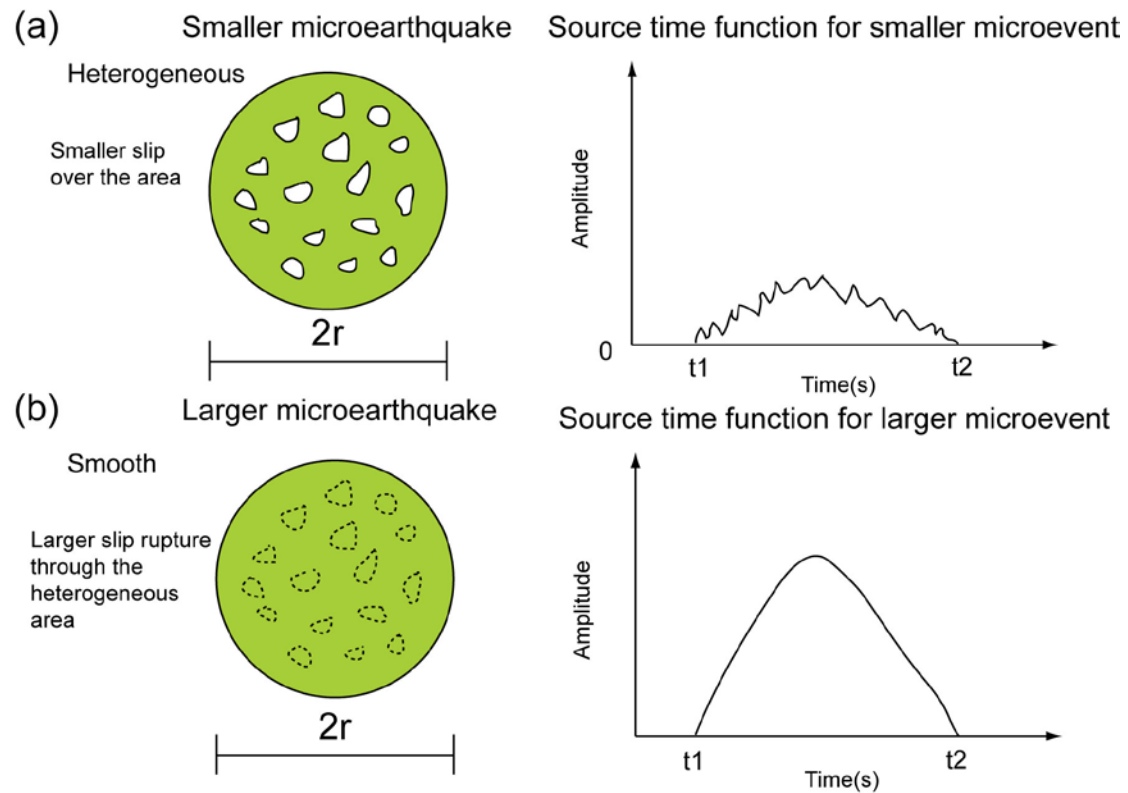
27

28

29

30

Evidence for Non-Self-Similarity of Microearthquakes Recorded at Taiwan Borehole Seismometer Array



1
2 **Fig. 16** Schematic of (a) small and (b) large earthquakes, motivated by the spectral
3 ratios shown in Fig. 9. The irregular white circles represent stronger asperities that
4 delay or stop rupture for small events, exciting high-frequency energy (illustrated in
5 the right column). Schematic of slip distributions are shown on the left, and the
6 associated STFs are shown on the right.
7

Appendix A

Records at the seven TCDPBHS stations, BHS1-BHS7, of the largest event

Fig. A1 shows the records at the seven TCDPBHS stations, BHS1-BHS7, of the unfiltered *Z*-component (P wave) velocity seismograms of the largest event. The records are similar for all the stations, as expected given the distance of 10-15 km to the sources and the spatial extent of the array of ~300 m.

Appendix B

Effect of band-pass filter on pulse width measurements.

Fig. A2(a) shows a series of triangular pulses representing STFs. Fig. A2(b) shows the waveforms after we applied a zero-phase band-pass filter with a pass band of 5 to 50 Hz to the STFs in Fig. A2(a). When the pulse width is small, the broadening is appreciable, but as the pulse width increases, broadening becomes relatively insignificant. Fig. A2(c) shows the relation between the given pulse width t_w (Fig. A2a) and the pulse width of the filtered waveforms (Fig. A2b) estimated using the method described in the text. Fig. A2(d) shows the difference between the estimated and given pulse widths.

Evidence for Non-Self-Similarity of Microearthquakes Recorded at Taiwan Borehole Seismometer Array

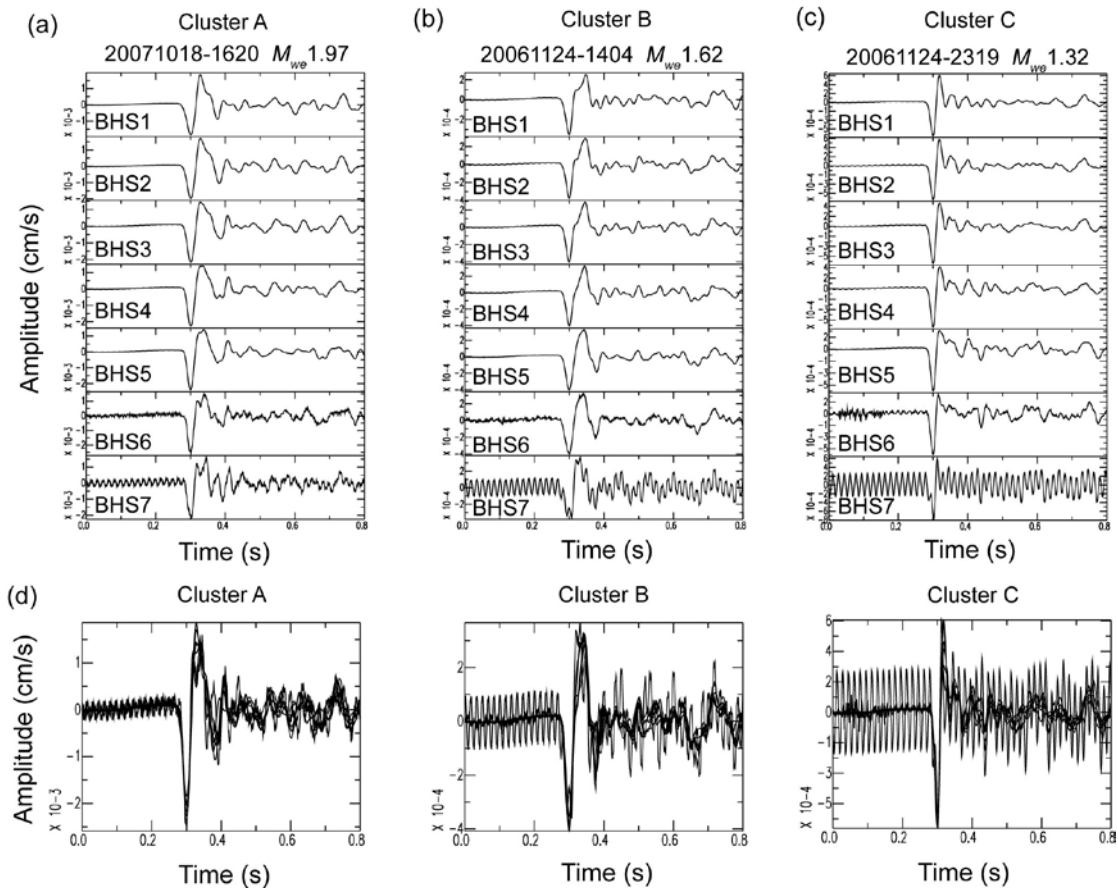


Fig. A1 Records at the seven TCDPBHS stations, BHS1-BHS7, of the unfiltered Z-component (P wave) velocity seismograms of the largest event of (a) Cluster A, (b) Cluster B, and (c) Cluster C. The records are similar for all the stations, as expected given the distance of 10-15 km to the sources and the spatial extent of the array of ~300 m. (d) Comparison of the P wave records for Clusters A (left), B (middle), and C (right).

Evidence for Non-Self-Similarity of Microearthquakes Recorded at Taiwan Borehole Seismometer Array

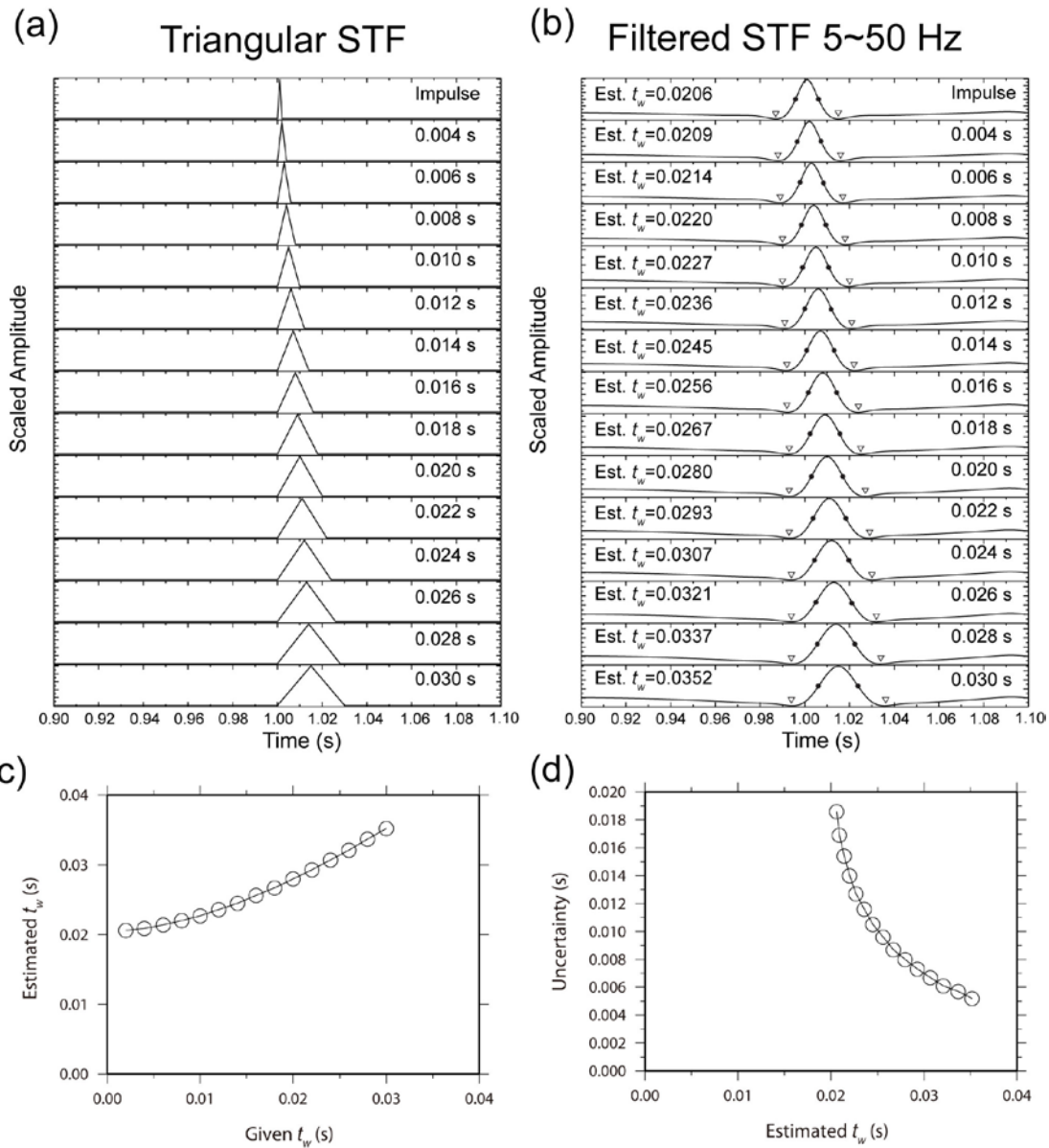


Fig. A2 Effect of the band-pass filter on the pulse width measurement of the STF. (a) Triangular STF with t_w ranging from 0.002 to 0.030 s. (b) Waveforms filtered by a zero-phase band-pass filter with a pass band of 5 to 50 Hz. The pulse widths are measured using the method illustrated in Fig. 6 and indicated on the left of the waveforms. (c) Comparison of the given t_w and estimated t_w . (d) The difference between the estimated and the given t_w as a function of the estimated t_w .