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Statistical properties of low-frequency earthquakes triggered by large earthquakes in southern Taiwan

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A B S T R A C T

The recent discovery of triggered tremors (TTs) and low-frequency earthquakes (LFEs) in various tectonic environments provides an opportunity for studying the fundamental properties and physical mechanisms of deep tectonic tremor. Here, we quantify the relationship between TTs and LFEs beneath the Central Range in southern Taiwan and their statistical properties during the teleseismic waves of six large distant earthquakes. Using waveforms of 11 LFEs triggered by the 2005 M_w 8.6 Nias earthquake as templates, we scan through 12 hours of waveform data around six mainshocks and identify a total of 783 LFEs. The LFEs were mainly located in a compact region between 12 and 36 km in depth near the Chaochou–Lishan Fault. Most of LFEs occurred within TT during the passage of large-amplitude surface waves, and the increase of the LFE rate during the surface waves is statistically significant. The LFE rates do not follow an Omori's type decay, but rather abruptly return to the background rate immediately after the surface-wave passage. These findings suggest that LFEs do not trigger any additional LFEs at later times and are primarily driven by an external forcing. Our observations are consistent with the inference that TTs consist of many reoccurring LFEs.

1. Introduction

Tremor away from volcanoes (Obara, 2002) and triggered by seismic waves, termed triggered 'non-volcanic' or 'deep tectonic' tremor (TT), reflects shear slip on deep active faults driven by transient dynamic stresses (Peng and Gomberg, 2010, and references therein). Relative to regular earthquakes, TTs have been exclusively observed near major plate boundaries: California (Ghosh et al., 2009; Peng et al., 2009, 2010), Japan (Miyazawa and Brodsky, 2008; Chao et al., 2013), New Zealand (Fry et al., 2011), Vancouver Island (Rubinstein et al., 2007, 2009), Haida Gwaii (or Queen Charlotte) Island (Aiken et al., 2013), Cuba (Peng et al., 2013) and Taiwan (Tang et al., 2010; Chao et al., 2012). It is well known that the durations of triggered and ambient tremors are longer than ordinary earthquakes, and they are dominated by low-frequency seismic energy of 1–10 Hz, as compared with regular earthquakes (Nadeau and Dolenc, 2005; Peng et al., 2008).

Low-frequency earthquake (LFE), a new class of seismic event, was first identified by the Japan Meteorological Agency (JMA) in their seismicity catalog in southwest Japan (Katsumata and Kamaya, 2003). Shelly et al. (2006) hypothesized that the reduction of effective stress due to elevated pore-fluid pressure might help promote LFE and tremor generation, and tremors could be considered as a swarm of many LFEs (Shelly et al., 2007). Recent studies of repeating LFEs families along the Parkfield–Cholame section of the San Andreas Fault also confirmed that TTs can be largely explained by the same LFE families that occurred during ambient tremor episodes (Shelly et al., 2011). Tang et al. (2010) identified LFEs within TTs beneath the southern Central Range of Taiwan and further confirmed that TTs consist of many LFEs. The hypocentral depths of these LFEs range from 12 km to 38 km and the epicenters are near the Chaochou–Lishan Fault (CLF), a major reverse fault in Taiwan. Local seismic tomography (Wu et al., 2007) reveals a relatively high ratio of P-to-S wave velocity (V_p/V_s ratio) area near the hypocenter of LFEs, suggesting the existence of pore fluids.

The primary objective in this study is to further investigate the statistical properties of LFEs in Taiwan around several large teleseismic mainshocks and their relationship with TTs observed in recent studies (Chao et al., 2012, 2013). *Note that in this paper*

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we do not attempt to resolve the issue as to whether LFEs are fundamentally different from TT, or simply higher amplitude or more resolvable parts of TT. Our observations, as will be shown below, are consistent with TT being composed of multiple LFEs. But for convenience, we will continue to use the term LFE for the repeatable waveforms that we identify with a matched-filter scanning approach.

We focus on the southern Central Range in Taiwan, mainly because tremors in this region have been repeatedly triggered by the surface waves of recent large teleseismic earthquakes (Peng and Chao, 2008; Tang et al., 2010; Chao et al., 2012). The conventional technique for locating tremor often uses differential S-wave arrival times from tremor envelopes (Obara, 2002; Chao et al., 2012), which have the difficulty of obtaining accurate tremor location, especially the depth. Identifying P- and S-waves of LFEs within tremors provides another way to accurately locate tremor sources (e.g., Shelly and Hardebeck, 2010). However, direct investigation of LFEs is difficult due to low amplitudes and overlapping P-wave arrivals. Hence, we use waveforms of LFEs triggered by the 2005 Nias M_w 8.6 earthquake (Tang et al., 2010) as templates to scan through several hours of data before and after triggering mainshocks. In the following section, we first describe briefly our LFE scanning technique and how we quantify the repeating LFEs families among different triggering events. Then we analyze their statistical properties and discuss the connection between the detected LFEs and TTs.

2. Data and analysis processes

We focused on six large teleseismic earthquakes (Table S1) that have triggered tremors in southern Taiwan (Chao et al., 2012). Their sources are close to centroid of LFEs triggered by the 2005 Nias earthquake (Tang et al., 2010). We utilized the waveform data recorded by the Broadband Array in Taiwan for Seismology (BATS), operated by the Institute of Earth Sciences, Academia Sinica and the Central Weather Bureau, and by the Central Weather Bureau Seismic Network (CWBSN) (Fig. 1). The BATS stations are equipped with broadband sensors (Trillium) and digital recorders (Q330), while the CWBSN stations have 1 Hz S13 short-period sensors (Teledyne Geotech).

The analysis process generally follows that of Tang et al. (2010) and is described here. We analyzed the waveform data 6 h before and after the origin time of the six mainshocks. The selected time period not only includes the large-amplitude teleseismic waves, but also spans a long-enough time period to provide an estimation of the background LFE rate. Because sampling rates of the stations of the BATS (20 sample s^{-1}) and the CWBSN (100 sample s^{-1}) are different, we first cut different time segments of data for each earthquake according to the duration of surface waves, apply a 2–8 Hz band-pass filter to remove long-period surface waves and then re-sample to 20 sample s^{-1}. We basically focused on TT signals within 2–8 Hz band-passed-filtered waveforms according to previous studies (Chao et al., 2012, 2013).

Next we used the same 11 LFEs templates triggered by the 2005 M_w 8.6 Nias earthquake (Tang et al., 2010) as templates to scan through the 12-h waveform for all six mainshocks. This is the so-called matched filter technique (Gibbons and Ringdal, 2006) that has been widely used to detect LFEs within tremor (Shelly et al., 2007; Brown et al., 2008, 2009; Tang et al., 2010) and missing early aftershocks following large earthquakes (Peng and Zhao, 2009). In detail, we first cut 3 s before and after the S-wave arrivals of each template. We then calculated the correlation coefficient (CC) values for the three-component seismograms among all stations for each time window that steps forward at 0.05 s (1 sample). We also computed a network averaged CC value for all time windows. Lastly, we applied a threshold based on five times the median absolute deviation (MAD) of each CC values to detect candidate events.

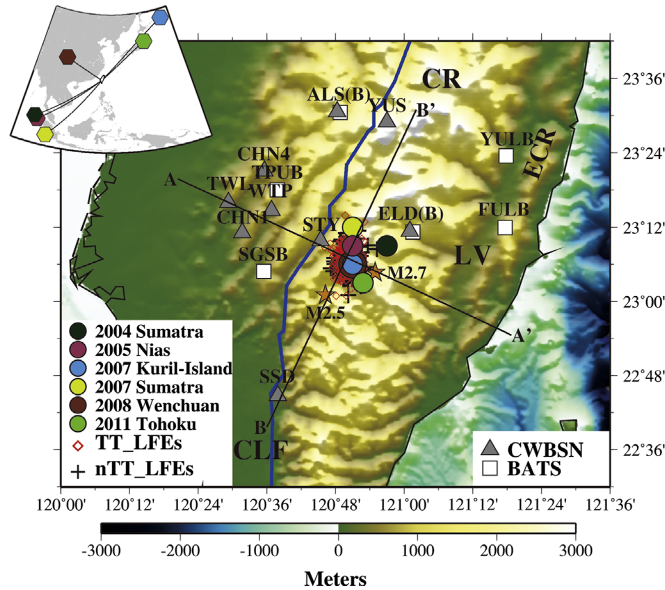

Fig. 1. Epicenters of low-frequency earthquakes (LFEs) within tremors triggered by large teleseismic events. The solid circles with different colors represent the location of triggered tremors reported by Chao et al. (2012). The open red diamonds and crosses correspond to location of LFEs within (TT_LFEs) and out of triggered tremors (TTs) segments (nTT_LFEs), respectively. Two orange stars represent local earthquakes occurred during the 2004 M_w 9.0 Sumatra earthquake and their origin times are marked in Fig. 3. The blue line represents the strike of Chaochou–Lishan fault (CLF). Profiles AA′ and BB′ are perpendicular and parallel to the Central Range (CR), respectively, and the cross-section plots are shown in Fig. 5. Seismic stations of BATS and CWBSN are denoted by white squares and gray triangles, respectively. The inset shows the epicentral locations (hexagons) and the ray paths of the six triggering mainshocks. LV: the Longitudinal Valley. ECR: Eastern Costal Range. (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)

After detecting candidate events, we added 2 s (1 s before and after the original 6-s window) to the waveforms of the newly detected events. We decreased the time window of the template waveforms to 4 s and scanned the template through the 8-s time segment to detect S-wave arrivals on the horizontal components. To identify as many LFEs as possible without sacrificing the robustness of the results, we set a threshold of 12 times the MAD for waveform detection. The same procedure was applied to search for P-wave arrivals in vertical component. Fig. 2 shows an example of detected P and S waves of a LFE for the 2004 M_w 9.0 Sumatra earthquake. More examples are shown in Fig. S1.

Once we obtained the P- and S-wave arrivals of the newly detected events, we determined the differential travel times of the P- and S-waves based on waveform cross-correlations from the 4-s window. We then used hypoDD double-difference algorithm (Waldhauser and Ellsworth, 2000) with a 1-D velocity model under southern Taiwan (Tang et al., 2010) to locate the detected LFEs. We required at least three stations with both P- and S-wave arrivals and eight link pairs of the P or S wave in locating LFEs.

To examine the statistical significance of detected LFEs during large-amplitude surface waves, we computed a β-statistic value (Aron and Hardebeck, 2009), which is a measure of the difference between the observed number of events after the mainshock and the expected number from the background rate before the mainshock. We use the 6-h window before the mainshock to compute the background LFE rate. The time window after the mainshock starts from the predicted P-wave arrival and stops at the end of TT period as reported by previous studies (Chao et al., 2012, 2013). The β-value is defined as follows:

$$\beta = \frac{\mathrm{Na} - N\mathrm{Ta}/T}{\sqrt{N(\mathrm{Ta}/T)(1 - \mathrm{Ta}/T)}} \quad (1)$$