Shallow Slow Earthquake Episodes Near the Trench Axis Off Costa Rica

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Key Points:

- Shallow very low frequency earthquakes (VLFEs) and tremors are detected off Costa Rica near the Middle America Trench
- Distribution of VLFEs and tremors is spatially correlated with slow slip events
- Scaled energy of shallow slow earthquakes off Costa Rica is 10^{-9}–10^{-8}, which is similar to those in Nankai

Abstract

Slow earthquakes are mainly distributed in regions surrounding seismogenic zones along the plate boundaries of subduction zones. In the Central American subduction zone, large regular interplate earthquakes with magnitudes of 7–8 occur repeatedly around the Nicoya Peninsula, in Costa Rica, and a tsunami earthquake occurred off Nicaragua, just north of Costa Rica, in 1992. To clarify the spatial distribution of various slip behaviors at the plate boundary, we detected and located very low frequency earthquakes (VLFEs) around the Nicoya Peninsula using a grid-search matched-filter technique with synthetic templates based on a regional three-dimensional model. VLFEs were active in September 2004 and August 2005, mainly near the trench axis, updip of the seismogenic zone. The distribution of VLFEs overlaps with large slip areas of slow slip events. Low frequency tremor signals were also found in high-frequency seismogram envelopes within the same time windows as detected VLFEs; thus, we also investigated the energy rates of tremors accompanied by VLFEs. The range of scaled energy, which is the ratio of the seismic energy rate of a tremor to the seismic moment rate of accompanying VLFE and related to the rupture process of seismic phenomena, was 10^{-9}–10^{-8}. The along-dip separation of shallow slow and large earthquakes and the range of the scaled energy off Costa Rica are similar to those in shallow slow earthquakes in Nankai, which shares a similar thermal structure along the shallow plate boundary.

Plain language summary

Slow earthquakes with slower rupture speeds compared to those of regular earthquakes generally occur on the plate boundaries of subduction zones. We detected and located very low frequency earthquakes (VLFEs), which are a type of slow earthquake, off Costa Rica. The VLFEs occurred at a depth range of 6–10 km, and their spatial distribution is correlated with slow slip events, another type of slow earthquakes. The spatial separation of slow and large regular earthquakes is common to the Nankai subduction zone. Low frequency tremor signals, which are also classified as slow earthquakes, are also found in seismograms at higher frequencies within the same time windows of detected VLFEs. We also estimated the ratio of energy rates of tremors to moment rates of VLFEs, which relates to the rupture process of seismic phenomena. The ratio is 10^{-9}–10^{-8} off Costa Rica, similar to that in shallow slow earthquakes in the Nankai subduction zone.

1. Introduction

Slow earthquakes are mainly observed in regions surrounding seismogenic zones, which are the areas that rupture in large regular earthquakes, along the plate boundaries of subduction zones (e.g., Obara & Kato, 2016) or strike-slip faults (e.g., Nadeau & Dolenc, 2005; Wang & Barbot, 2020). Various types of slow earthquakes, such as low frequency tremors (tectonic tremors; e.g., Obara, 2002), low frequency earthquakes (LFEs; e.g., Shelly et al., 2006), very low frequency earthquakes (VLFEs; e.g., Obara & Ito, 2005), and slow slip events (SSEs; e.g., Dragert et al., 2001) have been observed in many subduction zones. Although these slow earthquake phenomena occur without correlation in some cases (Hutchison, 2020; Hutchison & Ghosh, 2016, 2019), they often correlate spatiotemporally, which is termed episodic tremor and slip (ETS). ETSs were observed in deep Cascadia (e.g., Ghosh et al., 2015; Rogers & Dragert, 2003) and deep Nankai (e.g., Ito et al., 2007; Obara, 2011), for example. Recently, in the Nankai subduction zone, pore fluid pressure changes have been observed during tremor and VLFE activities and are considered to reflect shallow SSEs by offshore borehole observations (Araki et al., 2017; Nakano et al., 2018). The hypocenters and focal mechanisms of slow earthquakes are generally consistent with shear slip on the plate boundaries. VLFE episodes and SSEs occur in almost identical source regions and their temporal changes of moment release are similar during as ETS, therefore VLFE episodes are considered as proxies for SSEs (Ghosh et al., 2015; Ito et al., 2007; Nakano et al., 2018; Yokota & Ishikawa, 2020). In summary, the distribution of slow earthquakes is related to large earthquake slip areas, interplate coupling, or fluid distribution (e.g., Baba et al., 2020b; Ghosh et al., 2015; Obara & Kato, 2016).

In the Central American subduction zone, the Cocos plate subducts beneath the Caribbean plate at the Middle America Trench at a rate of approximately 80 mm/year (Figure 1b; referred from NUVEL1A; DeMets et al., 1994). In this subduction zone, large thrust-type earthquakes with a moment magnitude (M_W) of 7–8 occur with a recurrence interval of tens of years around the Nicoya Peninsula, in Costa Rica (light blue areas in Figure 1a; Protti, 1995; Yue et al., 2013). The coseismic slip areas of these large earthquakes are distributed at a depth range of 10–35 km beneath the peninsula and off the coast. The latest large earthquake with M_W of 7.6 occurred on 5 September, 2012 (green contour lines in Figure 1a; Yue et al., 2013). In the vicinity, a tsunami earthquake with M_W of 7.6 also occurred off Nicaragua, just north of Costa Rica, on 2 September, 1992 (dark blue area in Figure 1a; Satake, 1994).

In addition to large regular and tsunami earthquakes, slow earthquakes also occur around the Nicoya Peninsula. The Global Navigation Satellite System data revealed that SSEs with M_W of 6.6–7.2 occur at intervals of 21.7±2.6 months (Jiang et al., 2012; Xie et al., 2020). The large slip area of the SSE in 2007 was separated into downdip and updip areas by the seismogenic slip area (Jiang et al., 2012, 2017; Outerbridge et al., 2010). The spatiotemporal change in relation to the 2012 M_W 7.6 earthquake was investigated by previous studies (Dixon et al., 2014; Voss et al., 2017), and an SSE preceded the 2012 M_W 7.6 earthquake (Voss et al., 2018) in the almost same area of the 2007 SSE, similar to both the slow slip before the 2011 Tohoku earthquake in Japan (Ito et al., 2013; Kato et al., 2012) and the slow slip before the 2014 Iquique earthquake in Chile (Kato & Nakagawa, 2014; Ruiz et al., 2017).

By using high-frequency (>1 Hz) seismograms, Brown et al. (2009) and Outerbridge et al. (2010) located LFEs and tremors in 2007, respectively (Figure 1a). The tremors and LFEs were located in almost the same area, downdip of the seismogenic zone. Although tremors and LFEs were temporally correlated with the SSE, the location of tremors and LFEs were separated from the large slip area of the 2007 SSE. On the other hand, Walter et al. (2011) located many tremors in the offshore region from 2007 to 2009. Walter et al. (2013) also found that VLFEs appeared in seismograms in a frequency range of 0.02–0.05 Hz and were temporally correlated with tremors in the time period of the 2008 SSE. Based on beamforming analysis, they estimated the propagation direction and the propagation speed of VLFE signals and suggested that VLFEs also occurred in offshore areas. Due to the limitations of a conventional analysis, however, epicenters of VLFEs in offshore areas were not located. Therefore, the detailed spatial distribution of VLFEs off Costa Rica is still not well understood.

The spatial variation of slow and large regular earthquakes can reflect the spatial heterogeneity of the frictional conditions on the plate boundary (e.g., Baba et al., 2020b). To clarify the spatial relationship between slow and large regular earthquake distribution around the Nicoya Peninsula, an accurate spatial distribution of VLFEs is needed. Thus, we detected VLFEs around the Nicoya Peninsula using a temporary broadband seismic network from August 2004 to January 2006 because signals of VLFEs are less attenuated than those of tremors and propagate longer distances. The method is based on the matched-filter technique. Template waveforms from possible VLFE locations were evaluated by numerical simulations of seismic wave propagation using a regional three-dimensional (3D) velocity structure model. In addition, scaled energy is an informative parameter for the rupture process of seismic phenomena (Kanamori & Rivera, 2006). Although scaled energies of slow earthquakes around Japan were well investigated by previous studies (e.g., Ide & Yabe, 2014; Yabe et al., 2019; 2021), those in Costa Rica was not estimated. Therefore, we also estimated the seismic energy rate functions of tremors accompanied by VLFEs by using high frequency (2–8 Hz) seismograms to evaluate the scaled energy of slow earthquakes around the Nicoya Peninsula.

2. VLFE analysis

2.1. Data and method

2.1.1. Data

We used waveforms of a temporary seismic network, Tomography Under Costa Rica and Nicaragua (TUCAN; Abers & Fischer, 2003), recorded from August 2004 to January 2006. There were 49 broadband seismic stations in four lines (Figure 1b). In this study, we mainly used data from stations in Costa Rica (shown in Figure 1a) for VLFE analysis. After removing instrumental responses, the seismograms for VLFE detection were resampled at one sample per second. We applied a bandpass filter in the frequency range of 0.02–0.05 Hz (e.g., Ghosh et al., 2015; Ito et al., 2009; Takemura et al., 2019), because this frequency band is less affected by microseismic noises (e.g., Hasselmann, 1963; Kaneko et al., 2018). We verified that the large amplitude surface waves are generally matched well between observed and synthetic waveforms in a higher frequency range (Figure S1; 0.02-0.06 Hz).

2.1.2. Matched-filter technique

The detection procedure used for VLFEs is similar to that used in our previous study (Baba et al., 2020a). We used only the vertical component seismograms because the horizontal component seismograms of many stations were noisy, and it was difficult to find VLFE signals (Figure S2). We placed 175 virtual source grids on the Cocos Plate boundary at a uniform interval of 0.1° (Figure 2a) and computed synthetic waveforms from these source grids to the stations in Costa Rica using an open-source seismic wave propagation code (OpenSWPC; Maeda et al., 2017). We used a three-dimensional velocity structure model constructed by combining CRUST 1.0 (Laske et al., 2013), Slab2 (Hayes et al., 2018), and ETOPO1 (Amante & Eakins, 2009), setting the minimum S-wave velocity in the solid columns to 1.0 km/s. We adopted the values of a mean oceanic slab structure (Christensen & Salisbury, 1975) for the physical parameters of the subducting slab (Table S1). For the physical parameters of the other layers except for the slab, we used the values of CRUST 1.0, and the default parameter set of OpenSWPC. The model covered the region enclosed by the red line (Figure 1b), which was discretized by a uniform grid interval of 0.2 km. The assumed VLFE moment rate function was a Küpper wavelet with a source duration of 15 s and an M_W of 4.0 (Figure 4 of Maeda et al., 2017). Since focal mechanisms of VLFEs are consistent with shear slip on the plate boundaries in previous studies (Cascadia: Ghosh et al., 2015; Nankai: Ito et al., 2009; Nakano et al., 2018; Sugioka et al., 2012; Takemura et al., 2019), the focal mechanism at each source grid was assumed to be consistent with the geometry of the plate boundary of Slab2 and the plate motion model, NUVEL-1A (DeMets et al., 1994). The time window of each template was set to 150 s from the event origin time. Hereafter, we simply refer to these synthetic waveforms as template waveforms. Examples of template waveforms at updip and downdip source grids are shown in Figures 2b and 2c, respectively. The signal first arrives at MANS and the variation of amplitudes is small for the updip source, whereas signals first arriving at FINA exhibit amplitudes in or near the Nicoya Peninsula that are much larger than in other areas for the downdip source.

We then calculated cross-correlation coefficients (CCs) between the filtered template waveforms and observed seismograms every 1 s. We selected events with station-averaged coefficients larger than a threshold defined as 9.5 times the median absolute deviation (MAD) of the distributions. In order to decrease false detections by non-VLFE signals on the condition that only the vertical component can be used and the station coverage along the azimuth direction is poor, we adopted a strict detection threshold compared to previous studies (e.g., 8 × MAD in Shelly et al., 2007 and Baba et al., 2018 and 9 × MAD in Baba et al., 2020a). The changes of CCs when focal mechanisms or depths of assumed source models are different from the geometry of the plate boundary are shown in Figure S3.

2.1.3. VLFE location and discarding false detections

Although a strict detection threshold was employed, there are false detections that are caused by other signals, such as local or regional regular earthquakes or teleseismic events. To exclude local or regional earthquakes, we compared the origin time of detected events with a catalog of local and regional regular earthquakes constructed by El Observatorio Vulcanológico y Sismológico de Costa Rica, Universidad Nacional (Catálogo de Temblores de Costa Rica, 2004-2006; Protti, personal comm.). We discarded events whose epicentral distances were less than 150 km and origin times were within ±50 s from the local or regional earthquakes listed in this earthquake catalog. To discard false detections by teleseismic events, we removed the events detected between the *P*-wave arrivals and 600 s after *S*-wave arrivals of teleseismic events ($M_W \geq 5$) in the catalog of the United States Geological Survey. The event amplitudes and CCs are positively correlated in general, but events with high amplitudes and low average CCs occasionally appear. These events are considered to be false detections due to teleseismic events absent in the catalogs. Therefore, we did not count events with average CCs below 0.56 and relative amplitudes to templates higher than 0.4 (Baba et al., 2018; 2020a). If the amplitude relative to the template with M_W of 4.0 was smaller than 0.05, we did not count the event because the signal was too small to judge whether the event is truly existed or not.

For the remaining events, we calculated the variance reduction (VR) between the template and observed waveforms. We estimated VRs using only the vertical component seismograms of relatively quiet stations in and around the Nicoya Peninsula (MANS, CABA, FINA, CRUP, and PALM), because differences of amplitude distributions between updip and downdip events are large in these stations:

$$VR = \left[1 - \frac{\sum_i \int \{f_i(t) - c g_i(t)\}^2 dt}{\sum_i \int \{f_i(t)\}^2 dt}\right] \times 100\% \, , \quad (1)$$

where $f_i\,(t)$ and $g_i\,(t)$ are the observed and template waveforms at the i-th station, respectively, and c is the relative amplitude of the observed waveform to the template. We selected events whose VRs were larger than 30%. This threshold is set by trial and error based on visual identifications of VLFEs in the observed data.

After the above procedures, falsely detected events still remained because we only used the vertical component seismograms, and the array configuration was cross shaped. To discard the remaining false detections, we estimated the normalized-and-stacked amplitude, azimuth, and velocity of signal propagation by applying delay-and-sum beamforming (Section 3.1 of Rost & Thomas, 2002; Walter et al., 2013) to vertical component seismograms. After normalizing the waveform of each station by its maximum amplitude in the 150 s time window, we searched for the azimuth and velocity that maximized the stacked amplitude by performing a grid search for the azimuth between 135º – 315º with 1º intervals and the velocity between 2–5 km/s with 0.1 km/s intervals. We first used the along-strike stations in both Costa Rica and Nicaragua (brown inverted triangles in Figure 1b) to discard teleseismic events. The amplitudes of Costa Rican VLFEs at the Nicaraguan stations are generally very small compared with those in the Costa Rican stations due to geometrical spreading, but amplitudes for teleseismic events are similar. Therefore, we selected events whose stacked normalized amplitude normalized by the number of stations was smaller than 0.6 because events with large stacked signals are suspected to be teleseismic earthquakes (Figure S4). We then conducted another beamforming analysis for the remaining events using the same stations as the matched-filter analysis, and selected events whose azimuth was 200–230º. Finally, to avoid duplicate detection, only one event was counted every 60 s from the remaining VLFE candidates. We only counted the event whose averaged CC was the highest spatiotemporally.

2.1.4. Estimation of the moments of events

We estimated the source durations of detected VLFEs by comparing template waveforms with source durations of 10–50 s and an M_W of 4.0 with observed waveforms (e.g., Yabe et al., 2021). The source duration that resulted in the highest values of CC between the observed and template waveforms was adopted.

We also calculated the amplitude of an event relative to the template waveforms using the same method as Baba et al. (2020b). The relative amplitude can be used to calculate the seismic moment of each VLFE. The seismic moment rate of a VLFE was calculated by dividing its seismic moment by its source duration. To evaluate the estimation error of moment rates of VLFEs, we calculated moment rates by assuming the various source durations whose CCs between synthetic and observed waveforms are more than 90% of the maximum CC. Although there are errors in the order of 0.2, we verified that the order of moment rates does not change (Figure S5).

2.2. Results

We detected 68 VLFEs during the analysis period. Example traces of a VLFE located at 85.8°W and 9.4°N are shown in Figure3. The signal of this VLFE first arrives at MANS and propagates to inland stations (top panel of Figure 3). This feature was successfully modeled for the updip templates (Fig. 2b). There is a tremor signal in the frequency range of 2–8 Hz in the same time window (middle and bottom panels of Figure 3). The cumulative number of VLFEs showed significant increases in September 2004 and August 2005 (Figure 4a). In August 2005, an SSE was reported by Jiang et al. (2012); therefore, SSE and VLFE activities were temporally correlated. The M_W and source duration of VLFEs were mainly distributed in 3.4–4.2 and 10–30 s, respectively (Figures 5a, b). The M_W and source duration of VLFEs have a positive correlation (Figure 5c) like shallow VLFEs in Nankai, Japan (Sugioka et al., 2012; Takemura et al., 2019).

Most of the VLFEs (62 events) are distributed where the plate boundary is at a depth range of 6–10 km below the sea level, near the trench axis off the Nicoya Peninsula (Figure 4b), at the updip of the seismogenic zone. The distribution of these VLFEs is consistent with the VLFEs in 2008 suggested by Walter et al. (2013). When locating some events using both vertical and horizontal component seismograms whose signal to noise (SN) ratios are relatively high for the verification of the analysis by using vertical components only, the high CC areas overlap and the epicenters were also located near the trench axis, although there are differences of 0.1–0.2º (Figure S2). The area overlaps with the shallower part of the large slip area of the 2007 SSE (Jiang et al., 2017) or summed SSE slip in 2007–2012 (Dixon et al., 2014). Although the slip distribution of the 2005 SSE was not estimated in previous studies, our results suggest that the 2005 SSE can also have a large slip area near the trench axis, similar to the 2007 SSE. The distribution of VLFEs lies within the gap between large slip areas of thrust-type large interplate earthquakes with an M_W of 7–8 around the Nicoya Peninsula and the 1992 tsunami earthquake with an M_W of 7.6. The distribution of VLFEs also separated from the afterslip area of the 2012 M_W 7.6 earthquake (Malservisi et al., 2015).

The distribution of the CC shows the resolution of the location of VLFEs. By the distribution of CC, it is confirmed that most of the VLFEs were located near the trench axis. CCs for more than half of the events exceeded the threshold only for updip templates (Figure 6a). For several events, CCs exceeded the threshold both updip and downdip of the seismogenic zone with a larger CC in the updip region. The area where CCs are more than 90% of the maximum CC is concentrated only in the updip area (Figure 6b). On the other hand, 6 VLFEs were located at a depth of ~40 km at the downdip of large earthquakes (Figure 4b). Although focal mechanisms may not be thrust-type and the areas where CCs are larger than the threshold are widely distributed, we verified that regular earthquakes listed in the earthquake catalog by El Observatorio Vulcanológico y Sismológico de Costa Rica in the updip and downdip areas are located in the updip and downdip areas respectively by this method (Figure S6). However, we cannot exclude the possibility that such VLFEs occur in the updip region in real because, in such cases, two CC peaks tend to appear both in the updip and downdip (Figure 6c). Of course, there is a possibility that such VLFEs really occur in the downdip region because the locations of such VLFEs were near the locations of previously reported LFEs (Brown et al. 2009) and tremors (Outerbridge et al. 2010). In this study, the SN ratios of VLFEs detected in the downdip region are very low; hence, it is difficult to judge whether such VLFEs occur in downdip or updip, because it is hard to judge which station the signal of the VLFE arrival first due to the similar arrival times at updip stations. The reason for the small number and the low SN ratio of downdip events may be that slow earthquakes in the downdip region were inactive during 1.5 years of the temporary array. To investigate whether deep VLFEs really exist, an analysis with a longer dataset is needed in future work.

3. Estimations of seismic energy rates for tremors accompanied by VLFEs

3.1. Data and method

Tremor signals were also found in the frequency range of 2–8 Hz within the time windows of detected VLFEs (middle panel of Figure 3; Figure S7). It is difficult to locate tremors in the offshore region by using an onshore network because sources of tremors are distant from the network and signals of tremors attenuate strongly compared to VLFE (0.02–0.05 Hz) signals. Based on the spatiotemporal correlation between VLFEs and tremors reported in other regions (e.g., Ghosh et al., 2015; Maeda & Obara, 2009; Tamaribuchi et al., 2019) and the interpretation that VLFEs and tremors are components of broadband slow earthquake phenomena (Gomberg et al., 2016; Hawthorne & Bartlow, 2018; Ide & Maury, 2018), we estimated the energy rate functions of tremors accompanied by VLFEs by assuming that a tremor occurs at the same location as the VLFE (e.g., Yabe et al., 2019; 2021). We simulated the waveforms at the location of a VLFE using the same model which is described in Section 2.1.2 but discretized by a finer grid interval (0.04 km). The simulated envelope shapes are different from observed ones due to a simple pulse source time function (details in the caption of Figure S8), but the arrival times of maximum *S*-wave amplitudes in the frequency range of 2–8 Hz are consistent with observed tremor waveforms (Figure S8). Therefore, we supposed that a VLFE and the corresponding tremor occurred at the same location.

We also used waveforms of the TUCAN network similarly to the VLFE detection. After applying a bandpass filter of 2–8 Hz, the envelope waveforms were calculated by taking the root-mean-square of sums of three-component squared seismograms and a smoothing time window of 3 s (bottom panel of Figure 3). The envelope waveforms were resampled at one sample per second.

3.1.1. Quality factor of the apparent *S*-wave attenuation

To estimate the energy rate functions of tremors accurately, we estimated the quality factor of the apparent *S*-wave attenuation ($Q_{app.}$), based on the coda-normalization method (e.g., Aki, 1980; Yoshimoto et al., 1993). First, we selected some isolated regular earthquakes (Figure S9). To eliminate the effect of differences in source size and site amplification, observed maximum *S*-wave amplitudes were normalized by averaged coda amplitudes within a lapse time of 80-90 s. The coda-normalized maximum *S*-wave amplitude of the *i*-th earthquake at the *j*-th station (A_{ij}) and the distance between the hypocenter of the *i*-th earthquake and *j*-th station (L_{ij}) have the following relationship (Takemura et al., 2017):

$$\ln\left(L_{ij}A_{ij}\right) = -\frac{\pi f_c Q_{app.}^{-1}}{V_S} L_{ij} + C' , \quad (2)$$

where V_s is the *S*-wave velocity (assuming 3.5 km/s in this study; Maeda & Obara, 2009; Yabe et al., 2019; 2021), f_c is the central frequency (assuming 5 Hz in this study), and C' is a constant. By solving Equation (2) by the least-squares method, we estimated $Q_{app.}^{-1}$ as $10^{-2.42}$ (Figure 7a).

3.1.2. Site amplification factor

We estimated the site amplification factor at 2–8 Hz using relative coda amplitudes (e.g., Maeda and Obara, 2009). Coda amplitudes at a certain time window generally depend on the source size and site amplification (e.g., Chapters 2 and 3 of Sato et al., 2012). Therefore, the ratio of the coda wave amplitude at a station to that at a reference station for the same event depends only on the site amplification factor relative to a reference station.

We calculated the ratios of the coda amplitudes for each station to those of MANS (reference station) for each regular earthquake used in Section 3.1. The time window for evaluating relative coda amplitudes is the same as that in coda-normalization in Section 3.1. Then we calculated the average of the coda amplitude ratios of all earthquakes for each station. The estimated relative site amplification factors at each station used in the estimations of the energy rate functions of tremors are shown in Figure 7b. We compared coda amplitudes of regular earthquakes at MANS with those at the JTS, a permanent station of the Global Seismograph Network by Incorporated Research Institutions for Seismology and International Deployment of Accelerometers (Scripps Institution of Oceanography, 1986). The average ratio of coda amplitudes at MANS to those at JTS is 1.14, suggesting that the condition of MANS site is very similar to that of the JTS.

3.1.3. Seismic energy rate of tremors

By using apparent attenuation ($Q_{app.}^{-1}$) and site amplification in the previous subsections, we estimated the energy rate functions of tremors. The source energy rate function of a tremor

($E_j\ (t)$) using the amplitude of the j-th station is calculated by the following formula (Maeda & Obara, 2009):

$$E_j(t) = 2\pi V_S r_j^2 \rho A'^2_j(\mathrm{t}+t_j)\exp\left(2\pi f_c Q_{app.}^{-1} t_j\right), \quad (3)$$

where $A'_j\ (t)$ is the site-corrected amplitude of the envelope waveform of the j-th station, r_j is the hypocentral distance from the accompanying VLFE, t_j is the travel time from the VLFE source, and ρ is the density (assuming 2,700 kg/m^3). For calculating $E_j\ (t)$, we used a 180 s time window that started 60 s before the origin time of VLFEs. We calculated the CCs of all station pairs in Figure 7b. To estimate the source energy rate function of the tremor, we only used stations whose CCs with at least one other station exceeded 0.6.

The seismic energy rate W_j using the amplitude of the j-th station is given by the integration of the source energy rate function $E_j\ (t)$ in time:

$$W_j = \frac{1}{t_2 - t_1}\int_{t_1}^{t_2} E_j(t)dt, \quad (4)$$

where t_1 and t_2 are the start and end of the integration range, respectively. The integration range is defined as the period in which the values of $E_j\ (t)$ exceeded 20% of the maximum value of $E_j\ (t)$ (Figure 8). The seismic energy rate of a tremor (W_0) was obtained by calculating the average W_j of all stations. The error of W_0 was obtained by calculating the standard deviation of W_j.

3.2. Results

The energy rates of tremors were mainly distributed in 10^3–$10^{5.5}$ J/s (Figure 9). There is a positive correlation between the energy rates of tremors and the moment rates of the corresponding VLFEs. We estimated the scaled energy by calculating the ratio between the seismic energy rate of a tremor and the seismic moment rate of the corresponding VLFE. The scaled energy of slow earthquakes off Costa Rica is mainly distributed in the range of 10^{-9}–10^{-8} (dotted lines in Figure 9).

4. Discussion

4.1. shallow ETS off Costa Rica

The activation of VLFEs and tremors in August 2005 temporally correlates with the 2005 SSE reported by Jiang et al. (2012). VLFEs and tremors occurred mainly in the updip area in August 2005; hence, the slip area of the 2005 SSE can be distributed in the updip area near the trench axis, similar to the 2007 SSE. In areas where shallow VLFEs occurred, subseafloor hydrological observatories recorded pore fluid pressure transients in 2000 (Brown et al., 2005), 2003–2004 (Solomon et al., 2009), and 2007–2013 (Davis et al., 2011; 2015). They interpreted that pore fluid pressure transients were caused by SSEs. Spatial correspondence of pore fluid change in the periods of previous studies and VLFE activity in 2005 near the trench off Costa Rica suggests the occurrence of a shallow ETS, as with the Nankai subduction zone (Araki et al., 2017; Nakano et al., 2018).

4.2. Separation of slow earthquakes and other phenomena

Before the 2012 M_W 7.6 earthquake, the interplate coupling of the shallow slow earthquake area at a plate-boundary depth range of 6–10 km was expected to be very weak (Feng et al., 2012) unlike the coseismic slip area, which was strongly coupled (Protti et al., 2014). The average stress drop of small-to-moderate regular earthquakes inside the large slip area of the 1992 tsunami earthquake (surrounded by dark blue lines in Figure 4) was 1.2 MPa, which was smaller than that outside the large slip area (Bilek et al., 2016). The values of reported stress drops of slow earthquakes in the Nankai subduction zone were 0.1–200 kPa (e.g., Ito & Obara, 2006; Takagi et al., 2019); therefore, we consider that the stress drops of slow earthquake area are also much smaller than those of regular and tsunami earthquakes. The spatial variation of interplate coupling and stress drop of slip at the plate boundary results from the heterogeneous distribution of frictional properties at the plate boundary in the Central American subduction zone. In addition, a low stress drop suggests a high pore pressure generated by the existence of fluids (Yao & Yang, 2020). Therefore, the frictional strength of the slow earthquake area at a depth range of 6–10 km can be quite weak owing to the rich fluid compared to that in the regions with regular and tsunami earthquakes.

In Costa Rica, repeating earthquakes were activated after the 2012 M_W 7.6 earthquake around the large coseismic slip area of the earthquake (Chaves et al., 2020). Such activation after a large earthquake in the afterslip area was also observed in the Tohoku subduction zone (Uchida & Matsuzawa, 2013). The locations of repeating earthquakes separate from the areas where VLFEs occur. Such separation is also found in the Nankai (e.g., Takemura et al., 2020) and the Tohoku subduction zone (e.g., Nishikawa et al., 2019).

4.3. Comparison with other subduction zones

Our study revealed that shallow VLFEs and tremors occur near the trench axis off Costa Rica, in the updip of coseismic slip areas of thrust-type large earthquakes with an M_W of 7–8. In the updip area, SSEs also occurred in 2007–2012 (Dixon et al., 2017; Jiang et al., 2012). The depth range and the separate distribution between shallow slow earthquakes and large earthquakes off Costa Rica are similar to shallow slow earthquakes in the Nankai subduction zone, where slow earthquakes are spatially separated from high slip-deficit zones (e.g., Takemura et al., 2020). On the other hand, before the 2011 Tohoku earthquake, shallow slow slip events propagated to the initial rupture point of the great earthquake (Kato et al., 2012). Therefore, the characteristics of distribution of slow and large earthquakes differ between Tohoku and Costa Rica.

There are other common features in shallow slow earthquakes between Costa Rica and Nankai. Although the lower limit of M_W is large (~3.4) due to a strict threshold, the ranges of magnitudes and source durations of shallow VLFEs off Costa Rica are similar to those of shallow VLFEs in the Nankai subduction zones (e.g., Takemura et al., 2019). The recurrence intervals of activation of slow earthquakes are one to several years in Costa Rica (Jiang et al., 2012), which is similar to shallow slow earthquakes in the Nankai subduction zone, but different from the shorter intervals of deep slow earthquakes in Nankai (e.g., Baba et al., 2020b). Although the number of tremors whose energy rates are less than 10^4 J/s is small because of the strict detection threshold of the corresponding VLFEs, the upper limit of the energy rate range of tremors is similar to that observed for shallow tremors in Nankai (Yabe et al., 2019). The estimated scaled energy of slow earthquakes off Costa Rica is also similar to that of shallow slow earthquakes in the Nankai subduction zone (Yabe et al., 2019). The scaled energy is related to the rupture process of seismic phenomena (Kanamori & Rivera, 2006), therefore these results suggest that the frictional properties within the shallow slow earthquake areas are similar in both Costa Rica and Nankai. On the other hand, the scaled energy range in both regions is 0.5–1 orders of magnitude larger than that of shallow slow earthquakes in the Tohoku subduction zone (Yabe et al., 2021), and approximately 1 order of magnitude larger than that of deep slow earthquakes in Nankai (Ide et al., 2008; Ide &Yabe, 2014; Ide, 2016; Ide & Maury, 2018; Maeda & Obara, 2009). We note that scaled energies of shallow slow earthquakes were estimated for individual events, whereas those of deep slow earthquakes estimated by Ide &Yabe (2014), Ide (2016), and Ide & Maury (2018) were estimated for stacked events.

The range of scaled energy and distribution of shallow slow earthquakes off Costa Rica are more similar to those in shallow Nankai than shallow Tohoku. According to Syracuse et al. (2010), the age and thermal parameters of Costa Rica are 15.8 Ma and 1,010 km, respectively, which are closer to those of Nankai (20.0 Ma and 450 km, respectively) than Tohoku (115.2–130.5 Ma and 5,720–6,040 km, respectively). The thermal parameter, which is product of the incoming plate age, the convergence rate, and the sine of the slab dip angle, is used to predict the slab surface temperature at a given depth (e.g., Kirby et al., 1991; Syracuse et al., 2010). In addition, the temperatures of shallower parts of plate interfaces of these subduction zones where shallow slow earthquakes are not so different (Nankai: ~100 ºC in the depth range of 0–5 km from the seafloor; Tohoku: 65–110 ºC in the depth range of 6–12 km; and Costa Rica: 12–60ºC in the depth range of 0–10 km; Saffer & Wallace, 2015). On the other hand, the Central American subduction zone is subduction of fast convergence rate (~8 cm/year; DeMets et al., 1994), high dip angle, and erosional type (e.g., Bangs et al., 2016), which are more similar to Tohoku than Nankai. Although the characteristics of slow earthquake activity can be related to various factors, the thermal parameter and incoming plate age of Costa Rica is more similar to Nankai than Tohoku. The temperature structure of the shallow plate interface is probably most sensitive to incoming plate age (Maunder et al., 2019) and secondarily to thermal parameter (Syracuse et al., 2010). Hence, similar temperature conditions on the interface may explain the common features of shallow slow earthquakes off Costa Rica and in Nankai.

In previous studies, the large slip area of the SSE in 2007 was separated into deeper and shallower parts (Jiang et al., 2017), and deep LFEs and tremors were detected downdip of the seismogenic zone (Brown et al., 2009; Outerbridge et al., 2010). And, several VLFEs were located in the downdip area, in the similar area reported in previous studies of tremors and LFEs. If these deep VLFEs, LFEs and tremors occur in the downdip area, slow earthquakes might occur at separate depths along both shallower and deeper extensions of rupture zones of large earthquakes (Figure 10). This characteristic might also be the same as that of the Nankai subduction zone (Obara & Kato, 2016). This suggests that the tectonic property may be similar in the wide depth range in Costa Rica and Nankai. On the other hand, slow earthquakes are distributed only in the deeper part in the Cascadia subduction zone and only in the shallower part in the Tohoku subduction zone. The variation of the distribution of slow earthquakes may be attributed to the difference in tectonics or detection capability. The elucidation of the reason for the difference of the distribution in slow earthquake is future works.

5. Conclusions

Based on the grid-search matched-filter technique using synthetic templates in the regional 3D model, we detected and located VLFEs around the Nicoya Peninsula. Many VLFEs occurred in September 2004 and August 2005, and more than 90% of the VLFEs were located near the trench axis, where the plate boundary is at a depth range of 6–10 km, updip of the seismogenic zone, whereas several VLFEs were located in the downdip area at a depth of ~40 km. In this area, the occurrence of shallow SSEs is suggested by VLFE episodes. The region with VLFE activity overlaps with the shallower part of the large slip area of the 2007 SSE; therefore, the occurrences of shallow SSEs are suggested in September 2004 and August 2005 to occur in the same area as the shallower part of the 2007 SSE. The distribution of VLFEs lies in the gap surrounding coseismic slip areas of tsunami and large regular earthquakes. This separation reflects the spatial distribution of the frictional strength of the plate boundary in the Central American subduction zone. By using high-frequency seismogram envelopes, we also estimated the energy rates of tremors accompanying VLFEs. The ranges of magnitude and source duration of VLFEs, energy rate of tremors, and scaled energy off Costa Rica are similar to those in shallow slow earthquakes in the Nankai subduction zone.

Data Availability

We used seismograms of the TUCAN network (Abers & Fischer, 2003; https://doi.org/10.7914/SN/YO_2003) and Global Seismograph Network (Scripps Institution of Oceanography, 1986; https://doi.org/10.7914/SN/II). We used the earthquake catalog of the U.S. Geological Survey (https://earthquake.usgs.gov/earthquakes/search/). We used OpenSWPC code Version 5.0.2 (Maeda et al., 2017; https://doi.org/10.5281/zenodo.3712650) for the numerical simulations. Numerical simulations were conducted using the Fujitsu PRIMERGY CX600M1/CX1640M1 (Oakforest-PACS) at the Information Technology Center, the University of Tokyo. We used generic mapping tools (Wessel et al., 2013) and Seismic Analysis Code (Helfrich et al., 2013) to prepare the figures and process seismograms, respectively. The VLFE and tremor catalog constructed by this study is provided in an open access repository, zenodo (doi: 10.5281/zenodo.4435232).

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Figures

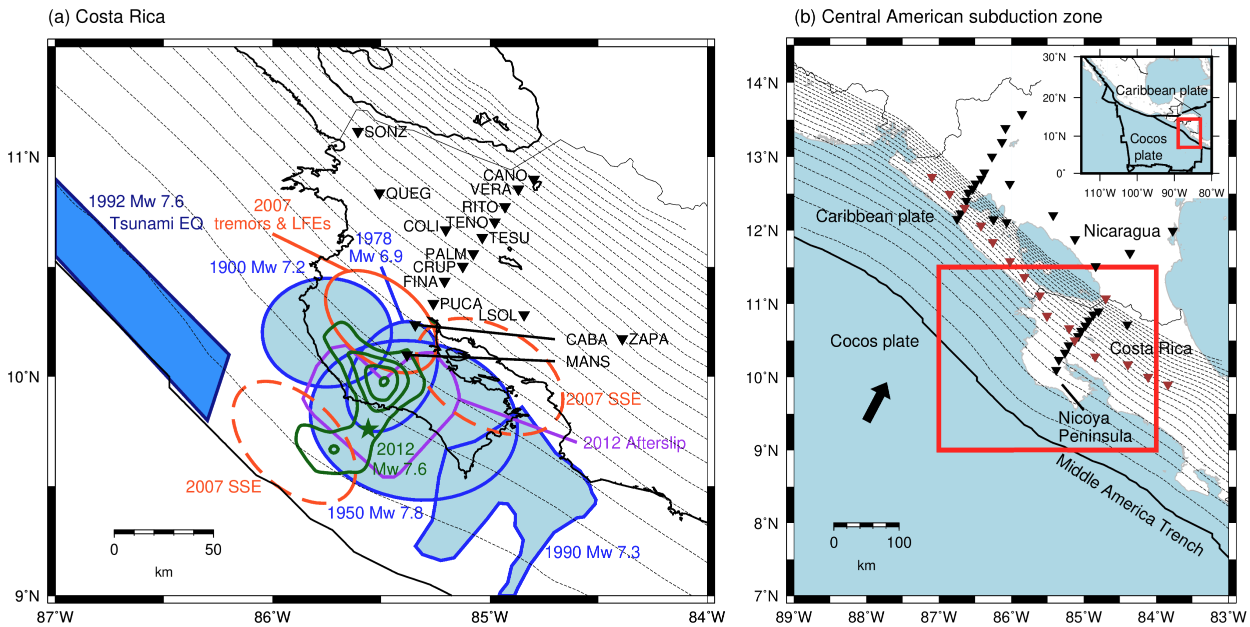

Figure 1. (a) Large regular and slow earthquake areas based on previous studies around the Nicoya Peninsula, in Costa Rica. Green contours show the coseismic slip distribution of the 2012 M_W 7.6 earthquake with a 1-m interval (Yue et al., 2013). Blue and dark blue areas show the slip areas of large and tsunami earthquakes (1990 M_W 7.3: Protti et al., 1995; others: Yue et al., 2013). Orange ellipses with dashed lines show large slip areas of the 2007 SSE (Jiang et al., 2017). The orange ellipse with a solid line shows the distributions of LFEs (Brown et al., 2009) and tremors (Outerbridge et al., 2010). The purple polygon shows the area whose afterslip of the 2012 M_W 7.6 earthquake is more than 150 mm (Malvervisi et al., 2015). Black inverted triangles show the station locations of the TUCAN network used in VLFE detection (Section 2.2). Dashed contours indicate the isodepths of the top of the Cocos Plate with 10 km intervals (Slab2; Hayes et al., 2018). (b) Map of the Central American subduction zone. Solid line represents the Middle America Trench (Slab2; Hayes et al., 2018). Dashed contours are the same as (a). Black arrow indicates the convergence direction of the Cocos Plate, which subducts below the Caribbean plate from the Middle America Trench (NUVEL-1A; DeMets et al., 1994). Inverted triangles show the locations of stations of the TUCAN network. Brown triangles are stations which were used in beamforming (Section 2.3). The black lines in the inset show plate boundaries.

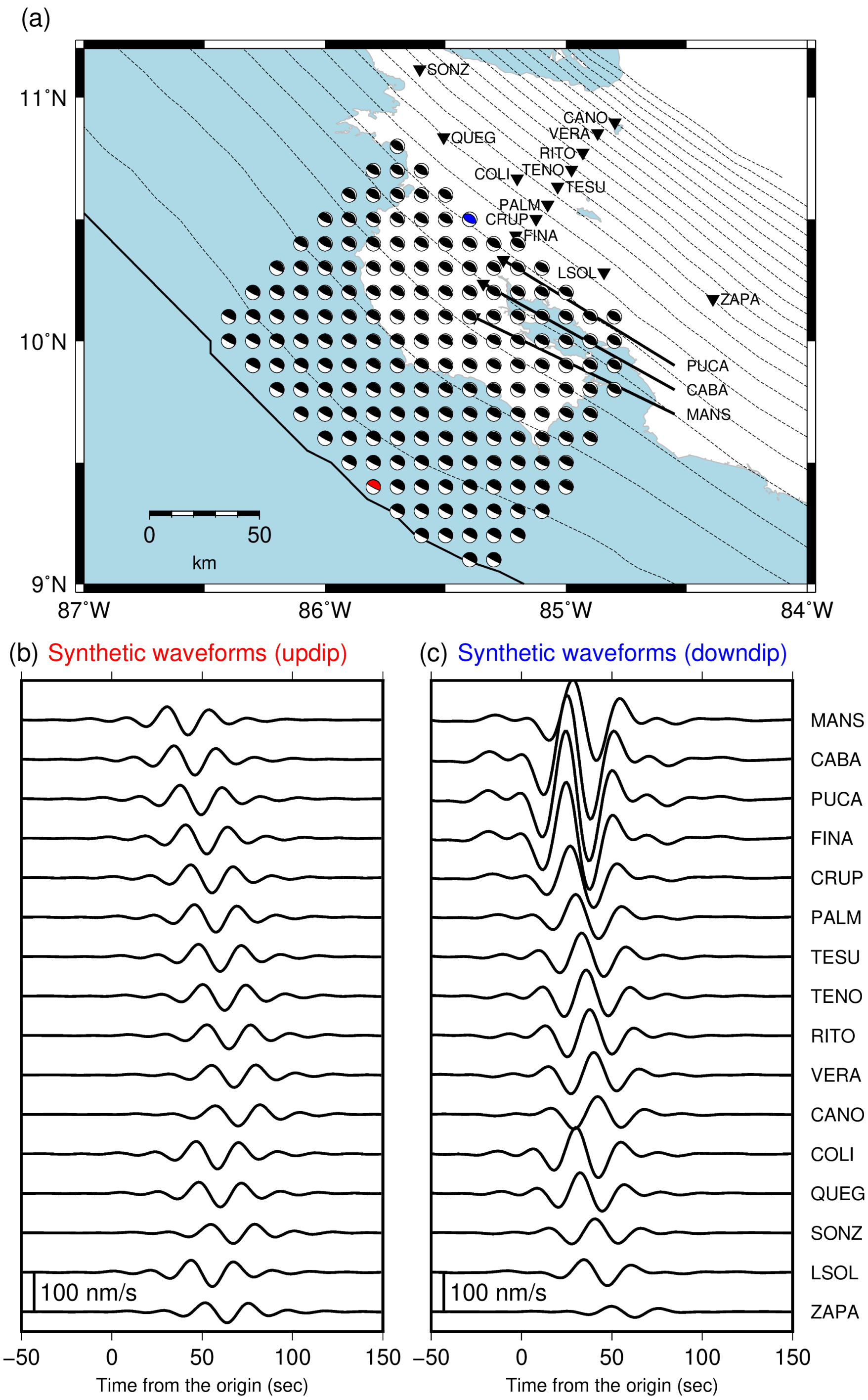

Figure 2. (a) Virtual source grids assumed in this study. Beach balls show the locations and focal mechanisms of the virtual sources. Inverted triangles and the black line are the same as in Figure 1. Dashed contours indicate the isodepths of the top of the Cocos Plate with 10 km intervals (Slab2; Hayes et al., 2018). Examples of waveforms of virtual sources in the (b) updip and (c) downdip areas. Sources of Figures 1b and 1c are shown by the red and blue beachballs in Figure 2a, respectively.

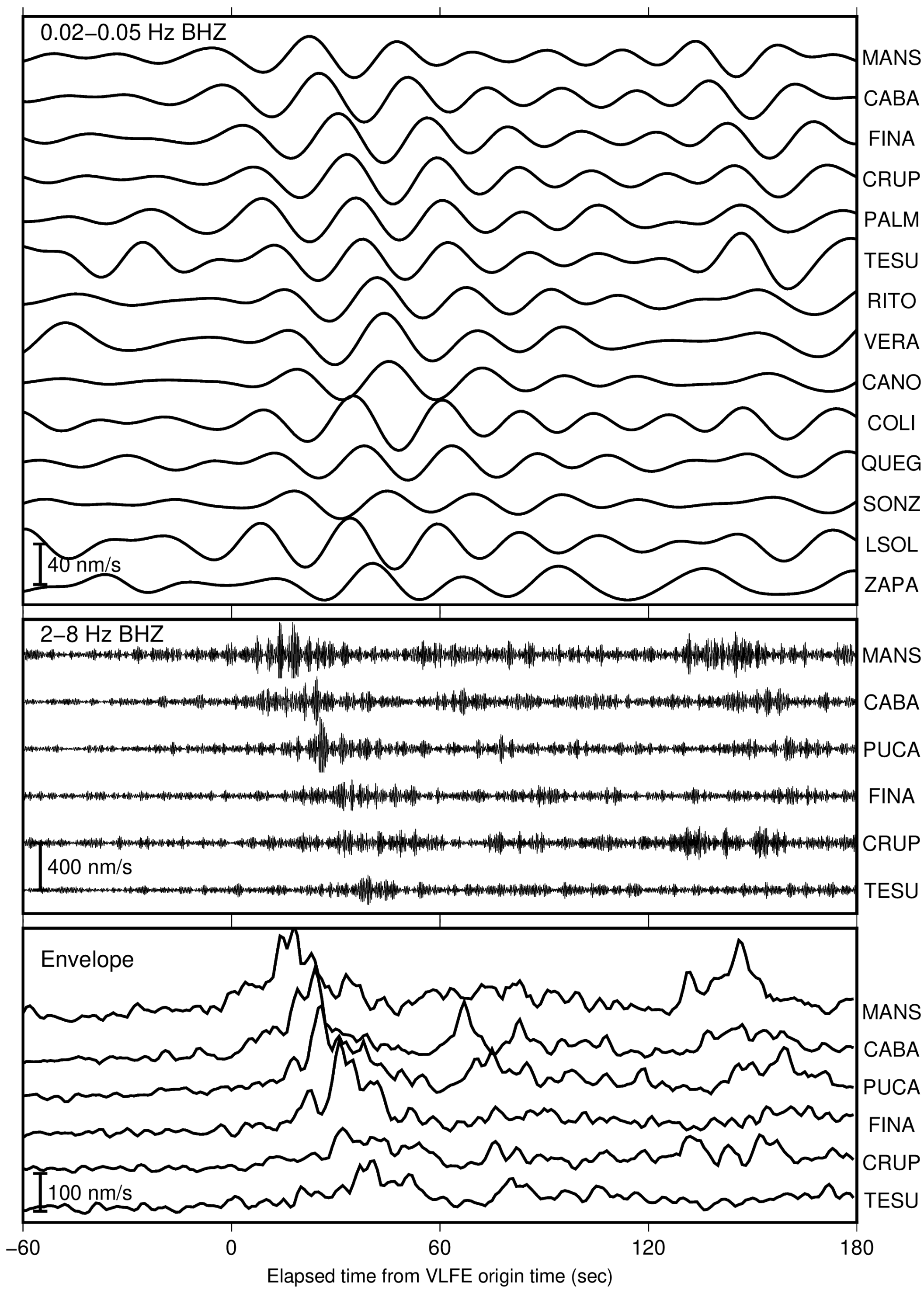

Figure 3. Example of waveforms of a VLFE and the corresponding tremor located at 85.8°W and 9.4°N (shown by a red beachball in Figure 2a) in the frequency range of 0.02–0.05 Hz and 2–8 Hz, and smoothed root-mean-square envelope in the frequency range of 2–8 Hz. Seismograms are shown from the origin time of the VLFE, 03:53:47 (UTC), August 10, 2005.

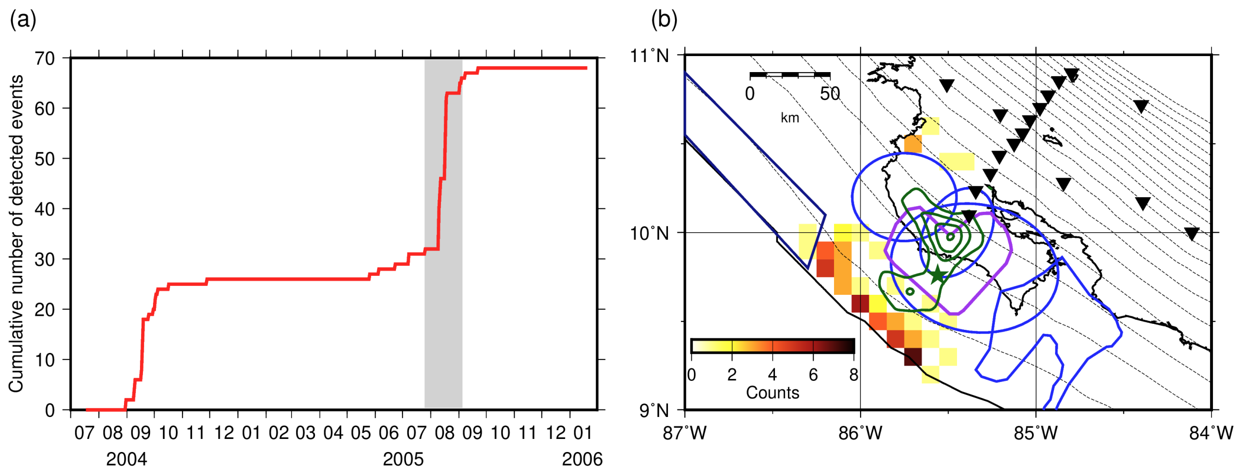

Figure 4. (a) Cumulative number of the VLFEs from July 2004 to January 2006. Gray shading shows the period of the 2005 SSE (Jiang et al., 2012). (b) Distribution of the number of detected events at each virtual source. Blue ellipses and polygons, dark blue quadrangle, inverted triangles, black line, the purple polygon, and dashed contours are the same as in Figure 1.

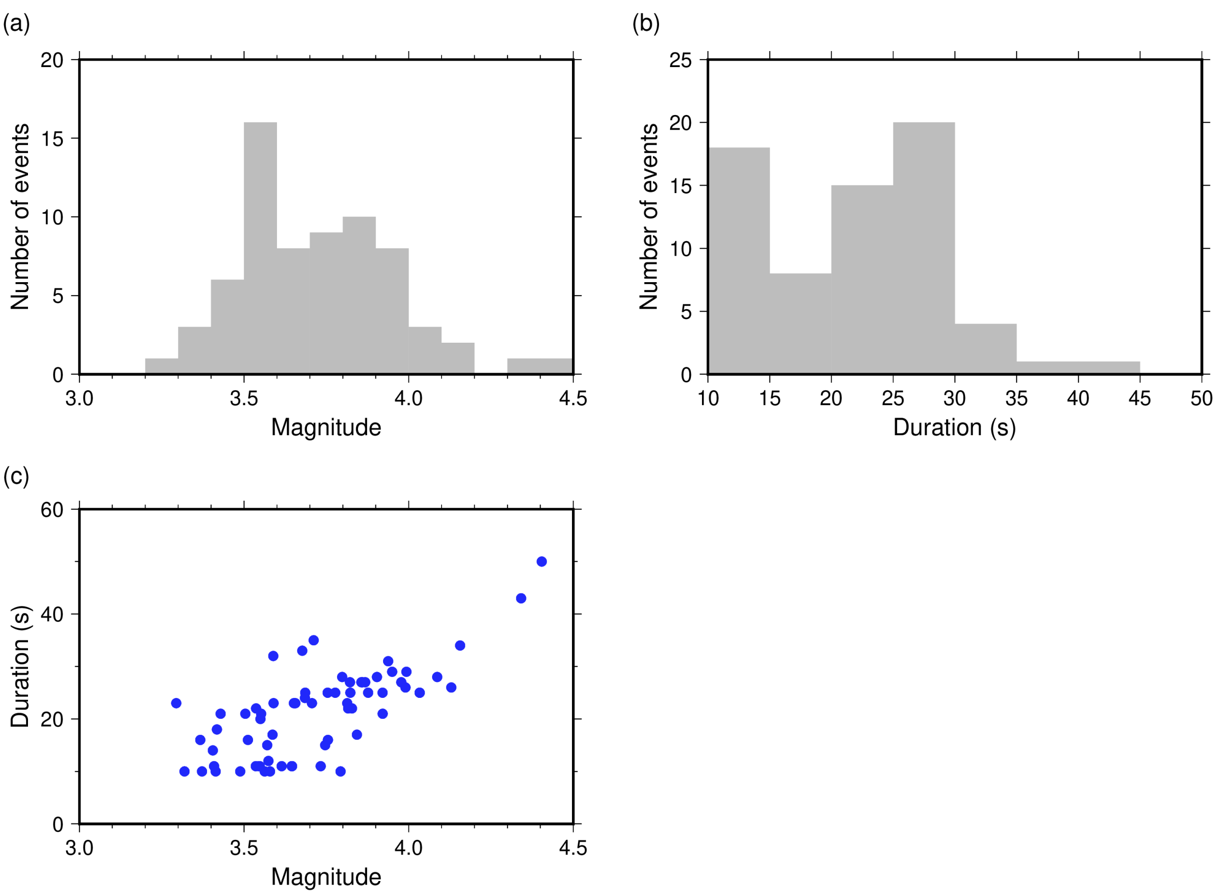

Figure 5. Distribution of (a) magnitudes and (b) source durations of VLFEs. (c) Relationship between source durations and magnitudes of VLFEs.

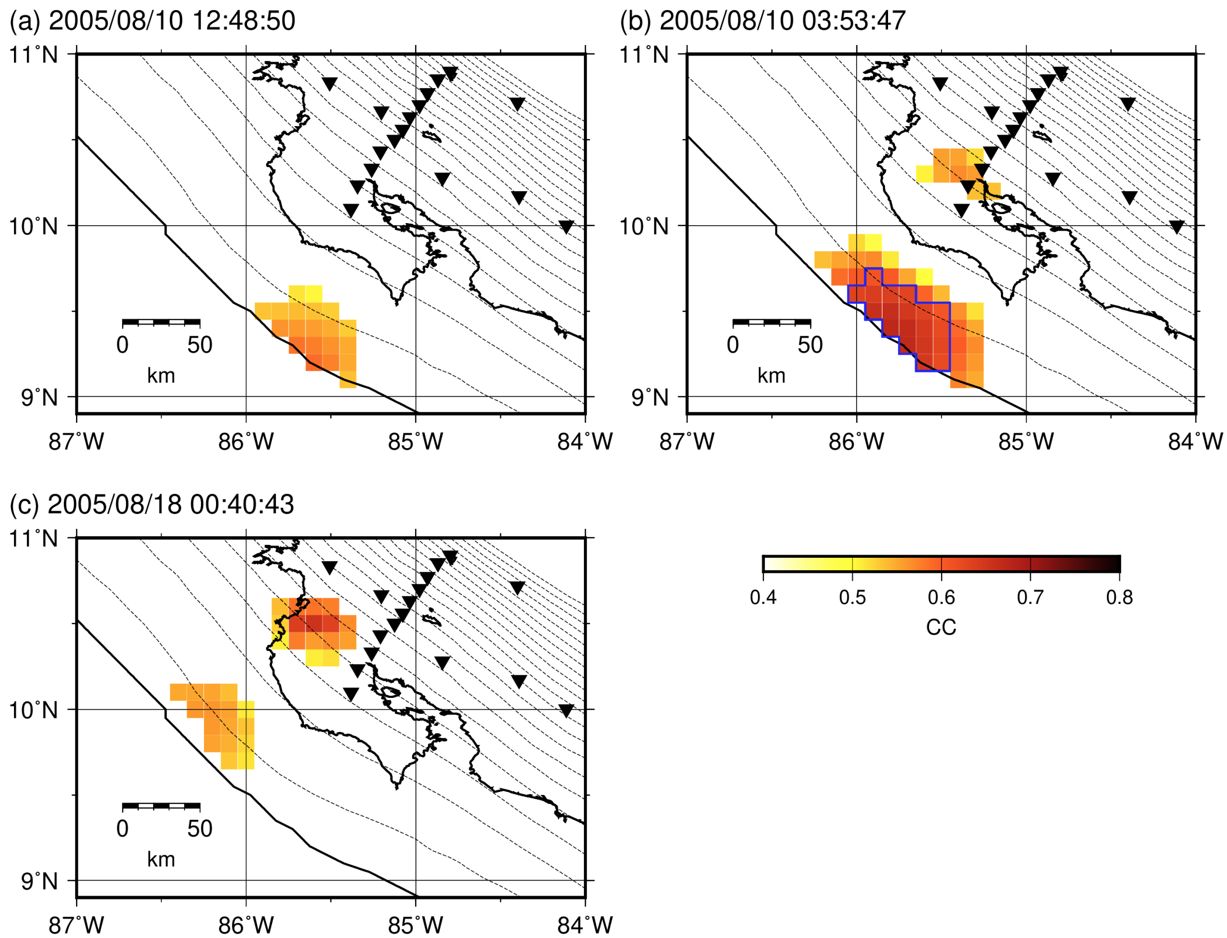

Figure 6. Examples of CC distributions of (a) an event which has large CCs only in updip grids, (b) an event which has large CCs both in updip and downdip grids but is located in an updip grid, and (c) an event which has large CCs both in updip and downdip grids but is located in a downdip grid. Inverted triangles, black line, and dashed contours are the same as in Figure 1. The blue polygon in (b) indicates the grids whose CC are more than 0.9 times of the maximum CC.

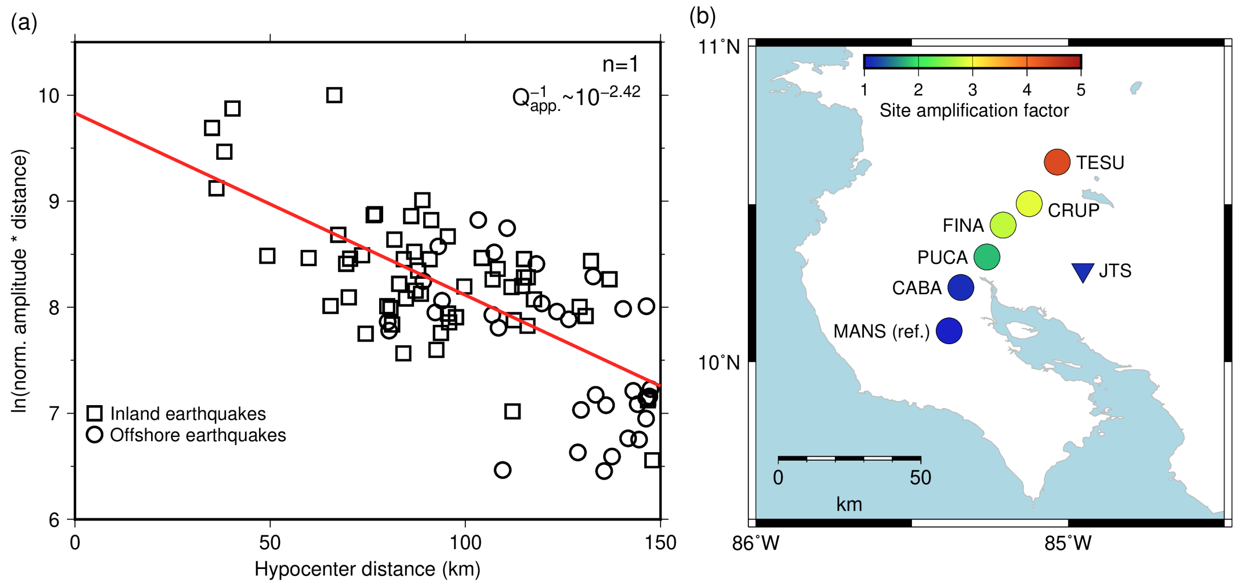

Figure 7. (a) Relationship between logarithm of coda-normalized maximum *S*-wave amplitudes and hypocentral distances. To eliminate effects of geometrical spreading of *S*-wave, coda-nomadized *S*-wave amplitudes were multiplied by their hypocentral distance. Red line shows the regression line using Equation (2). (b) Site amplification factors relative to MANS based on relative coda amplitude measurements.

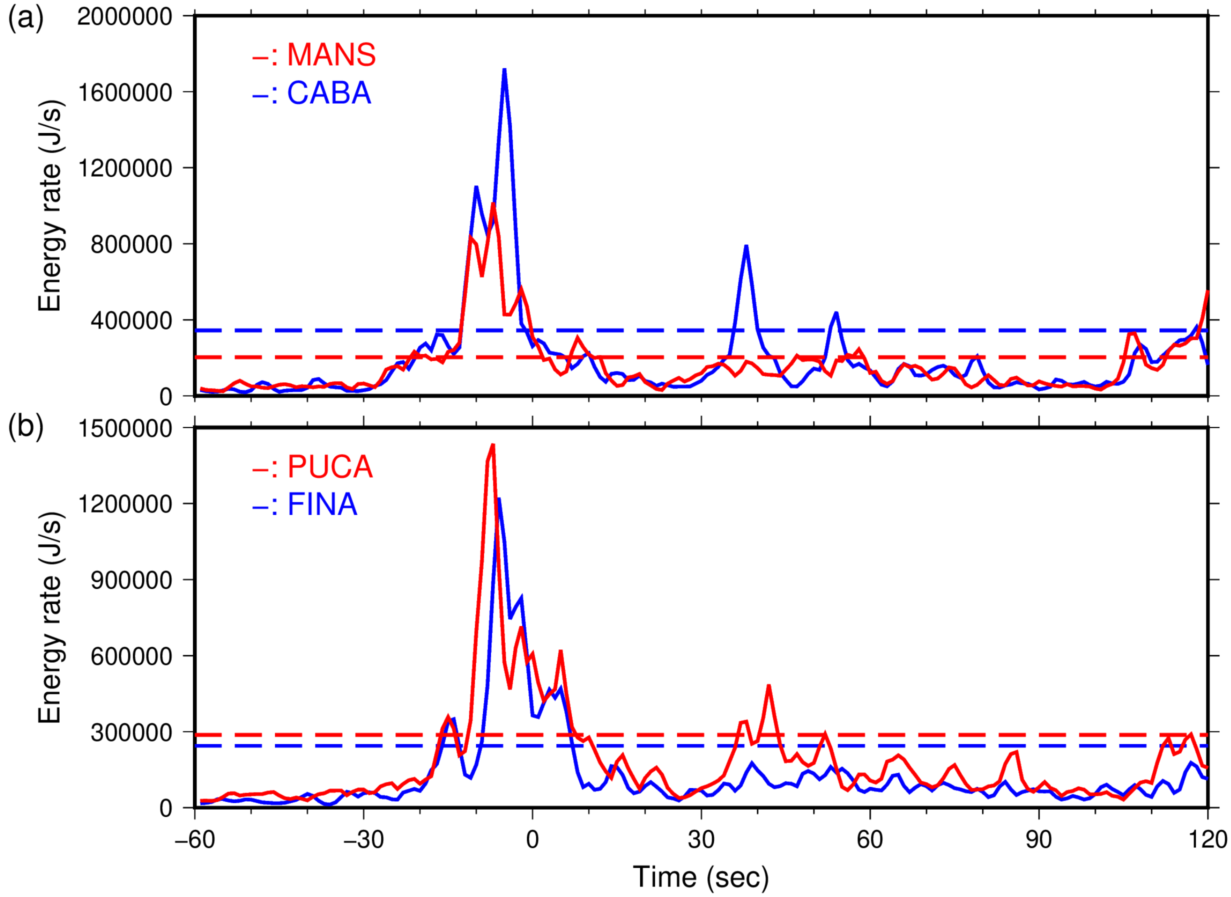

Figure 8. Temporal changes of energy rate functions of a tremor in (a) MANS and CABA and (b) PUCA and FINA. The corresponding VLFE occurs on 03:53:47 (UTC), August 10, 2012. Dashed lines indicate the threshold, which is set as 20% of the maximum value of the energy rate functions.

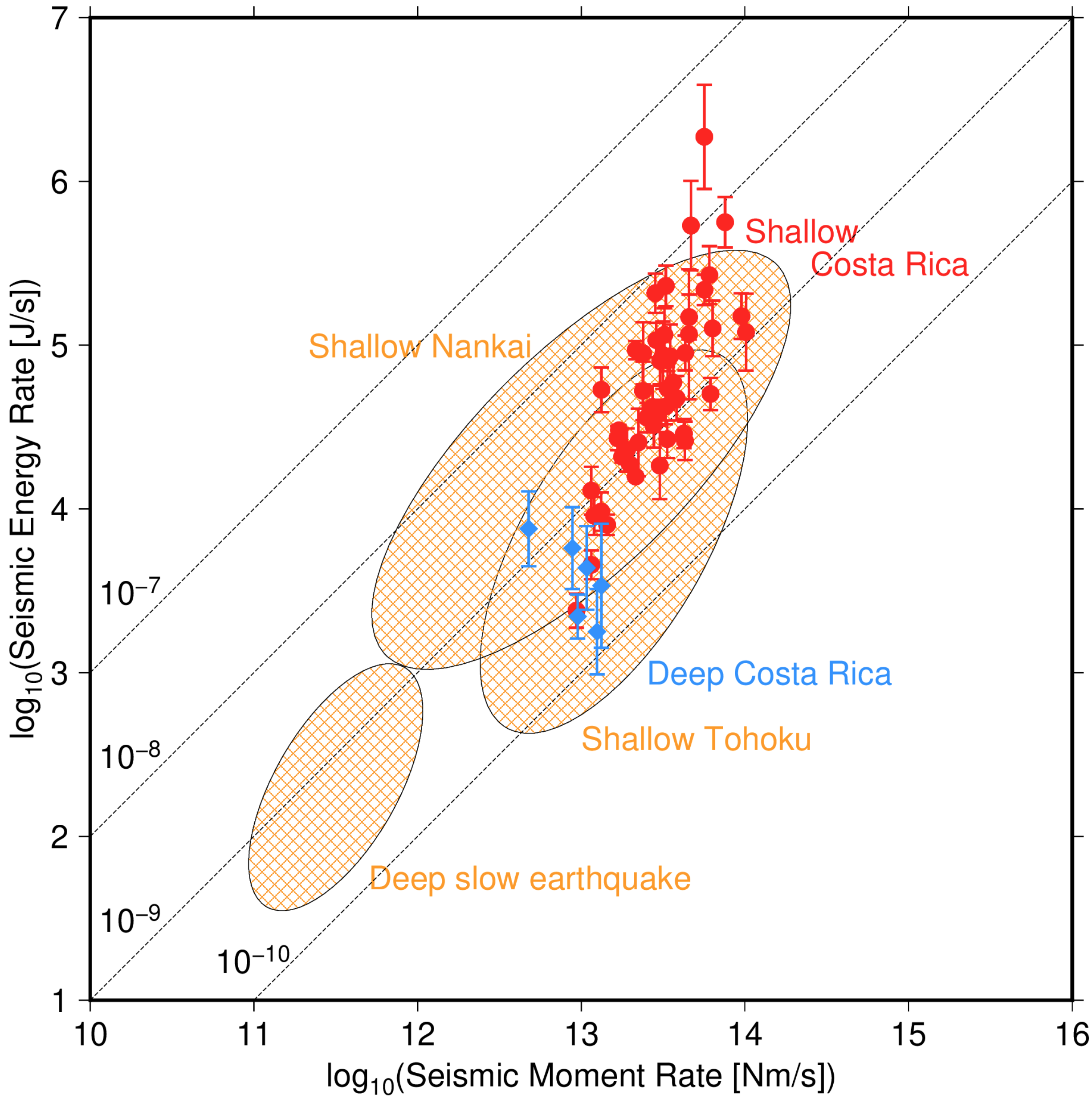

Figure 9. Relationship between seismic moment rates of VLFEs and seismic moment rates of tremors estimated in this study. Red circles and blue diamonds show the events of updip and downdip regions, respectively. Dashed lines show scaled energies of 10^{-7}, 10^{-8}, 10^{-9}, and 10^{-10}. Orange shadings show the relationships between seismic moment rates of VLFEs and seismic moment rates of tremors of shallow slow earthquakes in the Nankai (Yabe et al., 2019) and Tohoku subduction zones (Yabe et al., 2021), and deep slow earthquakes in southwest Japan (Ide & Yabe, 2014), Cascadia (Ide, 2016), and Mexico (Ide & Maury, 2018). We note that scaled energies of shallow slow earthquakes were estimated for individual events, whereas those of deep slow earthquakes were estimated for stacked events.

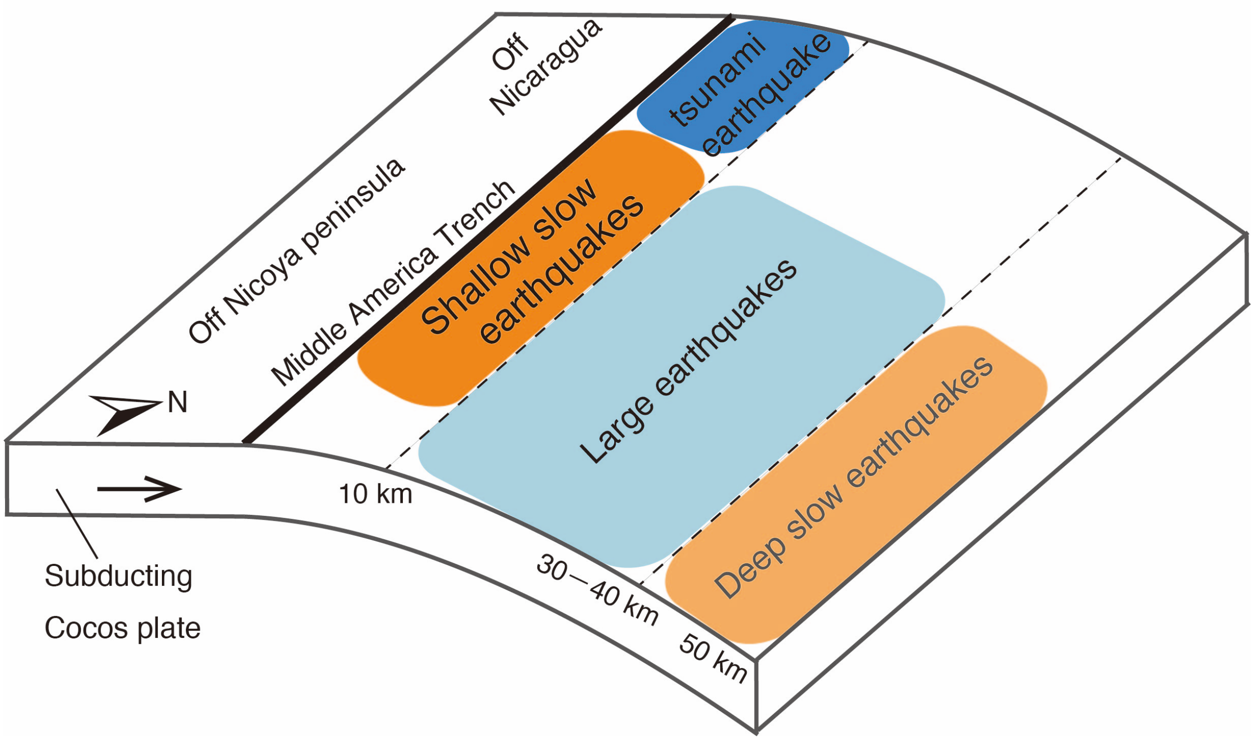

Figure 10. A schematic illustration showing the interpretation of distributions of slow, tsunami, and large regular earthquakes in the Central American subduction zone. The areas of large earthquakes, the 1992 tsunami earthquake, and deep slow earthquakes are referred from Yue et al. (2013), Satake (1994), and Outerbridge et al. (2010), respectively.