Cascading rupture of patches of high seismic energy release controls the growth process of episodic tremor and slip events

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Abstract

Slip phenomena on plate interfaces reflect the heterogeneous physical properties of the slip plane and thus exhibit a wide variety of slip velocities and rupture propagation behaviors. Recent findings on slow earthquakes reveal similarities and differences between slow and regular earthquakes. Episodic tremor and slip (ETS) events, a type of slow earthquake widely observed in subduction zones, likewise show diverse activity. We investigated the growth of 17 ETS events beneath the Kii Peninsula in the Nankai subduction zone, Japan. Analyses of waveform data recorded by a seismic array enabled us to locate tremor hypocenters and estimate the migration patterns and spatial distribution of the energy release of tremor events. Here we describe three major features in the growth of ETS events. First, independent of their start point and migration pattern, ETS events exhibit patches of high seismic energy release on the up-dip part of the ETS zone, suggesting that the location of these patches is controlled by inherent physical or frictional properties of the plate interface. Second, ETS events usually start outside the high-energy patches, and their final extent depends on whether the patches participate in the rupture. Third, we recognize no size dependence in the initiation phase of ETS events of different sizes with comparable start points. These features demonstrate that the cascading rupture of high-energy patches governs the
growth of ETS events, just as the cascading rupture of asperities govern the growth of
regular earthquakes.

**Key words:** Slow earthquakes, Tectonic tremor, Subduction dynamics, Rapid tremor
reversal

**Introduction**

Earthquakes feature heterogeneous slip distributions on fault planes that are
caracterized by one to several areas of large slip, or asperities. In general, earthquake
hypocenters are located outside of asperities. The complex growth and rupture processes
of earthquakes are fundamental features that have fueled a long-running debate on how
and when the eventual size of an earthquake is determined, specifically the existence of
a nucleation phase (e.g., Ellsworth and Beroza 1995; Ide 2019).

Recent findings about slow earthquakes have broadened our understanding of
subduction dynamics and the generation of megathrust earthquakes (e.g., Obara and
Kato 2016). In subduction zones, strain on the plate interface is released by slow slips in
slow earthquakes and high-speed ruptures in regular earthquakes. Interestingly, the
spatial distributions of slow and regular earthquakes are complementary: Slow
earthquakes are distributed around and outside the locked zones that are the source areas of megathrust earthquakes, both along dip (Obara and Kato 2016) and along strike (Nishikawa et al. 2019). There also appears to be a temporal relationship between slow and regular earthquakes. Kato et al. (2012) reported the occurrence of two slow slip events (SSEs) from earthquake activities which migrated toward the future hypocenter of the 2011 Tohoku earthquake before the mainshock. The migration speed of those hypocenters, 2–5 km/day, is close to the typical migration speed of episodic tremor and slip (ETS) events, 10 km/day (Dragert et al. 2001; Obara 2002). Slow slips preceding a dynamic rupture have also been reported in laboratory experiments and numerical simulations (e.g., Dieterich 1979; Ohnaka 1992; Shibazaki and Matsuzawa 1992).

Matsuzawa et al. (2010) showed that in a numerical simulation, the recurrence interval of SSEs grew shorter in the time interval preceding a megathrust earthquake. Therefore, investigating the features of slow earthquakes yields clues to the generation of megathrust earthquakes in subduction zones.

In some subduction zones, slow earthquakes occur as short-term SSEs in the form of ETS events, in which tectonic (nonvolcanic) tremor is synchronized with slip events in time and space (Rogers and Dragert 2003), such that tremor can be used to monitor ETS events. The growth process of ETS events is not simple and is typified by diverse
patterns of migration and eventual sizes (Obara et al. 2010), heterogeneous slip distributions (Hirose and Obara 2010), and brief episodes in which tremor migrates rapidly in the direction opposite to its overall migration (Houston et al. 2011). Ghosh et al. (2012) reported the existence of “tremor asperities,” areas of high radiated seismic energy, on the plate interface analogous to the asperities of regular earthquakes. In the Nankai subduction zone off southwestern Japan, the ETS zone has been divided into segments (Obara 2002) in which ETS events have similar recurrence intervals of several months. The short recurrence intervals of these ETS events make the Nankai subduction zone a promising area to elucidate what controls their growth.

Seismic arrays are a powerful tool for detecting and characterizing tremor (Ghosh et al. 2009; Imanishi et al. 2011). The Geological Survey of Japan, National Institute of Advanced Industrial Science and Technology (GSJ), installed a seismic array of 39 sensors on the Kii Peninsula from February 2011 to November 2016 (Fig. 1), producing a set of well-recorded waveforms that has been valuable for examining the detailed growth process of ETS events from tectonic tremor. Sagae et al. (2020) applied the MUSIC high-resolution frequency-wavenumber (f-k) method (Schmidt 1986) to a dataset from July 2012 to July 2014, revealing a more detailed view of tremor migration than the conventional envelope cross-correlation method. We report here the spatio-
temporal distribution of the radiated seismic energy of tectonic tremor during 17 ETS events between April 2011 and December 2014. We show that persistent patches of high seismic energy release are located on the plate interface in the ETS zone and that the growth process of ETS events is controlled by cascading ruptures of these patches.

Data

We analyzed waveform data recorded by the GSJ array network in the Kii Peninsula from April 2011 to December 2014 (Fig. 1). The array consisted of 39 three-component velocity seismographs with a natural frequency of 2 Hz, and the sampling frequency of waveform data was 200 Hz. For the semblance analysis, we used the N-S component of seismic data bandpass-filtered between 2 and 4 Hz, as the N-S component yielded the highest semblance value of the three components. For the calculation of the seismic energy of tremors, we used all three components of data between 2 and 8 Hz. We used the JMA2001 seismic velocity model (Ueno et al. 2002) for locating tremors.

Method

ETS event durations for semblance analysis
In general, the activity of an ETS event is correlated with the number of recorded tremors (e.g., Obara 2010). In this study, we treated a concentration of tremors in space and time as an ETS event. To estimate the duration of ETS events from tremor activity for our semblance analysis, we applied our clustering technique to the National Research Institute for Earth Science and Disaster Resilience (NIED) hybrid clustering catalog (Obara et al. 2010). We classified tremor events with hypocenters less than 20 km apart within time intervals shorter than 48 h as a group. Groups of more than 10 hypocenters were considered an ETS event. We estimated here the duration of an ETS event as the time from the start of the group’s first tremor to the end of its last tremor. Consequently, we detected 20 ETS events for our semblance analysis, of which 16 had previously been geodetically detected as short-term SSEs from analyses of strainmeter, tiltmeter, and water-level records (Kitagawa et al. 2011, 2012; Itaba et al. 2012, 2013a, 2013b, 2014a, 2014b, 2015; Ochi et al. 2015) (Figs. 1c and 1d). The semblance analysis included array data between two days before the start and one day after the end of each ETS event (Table 1).

Locating tremors from semblance analysis
Tremor has been shown to occur on the plate interface (e.g., Shelly et al. 2006; Ghosh et al. 2012; Suzuki et al. 2018). Thus, we assigned possible hypocenters to the grid points with horizontal intervals of 2 km on the plate interface, which is 5 km shallower than the Moho in the subducting Philippine Sea plate (Shiomi et al. 2008). We calculated the slowness of the optimum ray path between each grid point and the central station of the array network within the JMA2001 velocity model (Ueno et al. 2002).

Furthermore, we compensated for the arrival time shift due to the elevation difference between the stations. We used eight earthquakes with magnitude $\geq 1.0$ and the epicentral distance from the central station of the array network $\leq 10$ km from 2011 to 2017 (Fig. 2a). We calculated the arrival time differences between the central station and the other stations from the cross-correlation function of waveform data using the time window of 1 s around the arrival time of P-wave. The arrival time differences are correlated positively with the elevation differences for all the earthquakes. We assumed here that ray paths of P-wave were vertical directly beneath the array network, so that the arrival time differences depended only on the elevation differences. A linear regression line between the arrival time differences and the elevation differences was determined by the least-squares method with a constraint of passing through the origin. The slopes of the lines, as P-wave velocities, were highly variable because of a large
scatter of data. Therefore, we chose the data of the earthquake that showed the best
fitting. As a result, the estimated P-wave velocity was 3.0 km/s (Fig. 2b), providing the
S-wave velocity of 1.7 km/s with the assumption of $V_p/V_S = \sqrt{3}$. We applied this S-wave velocity above sea level for the semblance analysis. We confirmed that the
variation (±30%) in this S-wave velocity did not significantly modify the following
results and discussion.

We conducted a semblance analysis (Neidell and Taner 1971) of the continuous
records to detect and locate tremors. Compared to the conventional beamforming and f-k methods, semblance analyses can detect events with lower signal-to-noise ratios as the
method is based on waveform coherency rather than the signal power. From the
waveform data of the N-S component, band-pass filtered at 2–4 Hz, we calculated
semblance from the slowness values of the grid cells by

$\text{SEM}(S_{nx}, S_{ny}, T) = \frac{\sum_{i=T}^{T+W} \sum_{j=1}^{M} f_j(t_i-\tau_{nj})^2}{\sum_{i=T}^{T+W} \sum_{j=1}^{M} f_j(t_i)}$, \hspace{1cm} (1)

$\tau_{nj} = S_{nx} \cdot x_j + S_{ny} \cdot y_j - HC_{nj} - SC_j$, \hspace{1cm} (2)

where SEM is a semblance value, $W$ a time window of 1 min, $T$ the start time of a time
window, $M$ the number of stations for which the maximum cross-correlation value of
waveforms for time shifts of −0.3 to +0.3 s between the central station and the $j$-th
station is greater than 0.3, $f_j$ the waveform data at the $j$-th station, $\tau_{nj}$ the time
difference between the central station and the $j$-th station calculated from the horizontal slowness vector $(S_{nx}, S_{ny})$ of the $n$-th grid point, and $x_j$ and $y_j$ are the $x$ and $y$ coordinates of the $j$-th station, respectively. $HC_{nj}$ is the time correction for the elevation difference between the central station and the $j$-th station for the horizontal slowness vector of the $n$-th grid point, and $SC_j$ is the station correction that is estimated to match hypocenters from the catalog with those from the semblance analysis.

If the maximum semblance value was greater than 0.3 and the number of stations used for the analysis was greater than 25, the grid point with the maximum semblance value was designated the hypocenter of a tremor. By applying a clustering procedure to the estimated hypocenters, we removed isolated events that were not part of an ETS event. Our criteria for clustering two hypocenters were an epicentral distance less than 5 km and a time interval shorter than 12 h. We visually inspected the waveforms of detected tremors to exclude waveforms of regular earthquakes or impulsive noises.

**Estimation of the seismic energy of tremors**

We calculated seismic energy from waveforms of three components at the central station during 1 min time windows by following the formula (Maeda and Obara 2009)

$$E_R = \int_{t-T_0/2}^{t+T_0/2} \frac{2\pi V_0 r_i^2 \rho A_i^2 (t' + t_i)}{\exp[-2\pi f_c Q^{-1} t_i]} dt'.$$  

(3)
where $E_R$ is the seismic energy, $T_0$ the time window of 1 min, $V_0$ the average S-wave velocity of 3.5 km/s, $r_i$ the hypocentral distance, $\rho$ the average density of 2,700 kg/m$^3$, $A_i$ the sum of squared velocity amplitude of the three components, $t_i$ the travel time as calculated from the hypocenter and the average S-wave velocity, $f_c$ the center frequency of 5 Hz, and $Q$ the quality factor. We adopted $Q$ values, which depend on the hypocentral distance, reported by Yabe and Ide (2014) and corrected for the effect of the free surface.

Results and discussion

Detectability of tectonic tremor and redefinition of ETS events

In this study, we typically detected about three times as many tremor hypocenters as listed in the NIED hybrid catalog, making it possible to detect the detailed growth process of an ETS event. We located 52,776 tremor hypocenters during 20 ETS events from our semblance analysis of the array data. The spatio-temporal distribution of tremors is in overall agreement with the result of Sagae et al. (2020), which was based on the MUSIC high-resolution f-k method, although their result had a time resolution as fine as 10 s.
Figure 3 shows the space-time plot of tremor in ETS event number 17 (ETS 17), during 9–14 January 2014. The temporal variation in tremor activity we observed is consistent with that of the NIED catalog. The large number of hypocenters enabled us to recognize bilateral migration of the hypocenters along strike during ETS 17 (Fig. 3).

To examine the growth process of an ETS event, we redefined ETS events on the basis of hypocenter clusters. We defined clusters as sets of tremor events that occurred within 1 h and 10 km epicentral distance, and the cluster hypocenter was defined as the centroid of a cluster with more than 10 tremors. The start and end points of an ETS event were defined as the centroids of the first and last clusters, and these points in turn defined the duration of an ETS event (Table 2). Three ETS events (ETS 1, 8, and 11) contained too few clusters with centroids for accurate estimates, and these have been excluded from the following analyses and discussion.

Stationary patches of high seismic energy release

Figure 4 depicts the migration of tremor hypocenter clusters and the distribution of cumulative seismic energy of tremors for three ETS events with different growth patterns (the other 14 ETS events are shown in Additional file 1: Fig. S1). ETS 3 displayed unilateral migration from the southern end, ETS 20 displayed unilateral
migration from the northern end, and ETS 9 displayed bilateral migration starting from
the central part (upper panels in Fig. 4). All three events displayed patches with
relatively high cumulative seismic energy release (‘high-energy patches’ hereafter) that
appeared consistently in the same locations, regardless of the start point location and the
migration direction (lower panels in Fig. 4). Therefore, the location of these patches
appears to be controlled by physical or frictional properties of the fault plane that are
inherent to the plate interface.

Ghosh et al. (2009) reported that the cumulative moment of tremor in an ETS event
in Cascadia had a spatially heterogeneous distribution and was characterized by several
patches of high moment release. The distribution of these patches reported by Ghosh et
al. (2009) appeared to be complementary to the slip surfaces of regular earthquakes,
suggesting that inherent properties on the plate interface control the distribution of the
patches.

**Relationship between start point, rupture extent, and high-energy patches**

Figure 5 shows relationships between the location of high-energy patches, start points,
and along-strike rupture extent (defined by tremor cluster hypocenters) of the 17 ETS
events (see also Additional file 1: Fig. S2). It is notable that nearly all of the start points
are outside the high-energy patches. This feature is similar to the relationship between hypocenters and asperities of regular earthquakes (e.g., Mai et al. 2005) and suggests that the rupture of an ETS event initiates in a low-strength area. The implication is that the high-energy patches are areas with high frictional strength. It is also notable that start points tend to be on the deeper, down-dip side of the plate interface, whereas the high-energy patches are on the up-dip side, implying a depth-dependent variation in frictional strength on the plate interface (Obara et al. 2010). Moreover, the along-strike distribution of start points appears to be heterogeneous. Some start points are concentrated on the northeastern side of patch A. This concentration, together with the rupture of patches A and B, causes the dominantly southwestward migration of tremor in the analyzed area (Obara 2010).

Focusing on the rupture extent of ETS events, we found that ETS events tended to terminate in or around the high-energy patches. This suggests that patches have high frictional strength. In that case, the complete rupture of a high-energy patch would extend the growth of an ETS event. On the other hand, the termination of the rupture in or around a patch would prevent the further growth of an ETS event.

Relationship between rapid tremor reversal and high-energy patches
Figure 6 shows two examples of rapid tremor reversal (RTR) (Houston et al. 2011) observed in this study. During ETS 2, the rupture propagated from patch A to patch B and then propagated back to patch A. During ETS 20, the rupture propagated from patch B to patch C, then back to patch B. It appears in both cases that RTR occurred when a high-energy patch was reruptured. The migration speed during RTR reached approximately 250 km/day, which is of the same order as RTR velocities observed previously in the Kii Peninsula (Obara et al. 2012) and in Cascadia (Houston et al. 2011), whereas the migration speed during the first rupture of the patches was approximately 15 km/day, similar to other ETS events not showing RTR. Our observations show clearly that the patches release part of their strain energy in the first rupture and release another portion in the second rupture during RTRs. The implication is that the passage of the ETS front does not entirely control the strain release in high-energy patches. We suggest that the state of stress or of fluids, not only in the patches but also in surrounding areas, is an important factor in the degree of strain release in the patches. Instead, the high-energy patches govern a complex rupture process for each ETS event. Spatial overlap between tremor patches and RTR areas has also been reported in Cascadia (Thomas et al. 2013).
Cascading growth process of ETS events

Comparing the temporal evolution of ETS events with the same start points effectively removes one possible confounding variable in the growth process of ETS events. Figure 7 displays the temporal evolution of three ETS events (ETS 9, 12 and 17) that initiated only a few kilometers apart. These events started at the down-dip end, and the rupture propagated up dip at about 1 km/h, which is consistent with the migration speed of 1 km/h reported for this event by Sagae et al. (2020). ETS 12 ruptured patch B and terminated without rupturing patches A and C, whereas ETS 9 and ETS 17 ruptured patch B and then propagated bilaterally to patches A and C. In ETS 17, patch A ruptured before patch C. The rupture of ETS 17 continued longer than that of ETS 9, and the radiated seismic energy from patches A and C was larger in ETS 17 (Fig. 7b). These results show that differences in ETS growth are controlled by the rupture of the strong patches, even for ETS events with nearly identical start points.

For regular earthquakes, the existence of a nucleation phase, or slow initial phase, has been controversial, and some studies have reported that the eventual size of an earthquake is scaled as the cube of the duration of the nucleation phase (e.g., Ellsworth and Beroza 1995). We investigated whether such a phase can be found for ETS events, again using the trio of ETS 9, 12, and 17 (Fig. 7c). Given the possible uncertainty of the
start time of ETS events, we assigned a consistent start time for the three events by
aligning their energy release curves at the beginning of the rupture of patch B, then
considered the relation between energy release during the day before and the day after
this start time (Fig. 7c). Despite the large difference in their sizes (Fig. 7b), the three
events had very similar initial phases before the start time; that is, there was no size-
dependent growth process evident in the initial part of the ETS events. This result shows
that the growth of ETS events is controlled by how high-energy patches rupture and
suggests that ETS events have a cascading rupture process similar to that reported for
regular earthquakes (Ide 2019).

A logarithmic plot of the data shown in Fig. 7b (Fig. 8a) allows us to focus on the
long-term cumulative energy curve of ETS events after $10^4$ s, a time range that
emphasizes variations between events. The cumulative energy of slow earthquakes is
considered to be proportional to the seismic moment with a constant of approximately
$10^{-10}$ (Ide et al. 2008), and the cumulative energy curves in Fig. 8a appear roughly
consistent with a scaling relationship between seismic moment ($M_o$) and duration ($T$) of

$M_o \propto T$ for slow earthquakes (Ide et al. 2007), shown as the pink stripe in Fig. 8.

However, it is apparent that the slopes of the cumulative energy curves are not constant.
The slope is steep for the time interval $2\times 4 \times 10^4$ s to $5\times 9 \times 10^4$ s, when the cumulative
energy ($E$) scales approximately as the cube of $T$, or $E \propto T^3$. The slope becomes gentler after $5-9 \times 10^4$ s, when the cumulative energy scales approximately as the lapse time, $E \propto T$ (Fig. 8a). The first time interval corresponds to the rupture of patch B, and the second corresponds to the rupture propagation along strike from patch B to patch A or C. Therefore, we suggest that ETS events might grow as $M_O \propto T^3$ during rupture expansion in a high-energy patch and as $M_O \propto T$ during rupture propagation along strike within the ETS zone. If this is the case, the scaling relation of $M_O \propto T^3$ for SSEs in Cascadia reported by Michel et al. (2019) might indicate the rupture of a high-energy patch or asperity during an ETS event. Moreover, the transition we found in the scaling relation between $M_O$ and $T$ differs from the result of a numerical simulation of slow earthquakes with a Brownian walk model, which shows a transition from $M_O \propto T^2$ to $M_O \propto T$ (Ide 2008), although this difference might reflect fluctuations in the temporal evolution of ETS events. The cumulative energy curves for all 17 ETS events in this study are quite scattered (Fig. 8b), and they appear to have no common evolution process signified by a systematic transition of the scaling between $E$ and $T$. On the other hand, most of the events approximate the $M_O \propto T$ scaling relationship (Ide et al. 2007) when considered as whole events (Fig. 8b). This is consistent with the $M_O \propto T^{0.811}$
scaling relationship for 61 SSEs in Shikoku, the Nankai subduction zones (Hirose and Kimura, 2020).

Conclusions

We analyzed seismic array records of tectonic tremor beneath the Kii Peninsula in the Nankai subduction zone to trace the growth of ETS events in detail. Our examination of tremor in 17 ETS events revealed the existence of stationary patches on the plate interface that release high radiated seismic energy. The location of these high-energy patches is independent of the start point and migration pattern of ETS events, suggesting that the distribution of the patches reflects heterogeneities in frictional properties that are inherent to the plate interface. The start points generally lie outside the high-energy patches, similar to the spatial relationship between hypocenters and asperities of regular earthquakes. In some ETS events, the high-energy patches have a role in terminating the rupture; in others, the high-energy patches rupture twice in an RTR-like activity. We also compared the initial part of ETS events with different sizes but nearly identical start points, finding that the final size of ETS events appears to be controlled by the extent of the rupture of the high-energy patches; that is, the final size of an ETS event cannot be estimated from its initial growth phase. These features
indicate that ETS events have an essentially cascading rupture process similar to that of regular earthquakes. The growth of ETS events is therefore controlled by how they rupture high-energy patches.

Additional file 1.

**Fig. S1.** Plots showing (upper) the migration of tremors and (lower) the spatial distribution of cumulative seismic energy of all ETS events other than those in Fig. 4 (ETS 3, 9, and 20). All symbols are the same as in Fig. 4.

**Fig. S2.** Plots of all ETS events showing the spatial distribution of cumulative seismic energy release in the plane of the plate interface, with symbols showing the rupture extent along the strike direction (black bar) and the start point (red star). Black rectangles represent the location of high-energy patches as shown in Fig. 4.

**Abbreviations**

SSE: Slow slip event; ETS: Episodic tremor and slip; RTR: rapid tremor reversal.

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**Authors’ contributions**

All authors contributed to the design of this study. KI and TU acquired the array data. KN and YH conducted the analyses. All authors contributed to the interpretation of results. YH and KN drafted the manuscript. All authors read and approved the final manuscript.

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**Availability of data and materials**
Short-term SSE data are available from the slow earthquake database (http://www-solid.eps.s.u-tokyo.ac.jp/~sloweq/). Array data used in this study are available from KI and TU upon reasonable request.

Competing interests

The authors declare that they have no competing interests.

References

Dieterich JH (1979) Modeling of rock friction: 1. Experimental results and constitutive equations. J of Geophys Res 84: 2161-2168. https://doi.org/10.1029/JB084iB05p02161

Dragert H, Wang K, James TS (2001) A silent slip event on the deeper Cascadia subduction interface. Science 292:1525–1528.

Ellsworth WL, Beroza GC (1995) Seismic evidence for an earthquake nucleation phase. Science 268:851-855. https://doi.org/ 10.1126/science.268.5212.851

Ghosh A. Vidale JE, Sweet JR, Creager KC, Wech AG (2009) Tremor patches at Cascadia revealed by array analysis. Geophys Res Lett 36: L17316. https://doi.org/10.1029/2009GL039080
Ghosh A, Vidale JE, Creager KC (2012) Tremor asperities in the transition zone control evolution of slow earthquakes. J Geophys Res 117: B10301. https://doi.org/10.1029/2012JB009249

Hirose H, Kimura T (2020) Slip distributions of short-term slow slip events in Shikoku, southwest Japan, from 2001 to 2019 based on tilt change measurements. J Geophys Res 125: e2020JB019601. https://doi.org/10.1029/2020JB019601

Hirose K, Obara K (2010) Recurrence behavior of short-term slow slip and correlated nonvolcanic tremor episodes in western Shikoku southwest Japan. J Geophys Res 115: B00A21. https://doi.org/10.1029/2008JB006050

Houston H, Delbridge BG, Wech AG, Creager KC (2011) Rapid tremor reversals in Cascadia generated by a weakened plate interface. Nature Geosci 4: 404–409. https://doi.org/10.1038/ngeo1157

Ide S (2008) A Brownian walk model for slow earthquakes. Geophys Res Lett 35: L17301. https://doi.org/10.1029/2008GL034821

Ide S (2019) Frequent observations of identical onsets of large and small earthquakes. Nature 573: 112–116. https://doi.org/10.1038/s41586-019-1508-5

Ide S, Beroza GC, Shelly DR, Uchide T (2007) A scaling law for slow earthquakes. Nature 447: 76–79. https://doi.org/10.1038/nature05780
Ide S, Imanishi K, Yoshida Y, Beroza GC, Shelly DR (2008) Bridging the gap between seismically and geodetically detected slow earthquakes. Geophys Res Lett 35: L10305. https://doi.org/10.1029/2008GL034014

Imanishi K, Takeda N, Kuwahara Y, Koizumi N (2011) Enhanced detection capability of non-volcanic tremor using a 3-level vertical seismic array network, VA-net, in southwest Japan. Geophys Res Lett 38: L20305. https://doi.org/10.1029/2011GL049071

Itaba S, Kitagawa Y, Koizumi N, Takahashi M, Matsumoto N, Takeda N (2012) The variation of the strain, tilt and groundwater level in the Shikoku District and Kii Peninsula, Japan (from May to October 2011). Report of the Coordinating Committee for Earthquake Prediction 87: 399-418. (in Japanese)

Itaba S, Kitagawa Y, Koizumi N, Takahashi M, Matsumoto N, Takeda N, Kimura H, Kimura T, Matsuzawa T, Shiomi K (2013a). Short-term slow slip events in the Tokai area, the Kii Peninsula and the Shikoku District, Japan (from May to October 2012). Report of the Coordinating Committee for Earthquake Prediction 89: 226-238. (in Japanese)

Itaba S, Kitagawa Y, Koizumi N, Takahashi M, Matsumoto N, Takeda N, Kimura H, Kimura T, Matsuzawa T, Shiomi K (2013b) Short-term slow slip events in the Tokai
area, the Kii Peninsula and the Shikoku District, Japan (from November 2012 to April 2013). Report of the Coordinating Committee for Earthquake Prediction 90: 254-269. (in Japanese)

Itaba S, Koizumi N, Takahashi M, Matsumoto N, Kitagawa Y, Takeda N, Kimura H, Kimura T, Matsuzawa T, Shiomi K (2014a) Short-term slow slip events in the Tokai area, the Kii Peninsula and the Shikoku District, Japan (from May to October 2013). Report of the Coordinating Committee for Earthquake Prediction 91: 230-242. (in Japanese)

Itaba S, Koizumi N, Takahashi M, Matsumoto N, Kitagawa Y, Ochi T, Takeda N, Kimura H, Kimura T, Matsuzawa T, Shiomi K (2014b) Short-term slow slip events in the Tokai area, the Kii Peninsula and the Shikoku District, Japan (from November 2013 to April 2014). Report of the Coordinating Committee for Earthquake Prediction 92: 238-249. (in Japanese)

Itaba S, Koizumi N, Takahashi M, Matsumoto N, Kitagawa Y, Ochi T, Takeda N, Kimura H, Kimura T, Matsuzawa T, Shiomi K (2015). Short-term slow slip events in the Tokai area, the Kii Peninsula and the Shikoku District, Japan (from May to October 2014). Report of the Coordinating Committee for Earthquake Prediction 93: 330-335. (in Japanese)
Kato A, Obara K, Igarashi T, Tsuruoka H, Nakagawa S, Hirata N (2012) Propagation of slow slip leading up to the 2011 Mw 9.0 Tohoku-Oki Earthquake. Science 335: 705-708. https://doi.org/10.1126/science.1215141

Kitagawa Y, Itaba S, Koizumi N, Takahashi M, Matsumoto N, Takeda N (2011). The variation of the strain, tilt and groundwater level in the Shikoku District and Kii Peninsula, Japan (from November 2010 to May 2011). Report of the Coordinating Committee for Earthquake Prediction 86: 519-533.

Kitagawa Y, Itaba S, Koizumi N, Takahashi M, Matsumoto N, Takeda N, Kimura H, Kimura T, Matsuzawa T, Shiomi K (2012). Short-term slow slip events in the Tokai area, the Kii Peninsula and the Shikoku District, Japan (from November 2011 to April 2012). Report of the Coordinating Committee for Earthquake Prediction 88: 303-311. (in Japanese)

Maeda T, Obara K (2009) Spatiotemporal distribution of seismic energy radiation from low-frequency tremor in western Shikoku, Japan. J Geophys Res 114: B00A09. https://doi.org/10.1029/2008JB006043

Mai PM, Spudich P, Boatwright J. (2005) Hypocenter locations in finite-source rupture models. Bull Seismo Soc Am 95: 965-980. https://doi.org/10.1785/012004011
Matsuzawa T, Hirose H, Shibazaki B, Obara K (2010) Modeling short- and long- term slow slip events in the seismic cycles of large subduction earthquakes. J Geophys Res 115: B12301. https://doi.org/10.1029/2010JB007566

Michel S, Gualandi A, Avouac J. (2019) Similar scaling laws for earthquakes and Cascadia slow-slip events. Nature 574: 522-526. https://doi.org/10.1038/s41586-019-1673-6

Neidell NS, Taner MT (1971) Semblance and other coherency measures for multichannel data. Geophysics 36: 482–297. https://doi.org/10.1190/1.1440186

Nishikawa T, Matsuzawa T, Ohta K, Uchide N, Nishimura T, Ide S (2019) The slow earthquake spectrum in the Japan Trench illuminated by the S-net seafloor observatories. Science 365: 808-813. https://doi.org/10.1126/science.aax5618

Obara K (2002) Nonvolcanic Deep Tremor Associated with Subduction in Southwest Japan. Science 296: 1679–1681. https://doi.org/10.1126/science.1070378

Obara K (2010) Phenomenology of deep slow earthquake family in southwest Japan: Spatiotemporal characteristics and segmentation. J Geophys Res 115: B00A25. https://doi.org/10.1029/2008JB006048

Obara K, Kato A (2016) Connecting slow earthquakes to huge earthquakes. Science 353: 253-257. https://doi.org/10.1126/science.aaf1512
Obara K, Tanaka S, Maeda T, Matsuzawa T (2010) Depth-dependent activity of non-volcanic tremor in southwest Japan, Geophys Res Lett 37: L13306. https://doi.org/10.1029/2010GL043679

Obara K, Matsuzawa T, Tanaka S, Maeda T (2012) Depth-dependent mode of tremor migration beneath Kii Peninsula, Nankai subduction zone. Geophys Res Lett 39: L10308. https://doi.org/10.1029/2012GL051420.

Ochi T, Itaba S, Koizumi N, Takahashi M, Matsumoto N, Kitagawa Y, Takeda N, Kimura H, Kimura T, Matsuzawa T, Shiomi K (2015) Short-term slow slip events in the Tokai area, the Kii Peninsula and the Shikoku District, Japan (from November 2014 to April 2015). Report of the Coordinating Committee for Earthquake Prediction 94: 250-261. (in Japanese)

Ohnaka M (1992) Earthquake source nucleation: A physical model for short-term precursors. Tectonophys 211: 149-178. https://doi.org/10.1016/0040-1951(92)90057-D

Rogers G, Dragert H (2003) Episodic tremor and slip on the Cascadia subduction zone: The chatter of silent slip. Science 300: 1942–1943. https://doi.org/10.1126/science.1084783
Sagae K, Nakahara H, Nishimura T, Imanishi K (2020) High resolution location of deep
low-frequency tremors beneath the Kii Peninsula, Nankai subduction zone, Japan,
using data from a dense seismic array. submitted to Geophys J Int.

Schmidt RO (1986). Multiple emitter location and signal parameter estimation, IEEE
Transactions on Antennas and Propagation 34: 276-280.

https://doi.org/10.1109/TAP.1986.1143830

Shelly DR, Beroza GC, Ide S (2007) Non-volcanic tremor and low-frequency
earthquake swarms. Nature 446: 305–307. https://doi.org/10.1038/nature05666

Shibazaki B, Matsu’ura M (1992) Spontaneous processes for nucleation, dynamic
propagation, and stop of earthquake rupture. Geophys Res Lett 19: 1189-1192.

https://doi.org/10.1029/92GL01072

Shiomi K, Matsubara M, Ito Y, Obara K (2008) Simple relationship between seismic
activity along Philippine Sea slab and geometry of oceanic Moho beneath southwest
Japan, Geophys J Int 173: 1018–1029. https://doi.org/10.1111/j.1365-
246X.2008.03786.x

Suzuki S, Okubo M, Imanishi K, Takeda N (2018) Detection method for pairs of P and
S waves of deep low-frequency earthquakes using a 3D array in the Tokai area of the
Nankai subduction and its application to hypocenter determination. G3 19: 4522-4540. https://doi/10.1029/2018GC007479

Thomas TW, Vidale JE, Houston H, Creager KC, Sweet JR, Ghosh A (2013) Evidence for tidal triggering of high-amplitude rapid tremor reversals and tremor streaks in northern Cascadia, Geophys Res Lett 40: 4254-4259. https://doi.org/10.1002/grl.50832

Ueno H, Hatakeyama S, Aketagawa T, Funasaki J, Hamada N (2002) Improvement of hypocenter determination procedures in the Japan Meteorological Agency. Q J Seismol 65: 123–134 (in Japanese with English abstract)

Wessel P, Smith WHG (1998) New improved version of the generic mapping tools released. Eos Trans. AGU 79: 579.

Yabe S, Ide S (2014) Spatial distribution of seismic energy rate of tectonic tremors in subduction zones. J Geophys Res 119: 8171–8185. https://doi.org/10.1002/2014JB011383

Figure captions
Fig. 1 a Location map of southwestern Japan showing the study area (black rectangle; Figs 4–6) and the array network (red square). Gray dots are tremor hypocenters from the NIED hybrid clustering catalog. Contour lines represent the depth of the Moho in the subducting Philippine Sea plate (Shiomi et al. 2008). Dotted square represents the area shown in c. b Configuration of stations (triangles) in the seismic array. The black triangle is the central station. c The distribution of short-term SSEs detected geodetically by GSJ during the study period. Gray rectangles are surface projections of areas on the fault planes of the short-term SSEs; the black side of each rectangle is the up-dip edge. d Space-time plot of tectonic tremors from the NIED hybrid clustering catalog. Distance along strike or along dip is measured from the corner with a black circle of the black rectangle in a. Circles represent hypocenters of tremor clusters for different ETS events analyzed in this study, colored according to their ETS number. ETS events 1, 8, and 11 were excluded from further analysis. Gray dots are tremor hypocenters that are isolated or belong to small ETS events. Gray vertical bars represent short-term SSEs detected geodetically by GSJ.

Fig. 2 a The distribution of epicenters of regular earthquakes (circles and star) and the location of the array network (red square). The epicenters are colored according to their
depth.  

b Plot showing the relation between the arrival time differences and the elevation differences for the earthquake shown by the star in a. Black line shows a linear regression line determined by the least-squares method with a constraint of passing through the origin.

Fig. 3 Space-time plots of tremor hypocenters in event ETS 17 a from this study and b from the NIED hybrid catalog. Gray circles in a are isolated events that were determined to not be part of an ETS event. c Cumulative count of tremor hypocenters (red) from this study and (blue) from the hybrid catalog.

Fig. 4 Examples of three ETS events with contrasting histories. (upper) The migration of tremor hypocenter clusters from their starting point (red star). The red square is the seismic array location. (lower) The spatial distribution of cumulative seismic energy release. Black rectangles indicate the location of high-energy patches.

Fig. 5 (upper) Plot showing the rupture extent along the strike direction (black bars) and the start point (red stars) of each ETS event. (bottom) The spatial distribution of
cumulative seismic energy release of all ETS events and the start point of each ETS event (red stars). Black rectangles indicate the location of high-energy patches.

**Fig. 6** Evolution of RTRs in a ETS 2 and b ETS 20. The top three panels show the cumulative seismic energy release during the period of each snapshot (upper left corner). Black rectangles show the locations of high-energy patches on the subducting Philippine Sea plate interface. The bottom plot shows the spatio-temporal distribution of tremor hypocenters, with dotted lines indicating migration speeds.

**Fig. 7 a** Evolution of the energy release for three ETS events with nearly identical start points. Each snapshot encompasses the time period shown in the upper left corner and displays the cumulative seismic energy release during that period. Black rectangles represent the locations of high-energy patches on the subducting Philippine Sea plate interface. Red stars indicate the start point of the ETS events. b Cumulative seismic energy release curves of the three ETS events. c Comparison of cumulative seismic energy release during the initial parts of the three ETS events. The start of the rupture in patch B serves as the zero point on which the curves are aligned so as to better compare the earliest phase of the ETS events.
Fig. 8 Logarithmic plot of the evolution of cumulative seismic energy release \(a\) for the three ETS events shown in Fig. 7b and \(b\) for all ETS events. The pink zone represents a scaling relation of \(M_0 \propto T\) (Ide et al. 2007) with a proportional relation to seismic energy with a constant of \(10^{-10}\). Black lines indicate scaling relations of \(E \propto T\), \(E \propto T^2\), and \(E \propto T^3\).

Table titles

Table 1 Duration of each ETS event for semblance analysis.

Table 2 Duration and cumulative seismic energy release of ETS events, estimated from hypocenter clusters with more than 10 tremors.