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Abstract

We estimated the long-term average slip rate on the plate interface across the Nankai subduction zone during 2002–2013 using deep low-frequency tremors as a proxy for short-term slow slip events based on empirical relations between the seismic moment of short-term slow slip events and tremor activities. The slip rate in each region is likely to compensate for differences between the convergence rate and the slip deficit rate of the subducting Philippine Sea plate estimated geodetically, although the uncertainty is large. This implies that the strain because of the subduction of the plate is partially stored as the slip deficit and partially released by slow slip events during the interseismic period. The excitation efficiency of the tremors for the slow slip events differs among regions: it is high in the northern Kii region. Some events in the western Shikoku region show a somewhat large value. Antigorite serpentinite of two types exists in the mantle wedge beneath southwest Japan. Slips with more effective excitation of tremors presumably occur in high-temperature conditions in the antigorite + olivine stability field. Other slip events with low excitation efficiency are distributed in the antigorite + brucite stability field. Considering the formation reactions of these minerals and their characteristic structures, events with high excitation efficiency can be correlated with a high pore fluid pressure condition. This result suggests that variation in pore fluid pressure on the plate interface affects the magnitude of tremors excited by slow slip events.

1. Introduction

In southwest Japan, the Philippine Sea (PHS) plate subducts beneath the Eurasian plate. Great earthquakes occur with a 100–150 year recurrence interval at the Nankai trough (e.g. Ando, 1975). Deep slow earthquakes, including nonvolcanic deep low-frequency (DLF) tremors (Obara, 2002), short-term slow slip events (SSE) (Obara et al., 2004), and very low-frequency earthquakes (Ito et al., 2007) occurring on the plate interface in southwest Japan have been reported from measurements obtained using dense seismic and geodetic networks throughout Japan. In southwest Japan, DLF tremors accompany short-term SSEs on the plate interface of the PHS plate (e.g., Obara et al., 2004). Such a spatiotemporal correlation, which has also been reported in Cascadia, is designated as an episodic tremor and slip (ETS) (Rogers and Dragert, 2003). Previous reports of some studies have described a proportional relation of ETS events between the moment of short-term SSEs observed geodetically and the size of DLF tremors as the apparent moment from the reduced displacement (Hiramatsu et al., 2008), duration (Aguiar et al., 2009), and number (Obara, 2010).

Monitoring of the slip at the ETS zone is important to elucidate the seismogenic process of interplate major earthquakes because DLF tremors and short-term SSEs take place at the deeper extension of such great earthquake source areas. Gao and Wang (2017) proposed a thermo-rheological model of slip behavior for ETS zones. In their model, the ETS zone is not a monotonic spatial transition of seismic to aseismic frictional behavior, but is instead controlled by the thermo-petrological condition in the mantle wedge corner. They showed that the ETS zone is separated from the mega-thrust earthquake zone in the Nankai subduction zone, suggesting that the neighboring creep drives episodic slips at the ETS zone. DLF tremors can be a proxy for short-term SSEs because minor tremor episodes of short duration are often observed with undetectable SSEs by geodetic observations because of the lack of a sufficient number of stations to capture the crustal deformation (Obara and Hirose, 2006). Hiramatsu et al. (2008) proposed an empirical procedure to estimate the long-term average slip rate attributable to short-term SSEs at the ETS zone on the plate interface from DLF tremors. Applying this method, the average slip rate has been estimated in the subduction zone along the Nankai trough. In Shikoku and the Kii Peninsula, the average slip rate is approximately consistent with differences between the convergence rate and the slip deficit rate of the subducting PHS plate (Hirose et al., 2010b; Ishida et al., 2013).

In the southern Kii Peninsula, however, Ishida et al. (2013) reported that the slip rate differed significantly from the difference between the convergence rate and the slip deficit rate. They suggested that this discrepancy resulted from the lateral variation in the proportional relation between the sizes of DLF tremors and SSEs, or the existence of a steady or quasi-steady slip. Yabe and Ide (2014) reported a heterogeneous distribution of the sizes of tremors defined by the band-limited seismic energy rate, and pointed out that the proportional relation between the sizes of DLF tremors and SSEs does not always hold.

Previous studies have used short-term SSEs data, observed by tiltmeters operated by the National Research Institute for Earth Science and Disaster Resilience (NIED) (Sekine et al., 2010) to estimate the average slip rate. Recently, Nishimura et al. (2013) detected short-term SSEs using Global Navigation Satellite System (GNSS) data. Their catalog includes SSEs that are undetectable to a tiltmeter. They reported that short-term SSEs might be undetected by both methods. Observations of SSEs conducted with multiple geodetic sensors are necessary to elucidate SSE characteristics.

This study re-estimated the long-term average slip rate at the ETS zone on the plate interface by re-analyzing DLF tremors and using short-term SSEs detected by tiltmeters and the GNSS. Subsequently, we assess the relation between the re-estimated average slip rate, the convergence rate, and the slip deficit rate of the PHS plate at the ETS zone. We also report heterogeneity of the size of DLF tremors excited by short-term SSEs and discuss the relation to mineral assemblages, antigorite serpentinite of two types, at the ETS zone.

2. Data and method

We analyzed waveform data recorded by a nationwide high-sensitivity seismograph network (Hi-net) operated by NIED, during April 2002 – July 2013 in the Tokai region, the Kii Peninsula and the Shikoku region of southwest Japan. We used hypocenter catalogs of DLF tremors determined using the hybrid method (Maeda and Obara, 2009) and the hybrid clustering method (Obara et al., 2010). The former is a method that combines the envelope correlation method (Obara, 2002) with spatial distributions of amplitude data, assuming tremor locations at the plate interface estimated by Shiomi et al. (2008). The latter catalog is processed using a clustering technique to estimate centroid locations and to eliminate non-tremor sources from the hybrid method. We used the former catalogue for visual checking of waveform data of DLF tremors and the latter one for tremor location.

We divided belt-like distributions of DLF tremors in the Kii Peninsula and the Shikoku region, respectively, into three segments bounded by the low-activity zones of DLF tremors: the northern, central, and southern areas of the Kii Peninsula (Ishida et al., 2013); and the eastern, central, and western areas of the Shikoku region (Hirose et al., 2010b) (Fig. 1). For this analysis, we selected Hi-net stations, which provide high S/N waveform data of DLF tremors, in each region. We used the Hi-net stations ASHH, ASUH, HOUH, STRH, and TDEH in the Tokai region (Hiramatsu et al., 2008); HYSH, KAWH, and TKWH for the northern area, and HNZH, TKEH, and TKWH for the central and southern areas of the Kii Peninsula (Ishida et al., 2013); and SINH, SADH, and IKWH for the eastern area, and IKKH, SJOH, and GHKH for the central area, and HITH, KWBH, OOZH, TBEH, and IKKH for the western area of the Shikoku region (Hirose et al., 2010b) (Fig. 1a). We also used 62 SSEs estimated from the NIED Hi-net tiltmeter data (Sekine et al., 2010; NIED, 2014) and 68 SSEs estimated from the GNSS data (Nishimura et al., 2013) (Fig. 1b). Some SSEs, but not others, were detected from both the tiltmeter and GNSS data. The numbers of SSEs detected by either or both the tiltmeter and GNSS in each region are presented in Table 1. For SSEs observed by both tiltmeter and GNSS, we used their average seismic moment. We corrected the seismic moment of SSEs observed by either tiltmeter or GNSS to fit the trend of the average seismic moment.

We estimated the seismic moment of the slip on the plate interface from DLF tremors following the empirical procedure proposed by Hiramatsu et al. (2008). First, we calculated the root-mean-square (RMS) amplitude of the vertical component for ground displacement by application of a band-pass filter of 2–10 Hz, integrating velocity waveform by time, and a moving average with a time window of 6 s. We define a DLF tremor event as follows: (1) the start time of a tremor is the time at which the RMS amplitude becomes larger than the noise level; (2) the end time of a tremor is the time at which the RMS amplitude becomes smaller than the noise level; (3) the duration is > 1 min; and (4) the maximum RMS amplitude is greater than twice the noise level. Next, we convert the observed amplitudes to reduced displacements (D_{RQ}), which are the RMS amplitudes corrected for geometrical spreading and inelastic attenuation, although previous works applied correction for geometrical spreading only. We calculated D_{RQ} as $D_{RQ} = A \cdot r \cdot exp(\pi f_c Q^{-1} t_i)$ (m^2), where A represents the RMS amplitude, r stands for the distance between the tremor source and the receiver, f_c signifies the center frequency, Q^{-1} denotes the intrinsic attenuation factor, and t_i is the travel time from the tremor source to the receiver. An intrinsic attenuation factor $Q = 184$ is also used (Maeda and Obara, 2009). We set the center frequency $f_c = 6$ (Hz) from a 2–10 Hz band-pass filter. In addition, we estimate the apparent moment as the event size of DLF tremors from the average of the time integral of the D_{RQ} amplitude at each station assuming the envelope of D_{RQ} as the apparent moment rate function (Hiramatsu et al., 2008). We calculate the sum of the apparent moment of an episode of the DLF tremors excited by a corresponding short-term SSE.

We used short-term SSEs observed geodetically. Then we estimated the conversion factor from the apparent moment of DLF tremors to the seismic moment of SSEs in each region. After obtaining the conversion factor, we can estimate the seismic moment of the SSE from multiplication of the conversion factor by the apparent moment of DLF tremors, even if the event is not detected geodetically (Hiramatsu et al., 2008; Hirose et al., 2010b; Ishida et al., 2013). Figure 2 presents the relation between the seismic moment of SSEs observed geodetically and the cumulative apparent moment of a corresponding episode of DLF tremors. In estimating the conversion factor, because SSEs propagate over two segmentations between the seven areas we set, we divided the study areas into five regions: Tokai, northern Kii, central and southern Kii, eastern Shikoku, and central and western Shikoku. For this study, we combined several episodes that are subdivided to express the migration of the slip area in the SSE catalog of NIED.

We estimated the conversion factor using the non-parametric bootstrap method. The size of the bootstrap sample is the same as that of the original data. The standard error and 95% confidence range were calculated from 2000 bootstrap estimations. The conversion factors, shown as the slopes in Fig. 2, are $5.3 \pm 1.1\ (3.7 - 8.1)\ \times 10^{16}\ \mathrm{N/\ m/s}$ in the Tokai region, $3.0 \pm 0.3\ (2.5 - 3.7)\ \times 10^{16}\ \mathrm{N/\ m/s}$ in the northern Kii region, $4.9 \pm 0.4\ (4.4 - 5.8)\ \times 10^{16}\ \mathrm{N/\ m/s}$ in the central and southern Kii region, $7.0 \pm 0.7\ (5.8 - 8.4)\ \times 10^{16}\ \mathrm{N/\ m/s}$ in the eastern Shikoku region, and $5.1 \pm 0.5\ (4.2 - 6.1)\ \times 10^{16}\ \mathrm{N/\ m/s}$ in the central and western Shikoku region, where values after

± and in parentheses respectively represent the standard error and the 95% confidence range (Table 2).

In earlier studies, the conversion factor of the Kii Peninsula was estimated using SSEs detected in the northern area (Ishida et al., 2013). That of the Shikoku region was estimated using SSEs detected in the western area (Hirose et al., 2010b) because SSEs were rarely observed in the central and southern Kii region or the eastern Shikoku region during the analysis periods. For this study, the conversion factor in the central and southern Kii regions is estimated as larger than that in the northern area. The conversion factor in the eastern Shikoku region is apparently larger than that in the central and western Shikoku regions, although the 95% confidence range overlaps slightly. The variation in these values and the widely various uncertainties, reflecting the scatter of data presented in Fig. 2, of the conversion factors might reflect variation in the frictional properties on the plate interface in each region. We discuss this point in Section 4.

3. Average slip rate estimated from DLF tremors

Fig. 3 presents the temporal variation in the cumulative seismic moment, at the ETS zone on the plate interface during the analysis period, as estimated from DLF tremors and the conversion factor. The steady increase in the cumulative seismic moment provides the long-term average rate of the moment release attributable to short-term SSEs on the plate interface. We can estimate the average slip rate. Table 2 presents the estimated seismic moment release rates in each region.

In addition, Fig. 3 shows that the cumulative seismic moment of SSEs detected geodetically is smaller than that estimated from DLF tremors. The rates of the seismic moment of the geodetically undetected SSEs are approximately 40%, 30%, 90%, 50%, and 55%, respectively, in the Tokai, northern Kii, central and western Kii, eastern Shikoku, and central and western Shikoku regions. The rate of the entire Nankai subduction zone is estimated as approximately 55%. We show the size–frequency distribution of SSEs detected geodetically in the Nankai subduction zone in two ways in Fig. 4: exponential and power-law distributions. We used here the average and corrected seismic moment described above as the size of the SSEs. Fig. 4 shows the lack of detection capability of the SSEs detected geodetically below the seismic moment of around 1.0×10^{18} Nm. Therefore, we calculated the best-fit line in Fig. 4 from data greater than 1.0×10^{18} Nm for each distribution. If we were able to observe smaller SSEs, as expected by the extrapolation of the best-fit line of the exponential distribution, then the cumulative seismic moment of geodetically detected SSEs would be close to that estimated from DLF tremors. However, extrapolation of the power-law distribution gives a much larger cumulative seismic moment, which implies that the size distribution of short-term SSEs is approximated by an exponential distribution rather than by a power law distribution, as reported by Hirose et al. (2010b).

To estimate the slip rate $\dot{U}$ at the ETS zone from the moment release rate $\dot{M}_o$, we used the formula $\dot{M}_o = \mu \dot{U} S$, where S is the fault area related to the slip of short-term SSEs in each region and μ is the rigidity (Hiramatsu et al., 2008). We approximate that the fault area is equal to the active area of DLF tremors as follows. We set 3 km $\times$ 3 km blocks on the plate interface and projected on the surface with a dip of 20°, which is the average dip of source faults of short-term SSEs observed geodetically. Then we counted blocks in which the number of the epicenters of DLF tremors was greater than five. We calculated the area on the plate interface from the number of such blocks. The estimated fault areas in each region are shown in Table 2. Assuming rigidity of 40 GPa (Hiramatsu et al., 2008; Hirose et al., 2010b; Ishida et al., 2013), we obtain average slip rates of $2.0 \pm 0.4\ (1.4-3.0)\ \mathrm{cm/year}$ in the Tokai region; $2.5 \pm 0.3\ (2.0-3.0)\ \mathrm{cm/year}$, $3.3 \pm 0.3\ (2.9-3.9)\ \mathrm{cm/year}$, and $3.8 \pm 0.3\ (3.4-4.5)\ \mathrm{cm/year}$, in the northern, central, and southern Kii regions, respectively; $3.9 \pm 0.4\ (3.2-4.6)\ \mathrm{cm/year}$, $2.6 \pm 0.2\ (2.1-3.1)\ \mathrm{cm/year}$, and $4.5 \pm 0.4\ (3.8-5.4)\ \mathrm{cm/year}$ in the eastern, central, and western Shikoku regions, respectively (Fig. 5 and Table 2). It is noteworthy that the standard error (values after $\pm$) and the 95% confidence range (values in parenthesis) are estimated by those of the conversion factors in each region. However, these estimations depend clearly on how the fault areas are estimated. For example, $\pm 20\%$ changes of the threshold of the number of DLF tremors per block results in $\pm 4-12\%$ changes of the slip rate in each area. Another factor on the uncertainty of the average slip rates is the seismic moment of SSEs. Sekine et al. (2010) reported that the uncertainty of the seismic moment of SSE was 17–77%. In addition, our dataset of SSEs includes the difference of estimations depending on the analyzed data and methods. Therefore, a considerable uncertainty of slip rate inferred from that of the seismic moment of SSEs can be larger than that of the fault area.

The depth distribution of the slip deficit rate on the plate interface is an important parameter for elucidating the seismogenic processes of great earthquakes in the subduction zone. Hirose et al. (2010b) emphasized that the average slip rates estimated from DLF tremors compensate for differences between the slip deficit rate at the ETS zone and the convergence rate of the PHS plate beneath the Shikoku region. Ishida et al. (2013) estimated the average slip rates at the ETS zone beneath the Kii Peninsula, and reported that the average slip rates are consistent with differences between the slip deficit rate and the convergence rate of the PHS plate, except in the southern area of the Kii Peninsula. We re-estimate the average slip rate from DLF tremors throughout southwest Japan and compare the balance between these rates: the average slip rate, the slip deficit rate, and the convergence rate of the PHS plate in each region.

The slip deficit rates are 1.5 cm/year in the Tokai region (Suito and Ozawa, 2009) ; 2.6 cm/year, 2.2 cm/year, and 1.0 cm/year in the northern, central, and southern Kii regions (Kobayashi et al., 2006); and 2.1 cm/year, 3.4 cm/year, and 2.6 cm/year in the eastern, central, and western Shikoku regions (Tabei et al., 2007). The convergence rates of the PHS plate are 3.0–4.0 cm/year in the Tokai region, 5.0–6.5 cm/year in the Kii Peninsula, and 6.5–6.8 cm/year in the Shikoku region (Miyazaki and Heki, 2001). The differences between the slip deficit rate and the convergence rate of the subducting plate are calculated as the following: 1.5–2.5 cm/year in the Tokai region; 2.4–3.9 cm/year, 2.8–4.3 cm/year, and 4.0–5.5 cm/year in the northern, central, and southern areas of the Kii

Peninsula; and 4.4–4.7 cm/year, 3.1–3.4 cm/year, and 3.9–4.2 cm/year in the eastern, central, and western areas of the Shikoku region. These values are approximately equal to the slip rates attributable to short-term SSEs estimated from DLF tremors in this study, except for the central Shikoku region in southwest Japan (Fig. 5), although the uncertainty of the slip rate is large (Table 2) and we cannot deny the possibility of a mismatch between these quantities. We infer that the strain at the ETS zone accumulated by the subduction of the PHS plate is partially stored as a slip deficit and partially released by short-term SSEs in the interseismic period, except for long-term SSE periods, beneath southwest Japan. The slip deficit rates cited in this study were estimated in the inter-long-term SSE period. Therefore, the slip rate and the slip deficit rate averaged over much longer period would be, respectively, larger and smaller.

Our result might resolve the inconsistency in the southern Kii region reported by Ishida et al. (2013) attributable to estimation of the typical conversion factor in the central and southern Kii regions because we re-analyze DLF tremors and use SSEs data detected by the GNSS in addition to the tiltmeter data. This result also emphasizes the importance of along-strike variation in the conversion factor for estimation of the average slip rate using the DLF tremors. In other words, simple estimation of slip rate over a subduction zone based on the activity of DLF tremors provides incorrect estimations. However, confirmation of this result requires more SSEs together with DLF tremors. The need exists for reanalysis to obtain more reliable estimation of the conversion factor and slip rate in future works.

In the central Shikoku region, the average slip rate we estimated might be less than the difference between the convergence rate and the slip deficit rate of the PHS plate. We assumed a constant conversion factor in the central and western Shikoku regions because nine SSEs share the fault area in these regions and only two SSEs exist in the central Shikoku region. However, the average frictional properties on the plate interface might be different in each region, implying a larger conversion factor in the central Shikoku region than in the western Shikoku region. If we estimate the conversion factors in the central and western Shikoku regions simultaneously using the least squares method for all DLF tremors and SSEs in these regions, then we obtain conversion factors of $9.7 \pm 1.7 \times 10^{16}\ \mathrm{N/m/s}$ and $4.6 \pm 0.4 \times 10^{16}\ \mathrm{N/m/s}$, respectively, giving an average slip velocity of $4.7 \pm 0.9\ \mathrm{cm/year}$ and $4.1 \pm 0.3\ \mathrm{cm/year}$ in the central and western Shikoku regions. As expected, the conversion factor in the central Shikoku region is estimated as larger than that in the western Shikoku region. However, the average slip rate in the central Shikoku region in this case is greater than the difference between the convergence rate and the slip deficit rate of the PHS plate in this region. This relation arises again from the small number of SSEs in the central Shikoku region. A sufficient number of SSEs in the central Shikoku region might enable us to estimate a reliable conversion factor and a slip rate in this region.

In the manner described above, we can estimate the conversion factors in the central and southern Kii regions simultaneously using the least squares method. The obtained conversion factors are $4.4 \pm 0.4 \times 10^{16}\ \mathrm{N/m/s}$ and $5.7 \pm 0.6 \times 10^{16}\ \mathrm{N/m/s}$, respectively, in the central and southern Kii regions. The average slip velocities are $3.0 \pm 0.3\ \mathrm{cm/year}$ and $4.4 \pm 0.5\ \mathrm{cm/year}$. These estimations imply a smaller conversion factor in the northern Kii region. The average slip velocities estimated here are apparently coincident with the difference between the convergence rate and the slip deficit rate of the PHS plate in this region. However, we emphasize that more SSEs are needed in the central and southern Kii regions for reliable estimations.

4. Heterogeneous distribution of the excitation efficiency of DLF tremors by short-term SSEs

In Section 2, we estimated conversion factors correlating the apparent moment of DLF tremors with the seismic moment of SSEs as the average values for seven regions to obtain the local average slip rates in the regions. We emphasize the variation in the conversion factor of each SSE, which is the ratio of the seismic moment of an SSE to the apparent moment of a coincident episode of DLF tremors. First, we estimate the average value of the conversion factor over southwest Japan, using all the SSEs analyzed in this study, to be $4.0 \pm 0.2 \times 10^{16}\ \mathrm{N/\ m/s}$ by least squares fitting. Next, we normalize a conversion factor for an individual SSE by the average one. Fig. 6 portrays a spatial distribution of the normalized conversion factors for SSEs in southwest Japan.

The DLF tremors sometimes migrate with a velocity of approx. 10 km/day (e.g. Obara, 2002). Obara (2010) reported that tremor migration is induced by the rupture initiation of the SSE because of the very sharp front of tremor migration, and sustained tremor activity behind the front.

Hirose and Obara (2010) estimated the slip propagation as 8–18 km/day for SSEs from time evolution inversion analysis of the slip on the plate interface. The coincidence of the tremor migration velocity with the slip propagation rate for SSEs indicates that the migrating tremors are caused by the rupture front of the SSE. Ando et al. (2010) showed that a physical model composed of heterogeneously and sparsely distributed unstable patches in a stable background can reproduce the migration pattern of the DLF tremors in terms of the rupture of patches triggered by the slow slip front.

The conversion factor represents the inverse of the excitation efficiency of DLF tremors by short-term SSEs. Our results (Fig. 6) demonstrate that the excitation efficiency differs among locations along the strike of the subducting PHS plate. It is particularly large in the northern Kii region. In the western Shikoku region, we find some events with the excitation efficiency as high as in the northern Kii region, although the scatter is large. Modification of the Q values by higher (+20%) and lower (-20%) values provides a similar along-strike variation in the excitation efficiency. Furthermore, we examine the effect of a heterogeneous Q structure on the spatial variation of the excitation efficiency by combining the results of the reference and a higher or lower, Q value. Figure 7 presents the variation in excitation efficiency for cases of a higher Q only in the northern Kii region (upper panels in Fig. 7) and only in the central and western Shikoku regions (lower panels in Fig. 7). The results show that the spatial variation in the excitation efficiency is not altered to any remarkable degree by a heterogeneous Q, although the contrast becomes weak in the western Shikoku region for the cases of a lower Q in the northern Kii region and a higher Q in the central and western Shikoku regions.

Ide and Yabe (2014) analyzed very low-frequency (0.02–0.05 Hz) (VLF) events that occurred in active tremor areas in the Nankai subduction zone. They estimated the scaled energy, the energy rate divided by seismic moment rate, of VLF events. They demonstrated the spatial distribution of the scaled energy of VLF events, as high in the entire Kii region and low in the Tokai and western Shikoku regions. The normalized conversion factors estimated in this study are low in the northern Kii region. Hiramatsu et al. (2008) reported that the conversion factor is proportional to the inverse of the scaled energy as a ratio of seismic moment of SSEs to radiated energy of DLF tremors. Therefore, it is interesting that the spatial distribution of the scaled energy of VLF events is almost coincident with that of the normalized conversion factors in spite of the difference in frequency band of the analyzed events. However, in the western Shikoku region, the normalized conversion factor might not be so high as expected by the inverse of the scaled energy of VLF events. If this is the case in the western Shikoku region, then this fact suggests that the frequency content radiated as seismic waves in a low-frequency to very low-frequency band relative to slow deformation differs from that in other regions.

High excitation efficiency indicates that the SSE tends to excite large magnitudes of DLF tremors in a slip, and vice versa. The heterogeneity in the excitation efficiency of DLF tremors by SSEs might derive from inhomogeneity of frictional properties of the slip planes, probably on the plate interface. In general, the frictional behavior of slow slip phenomena can be changed depending on the temperature conditions, the fluid amount, and the structural and rheological properties of fault gouge and rocks (Obara, 2011). What is the most likely cause for the heterogeneous tremor excitation?

DLF tremors and SSEs are regarded as occurring on the plate interface because of the high pore fluid pressure. As supporting evidence, seismic observations have revealed concentrations of fluid released from the oceanic crust of the subducting plate and/or serpentinization at the hanging wall mantle. In the Nankai subduction zone, three-dimensional seismic velocity structures presented by Matsubara et al. (2008) show high V_p/V_s zones along the distribution of DLF tremors at depths of 30 km. They suggested that DLF tremors might occur at all parts of the high V_p/V_s zone where the oceanic crust of the PHS plate encountered the serpentinized mantle wedge beneath the Eurasian plate. Shelly et al. (2007) and Kato et al. (2010) respectively reported that DLF tremors occurred along the top of a high V_p/V_s zones beneath the Shikoku and Tokai regions. Shibutani et al. (2009) and Saiga et al. (2013) also respectively reported the existence of a thin serpentinized layer on the plate interface beneath the Kii region from the receiver function and S-wave splitting analyses.

To consider the more detailed structures detected by our analyses, we particularly examine the mineral assemblages of serpentinite in the mantle wedge. Mizukami et al. (2014) discussed a possible causal relation between the characteristics of long-term and short-term slow slip behaviors in the Shikoku region, and two distinct types of antigorite serpentinite inferred in the hanging wall mantle. The formation of the antigorite (Atg) + brucite (Brc) serpentinite (hydration reaction: $\mathrm{olivine} + H_2O \rightarrow \mathrm{antigorite} + \mathrm{brucite}$) absorbs large amounts of H_2O. It can reduce pore fluid pressure, although the formation of the Atg + olivine (Ol) serpentinite (hydration reaction: $\mathrm{olivine} + H_2O + SiO_2 \rightarrow \mathrm{antigorite} + \mathrm{olivine}$) might generate high pore fluid pressure because the reaction is controlled by the amount of SiO_2 (Mizukami et al., 2014).

Short-term SSEs and deep low-frequency tremors are known to be active in the down dip portion of the source region of long-term SSE in the western Shikoku region. The source region for the long-term SSE is inferred to form a composite structure in which slip characteristics of the plate boundary change with depth. Nakata et al. (2017) fitted a two-segment model for heterogeneous slip behaviors of the long-term SSEs recorded in GEONET data. This fitting is coincident with the petrological model of Mizukami et al. (2014): the shallower portion of the long-term SSE region on the subduction boundary is inferred to be overlain by a mantle wedge with the Atg + Brc mineral assemblage. The mantle overlying the down dip portion is inferred to be near/within the Atg + Ol stability. As serpentinization of mantle wedge peridotite proceeds, the hydrated mantle weakens its ability to absorb slab-derived fluid. As a result, build-up of pore pressure is enhanced even in the Atg + Brc stability. The hybrid slip nature will be obscured because of antigorite serpentinization of two types. However, our petrological interpretation based on a recent thermal model of Nankai subduction zone (Fig. 8) shows that short-term SSEs in the Atg + Brc regions are less active in terms of frequency and magnitude, implying that slip planes for short-term SSE might be relatively minor and sparse in Atg + Brc stability. The heterogeneous occurrence of short-term SSEs indicates that the mantle beneath southwest Japan is not mature. It still affects fluid pressures in the Nankai subduction zone.

Fig. 8 presents a phase diagram for hydrous ultramafic rocks compiled by Mizukami et al. (2014). We compare the subduction geotherm of each region (Ji et al., 2016) with the petrological change along the plate interface. Ji et al. (2016) estimated the thermal structure on the plate interface of the PHS plate using a 3-D parallelepiped model of the subducting plate in the thermal convection. Events with a high excitation efficiency in the northern Kii and western Shikoku regions occur predominantly in the Atg + Ol stability field (Fig. 8). By contrast, events with a low excitation efficiency occur mostly in the Atg + Brc stability field. As an exception, events in the central Shikoku region, for which the excitation efficiency is low, lie in the Atg + Ol stability field (Fig. 8). The configuration of the subducting PHS plate bends greatly beneath the central Shikoku region (Shiomi et al., 2008). The 3D thermal simulation might not completely reproduce the effect of such complexity, causing a bias of thermal conditions on the plate interface. Perhaps for this reason, events in the central Shikoku region belong to the Atg + Ol stability field.

Magnitudes of DLF tremors detected on the surface are ascertained through propagation processes of seismic waves in overlying layers as well as source processes associated with short-term SSEs. The existence of brucite in mantle wedge reduces the elastic stiffness (Jian et al., 2006) and the mechanical strength upon shear (Moore and Lockner, 2007). From seismological observations, it is inferred that the tremor source is clusters of fluid-filled patches in which shear stress and pore fluid pressure are accumulated (Obara, 2011; Ando et al., 2010), and furthermore, that the fundamental structure of the tremor source may be common in the Atg + Ol and Atg + Brc regions. Important variables affecting the total moment release of DLF tremors are elastic energy in a single patch and spatial density of the patches on a slip plane of SSE and activation rate. Given that weak frictional strengths of the patches define critical shear stress to cause tremors, the variable rigidity of the wall rock mantle, that is mechanical strength, might not have marked effects on shear strain energy. Highly elevated fluid pressures reduce the effective normal stress and the fracture strength of the fault plane, and engender frequent SSE activities by stress perturbations (Obara, 2011). In this case, DLF tremors might also be excited easily by stress perturbation accompanied by short-term SSEs under high pore fluid pressure in the Atg + Ol stability field. If the collapse of fluid-filled patches is a fundamental process of tremor formation (Ando et al., 2010), then the extent of the elevation of pore pressure might be related directly to the tremor magnitude. Therefore, we suggest that the heterogeneity of the pore fluid pressure on the plate interface, induced by the phase transition of antigorite serpentinite, causes a heterogeneous distribution of the magnitude of DLF tremors excited by short-term SSEs.

5. Conclusions

After analyzing waveform data of DLF tremors recorded by Hi-net in southwest Japan, we have estimated the conversion factor from the apparent moment of DLF tremors of a corresponding episode to the seismic moment of short-term SSEs based on the proportional relation between their magnitudes. The cumulative seismic moment, as estimated from DLF tremors using the conversion factor, increases constantly over the long term and shows a steady moment release on the plate interface. We estimated the average slip rate at the ETS zone on the plate interface of the subducting PHS plate from DLF tremors. The average slip rates are apparently coincident with differences between the convergence rate and the slip deficit rate of the PHS plate, as estimated geodetically except for the central Shikoku region, although the uncertainty is large. Therefore, the short-term SSEs release the strain, except for slip deficit, accumulated through subduction of the PHS plate in the interseismic period at the Nankai subduction zone. We also investigated the spatial distribution of the conversion factor of each SSE. Actually, the conversion factor can be interpreted as the inverse of the excitation efficiency of DLF tremors by short-term SSEs. The excitation efficiency is distributed heterogeneously along the strike of the subducting PHS plate. Higher ones are distributed dominantly in the northern Kii. Slightly higher ones are also found in the western Shikoku region. We have compared the subduction geotherm beneath each region to the mineral assemblages of serpentinite in the mantle wedge. Events in the northern Kii and western Shikoku regions are predominant in the Atg + Ol assemblage, providing high pore fluid pressures. Others are predominant in the Atg + Brc assemblage. We suggest the variations in the pore fluid pressure and the mechanical strength of the hanging wall mantle as possible causes of the difference in the magnitudes of DLF tremors excited by short-term SSEs.

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Figure captions

Fig. 1. (a) Distribution of DLF tremors in southwest Japan. Dots show epicenters of DLF tremors (Obara et al., 2010). Triangles show Hi-net stations used for this study. The analysis area is divided into three segments in the Kii Peninsula and the Shikoku region, respectively, shown by rectangles with thick lines. (b) Dark gray and light gray rectangles respectively show locations of the fault models of SSEs observed by tiltmeter and GNSS.

Fig. 2. Relation between the seismic moment of short-term SSEs observed geodetically and the cumulative apparent moment of the corresponding episode of DLF tremors. Dotted, solid, and dashed lines respectively present the best-fit lines for the data in the Tokai region, Kii Peninsula, and Shikoku region.

Fig. 3. Temporal variation in the cumulative apparent moment (left axis) and in the seismic moment (right axis) from DLF tremors (red line). The light green and blue lines and arrows respectively indicate the cumulative seismic moment and the occurrences of SSEs observed by tiltmeter and GNSS. Pink hatched zones show SSEs observed by both tiltmeter and GNSS. Gray hatched zones show the period of long-term SSEs in Tokai (Ozawa et al., 2002; Suito and Ozawa, 2009), and in the Bungo channel (Hirose and Obara, 2005; Hirose et al., 2010a).

Fig. 4. Size–frequency distribution of geodetically detected SSEs: (a) exponential distribution and (b) power-law distribution. The solid line shows the best-fit line for data greater than 1.0×10^{18} Nm. R^2 is the coefficient of determination of the fitting.

Fig. 5. Average slip rates at the ETS zone on the plate interface estimated from DLF tremors in southwest Japan. Solid and open red arrows respectively indicate the slip rates attributable to short-term SSEs estimated from DLF tremors in this study and in previous studies (Hirose et al., 2010b; Ishida et al., 2013). Black arrows indicate the slip deficit rates at the ETS zone on the plate interface of the subducting PHS plate from the Nankai Trough (Tabei et al., 2007; Kobayashi et al., 2006; Suito and Ozawa, 2009). Blue arrows show the convergence rates of the PHS plate (Miyazaki and Heki, 2001).

Fig. 6. Distribution of the normalized conversion factors of each SSE by the average one estimated from all SSEs in southwest Japan depicted in Fig. 3. Circles and thick lines respectively portray the central points and the upper edges of the fault plane of SSEs. The dashed lines represent depth contours of the oceanic Moho discontinuity (Shiomi et al., 2008).

Fig. 7. Distributions of the normalized conversion factors of each SSE by the average one estimated

from all SSEs for heterogeneous Q structures, with lower (-20%) and higher (+20%) Q values in the northern Kii region and in the central and western Shikoku regions. All symbols are the same as those presented in Fig. 5.

Fig. 8. (a) Phase diagram of serpentinite (modified from Mizukami et al., 2014). The dashed curves, which correspond to the lines in Fig. 6b, indicate the temperature structure beneath southwest Japan. The stars show high-activity areas of tremors and SSEs in each region. (b) Red curves show isotherms of the upper surface of the PHS plate (Ji et al., 2016).

Table 1. Summary of the number of SSEs, as detected with either or both the tiltmeter and GNSS data, in each region used for this study.

Table 2. Results for estimation of the average slip rate and each rate of the subducting PHS plate in the Tokai, northern Kii, central Kii, southern Kii, eastern Shikoku, central Shikoku, and western Shikoku regions. Values after ± and in parentheses respectively represent the standard error and the 95% confidence range.

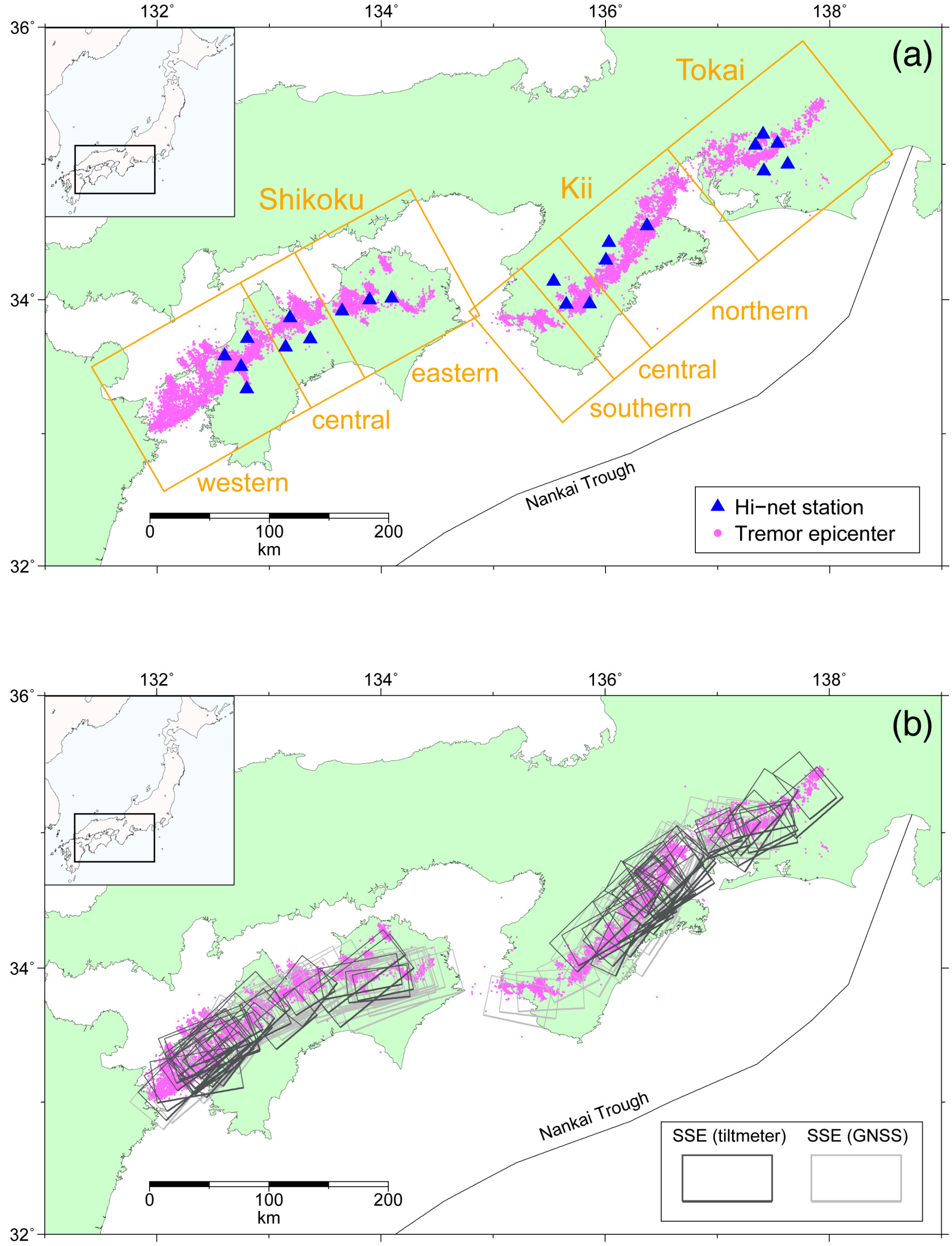

(a)
132°
134°
136°
138°
36°
34°
32°
Tokai
Shikoku
Kii
northern
central
southern
eastern
central
western
Nankai Trough
0
100
200
km
Hi−net station
Tremor epicenter
(b)
SSE (tiltmeter)
SSE (GNSS)

Figure 1.

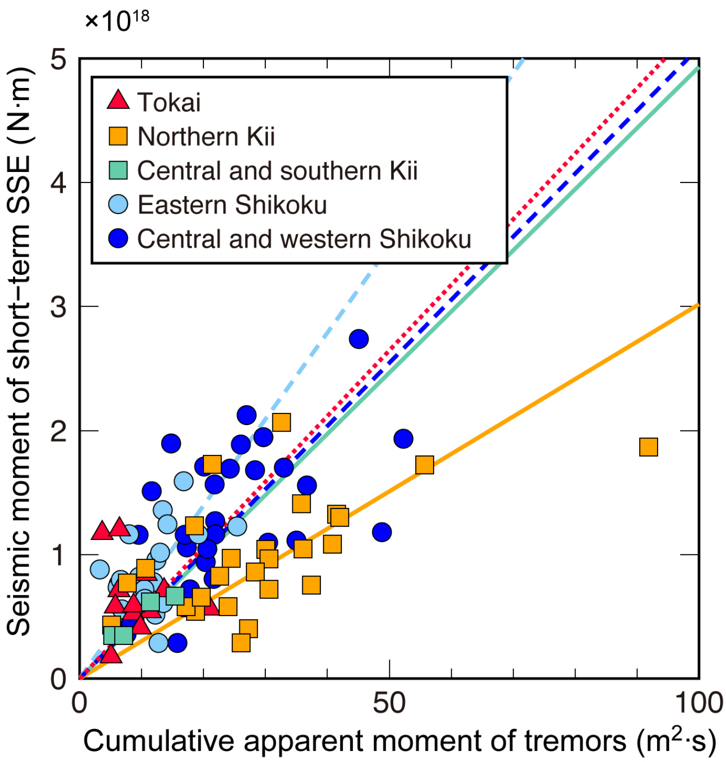
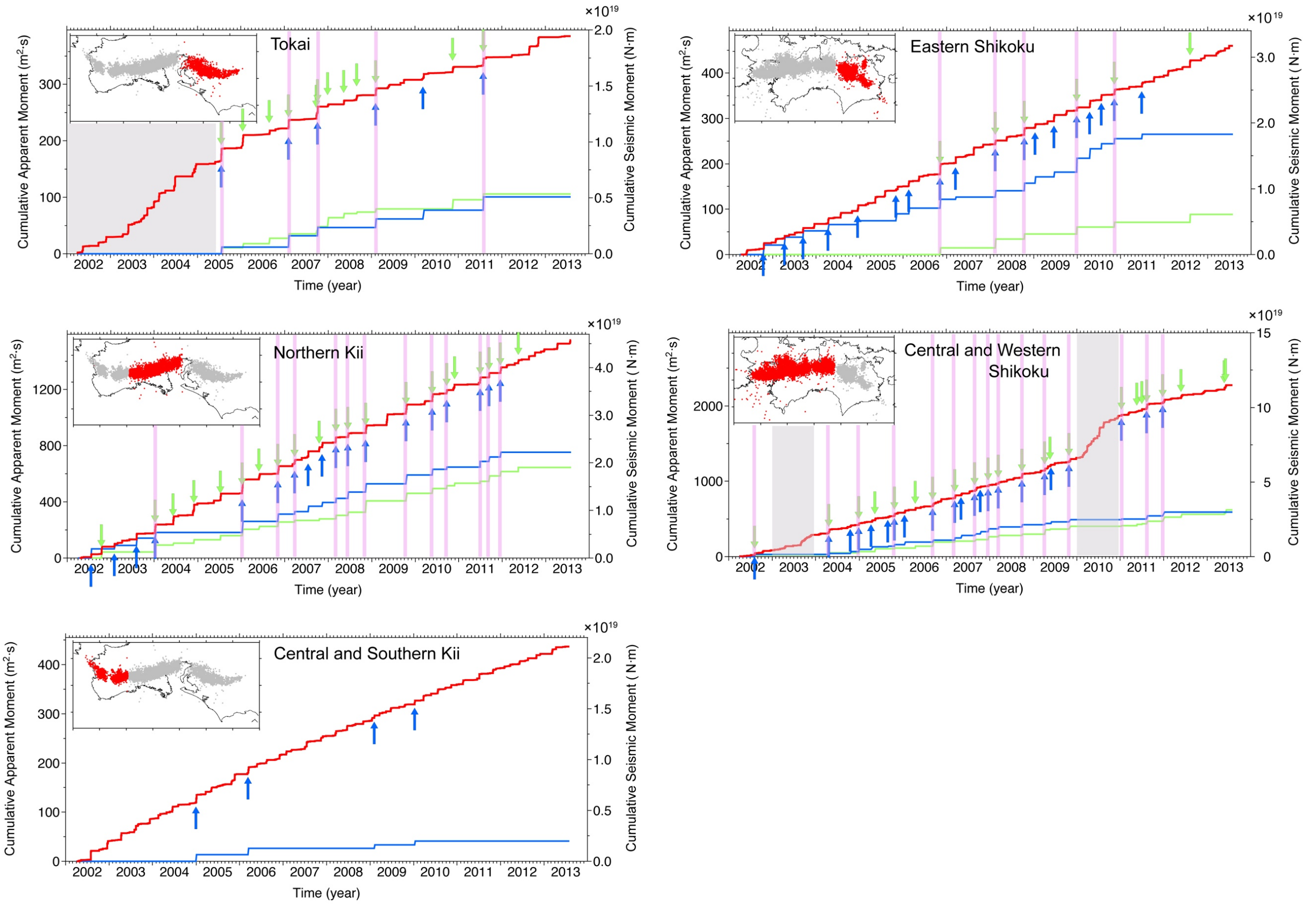

×10^18
5
4
3
2
1
0
Seismic moment of short−term SSE (N·m)
Tokai
Northern Kii
Central and southern Kii
Eastern Shikoku
Central and western Shikoku
0
50
100
Cumulative apparent moment of tremors (m2·s)

Figure 2.



Tokai
Eastern Shikoku
Northern Kii
Central and Western Shikoku
Central and Southern Kii
Cumulative Apparent Moment (m^2·s)
Cumulative Seismic Moment (N·m)
×10^{19}
Time (year)

Figure 3.

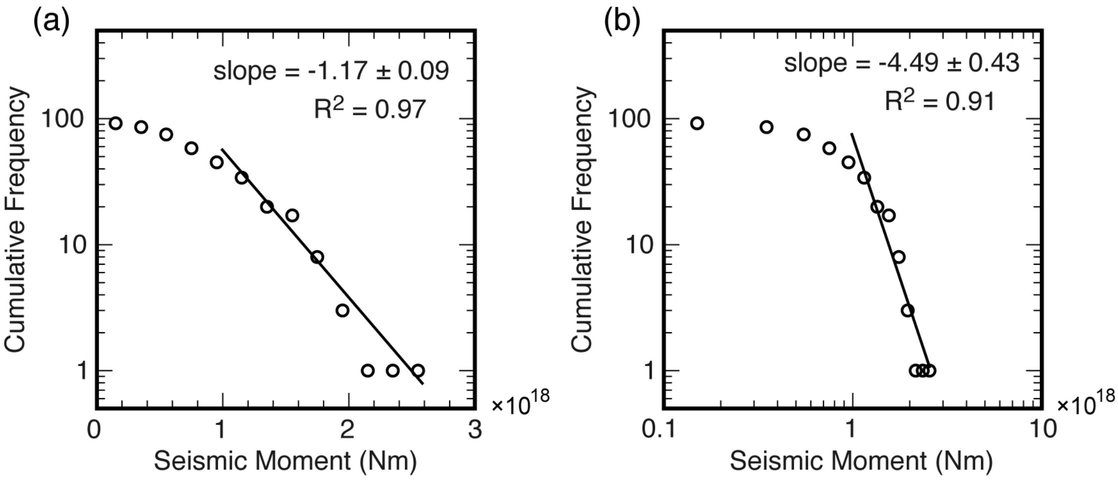

(a)
slope = -1.17 ± 0.09
R^2 = 0.97
Cumulative Frequency
Seismic Moment (Nm)
×10^18
(b)
slope = -4.49 ± 0.43
R^2 = 0.91
Cumulative Frequency
Seismic Moment (Nm)
×10^18

Figure 4.

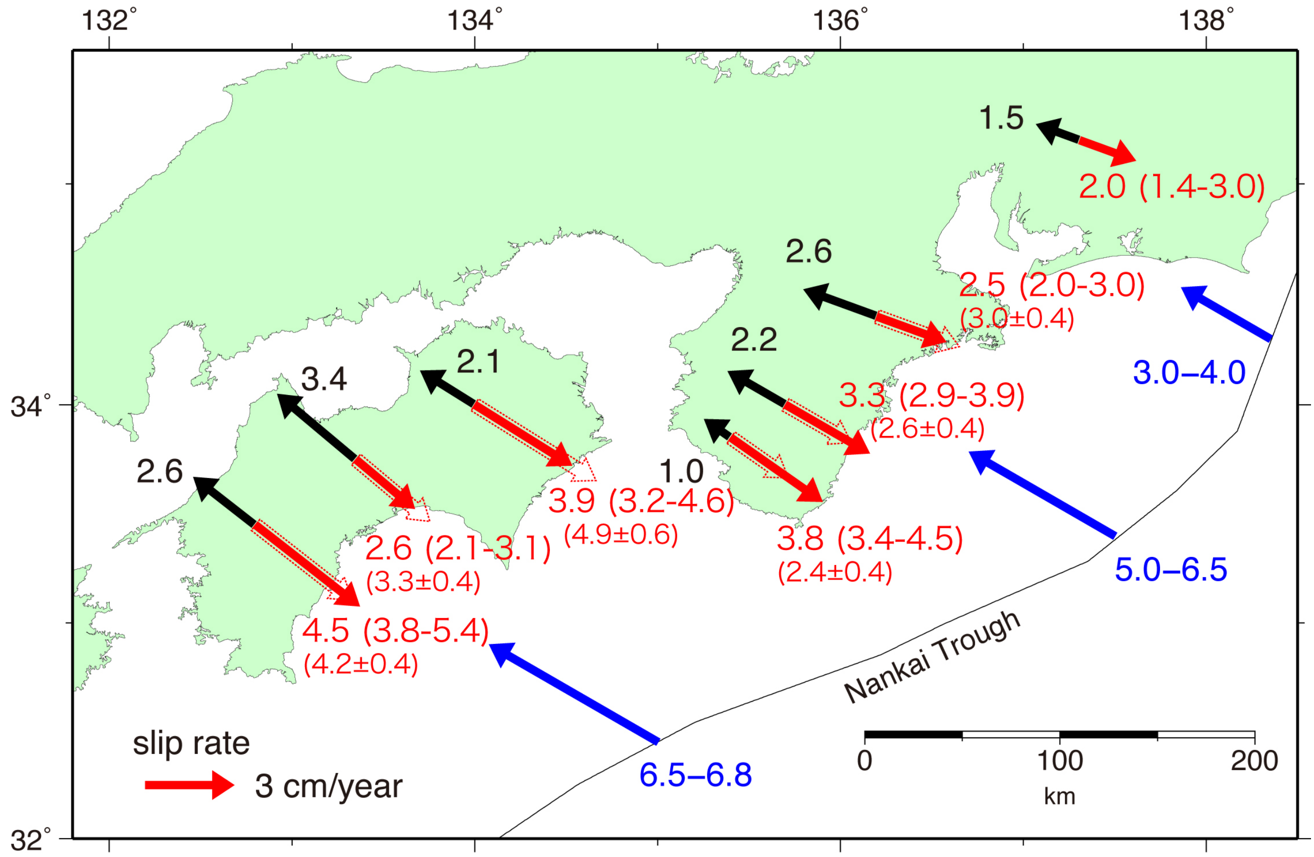

132°
134°
136°
138°
34°
32°
1.5
2.0 (1.4-3.0)
2.6
2.5 (2.0-3.0)
(3.0±0.4)
3.0–4.0
2.2
3.3 (2.9-3.9)
(2.6±0.4)
2.1
3.4
1.0
2.6
3.9 (3.2-4.6)
(4.9±0.6)
3.8 (3.4-4.5)
(2.4±0.4)
5.0–6.5
2.6 (2.1-3.1)
(3.3±0.4)
4.5 (3.8-5.4)
(4.2±0.4)
Nankai Trough
slip rate
3 cm/year
6.5–6.8
0
100
200
km

Figure 5.

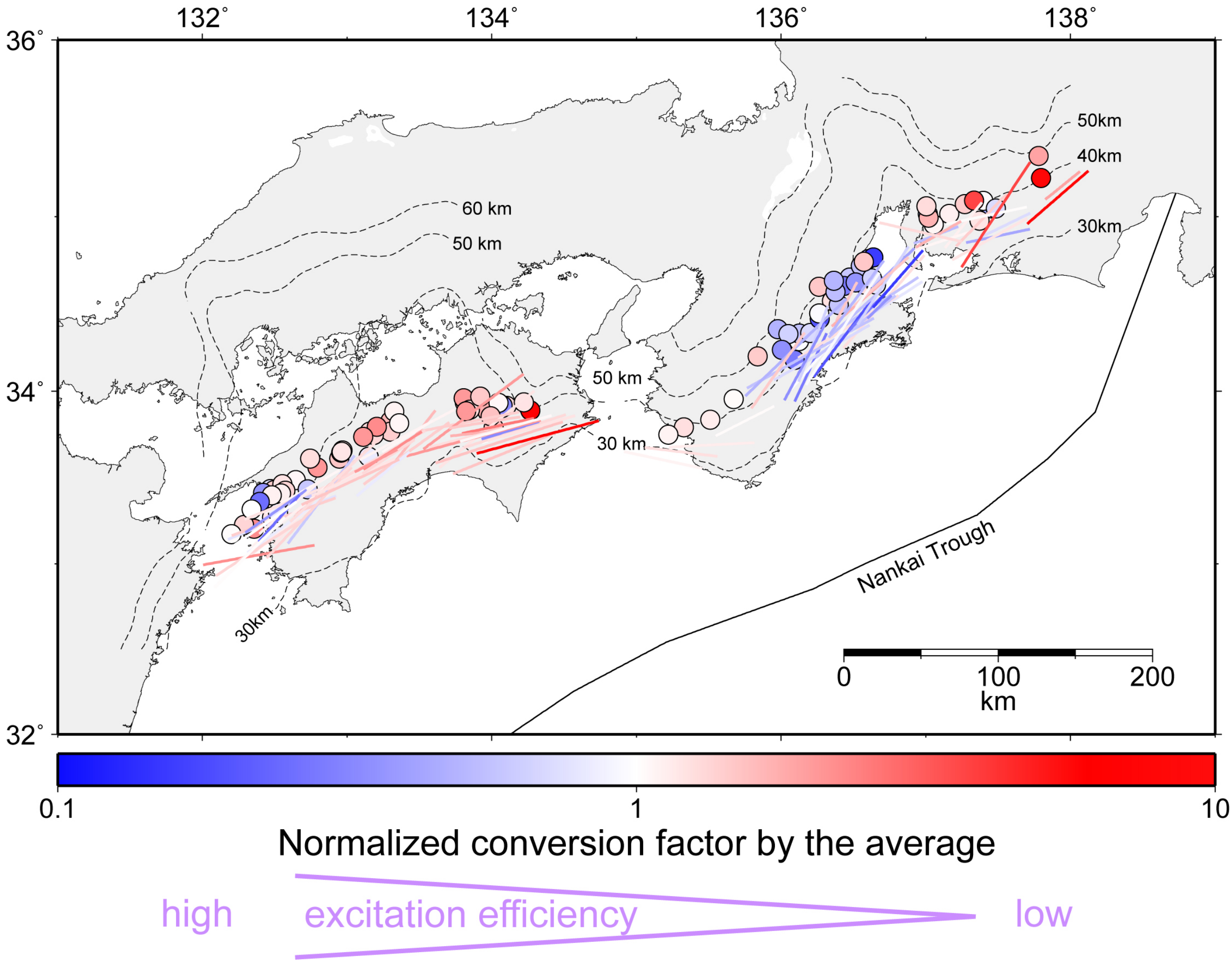
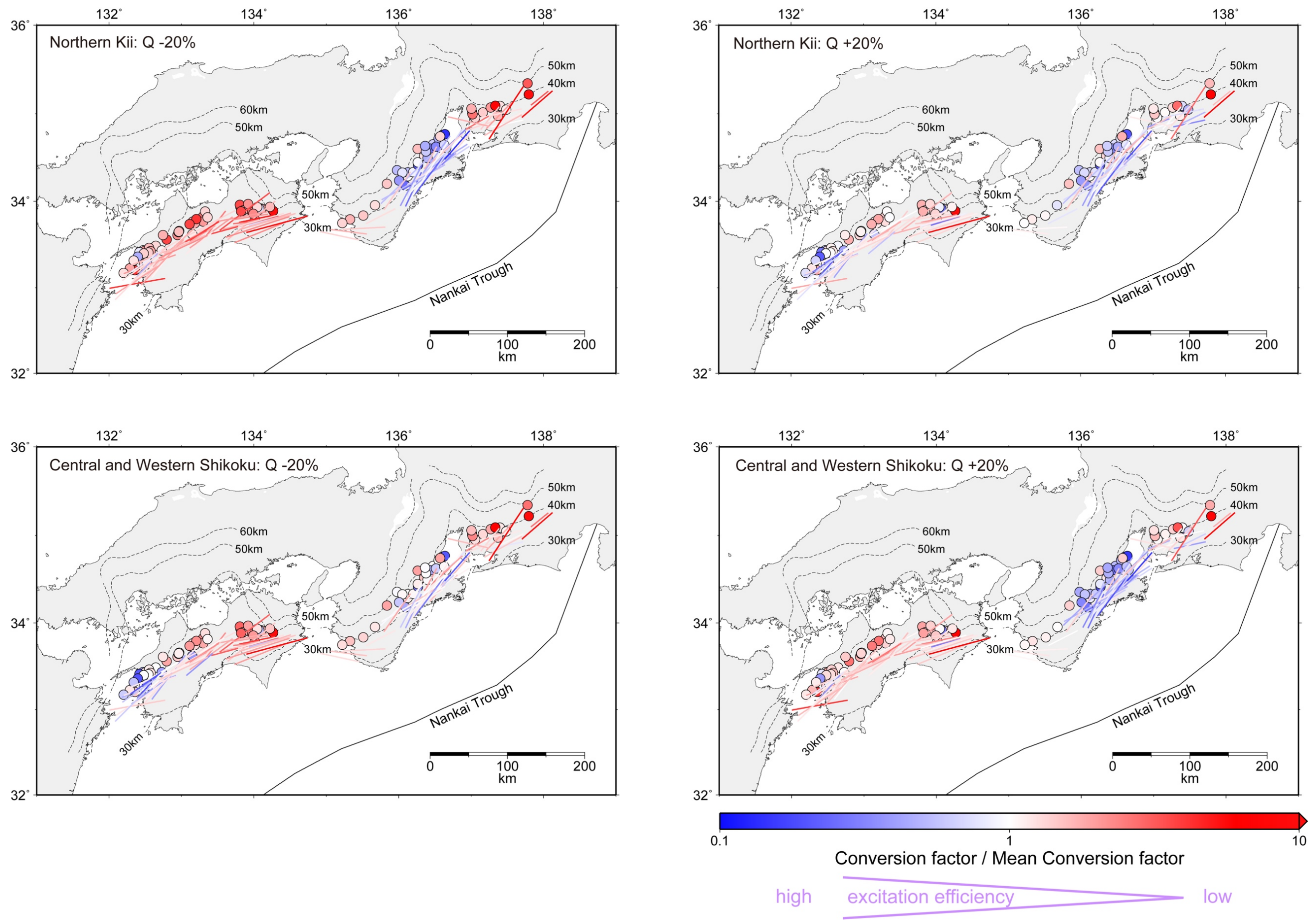

132°
134°
136°
138°
36°
34°
32°
60 km
50 km
50 km
30 km
30km
50km
40km
30km
Nankai Trough
0
100
200
km
0.1
1
10
Normalized conversion factor by the average
high
excitation efficiency
low

Figure 6.



Northern Kii: Q -20%
Northern Kii: Q +20%
Central and Western Shikoku: Q -20%
Central and Western Shikoku: Q +20%
132°
134°
136°
138°
36°
34°
32°
60km
50km
40km
30km
Nankai Trough
0
100
200
km
0.1
1
10
Conversion factor / Mean Conversion factor
high
excitation efficiency
low

Figure 7.

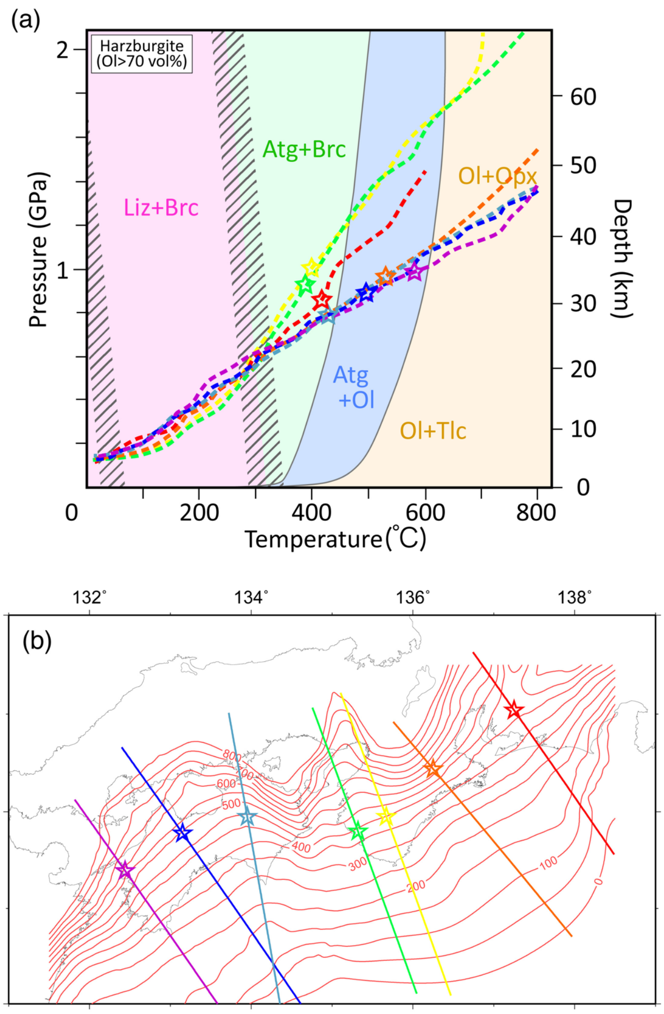

(a)
Harzburgite
(Ol>70 vol%)
Liz+Brc
Atg+Brc
Atg
+Ol
Ol+Opx
Ol+Tlc
Pressure (GPa)
Depth (km)
Temperature(°C)
(b)
132°
134°
136°
138°
36°
34°
32°

Figure 8.

	SSE from both tiltmeter and GNSS	SSE from tiltmeter	SSE from GNSS
Tokai	5	7	1
northern Kii	13	8	5
central and southern Kii	-	-	4
eastern Shikoku	5	1	13
central and western Shikoku	15	8	7

Table 1.

	conversion factor (10^{16} N/m/s)	moment release rate (10^{18} Nm/yr)	fault area (10^9 m^2)	slip rate (cm/yr)	convergence rate (cm/yr)	slip deficit rate (cm/yr)	convergence rate - slip deficit rate (cm/yr)
Tokai	5.3 ± 1.1 (3.7-8.1)	1.5 ± 0.3 (1.0-2.2)	1.8	2.0 ± 0.4 (1.4-3.0)	3.0 - 4.0[a]	1.5[b]	1.5 - 2.5
northern Kii	3.0 ± 0.3 (2.5-3.7)	3.1 ± 0.3 (2.5-3.8)	3.1	2.5 ± 0.3 (2.0-3.0)	5.0 - 6.5[a]	2.6[c]	2.4 - 3.9
central Kii	4.9 ± 0.4 (4.4-5.8)	1.1 ± 0.1 (1.0-1.3)	0.8	3.3 ± 0.3 (2.9-3.9)	5.0 - 6.5[a]	2.2[c]	2.8 - 4.3
southern Kii	4.9 ± 0.4 (4.4-5.8)	0.8 ± 0.1 (0.7-1.0)	0.5	3.8 ± 0.3 (3.4-4.5)	5.0 - 6.5[a]	1.0[c]	4.0 - 5.5
eastern Shikoku	7.0 ± 0.7 (5.8-8.4)	2.8 ± 0.3 (2.4-3.4)	1.8	3.9 ± 0.4 (3.2-4.6)	6.5 - 6.8[a]	2.1[d]	4.4 - 4.7
central Shikoku	5.1 ± 0.5 (4.2-6.1)	1.5 ± 0.1 (1.3-1.8)	1.5	2.6 ± 0.2 (2.1-3.1)	6.5 - 6.8[a]	3.4[d]	3.1 - 3.4
western Shikoku	5.1 ± 0.5 (4.2-6.1)	6.7 ± 0.6 (5.6-7.9)	3.7	4.5 ± 0.4 (3.8-5.4)	6.5 - 6.8[a]	2.6[d]	3.9 - 4.2

[a]Data from Miyazaki and Heki (2001), [b]Data from Suito and Ozawa (2009), [c]Data from Kobayashi et al. (2006), [d]Data from Tabei et al. (2007).

Table 2