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Non-volcanic deep low-frequency tremors accompanying slow slips in the southwest Japan subduction zone

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Abstract

Non-volcanic deep low-frequency tremors in southwest Japan exhibit a strong temporal and spatial correlation with slow slip detected by the dense seismic network. The tremor signal is characterized by a low-frequency vibration with a predominant frequency of 0.5–5 Hz without distinct P- or S-wave onset. The tremors are located using the coherent pattern of envelopes over many stations, and are estimated to occur near the transition zone on the plate boundary on the forearc side along the strike of the descending Philippine Sea plate. The belt-like distribution of tremors consists of many clusters. In western Shikoku, the major tremor activity has a recurrence interval of approximately six months, with each episode lasting over a week. The tremor source area migrates during each episode along the strike of the subducting plate with a migration velocity of about 10 km/day. Slow slip events occur contemporaneously with this tremor activity, with a coincident estimated source area that also migrates during each episode. The coupling of tremor and slow slip in western Shikoku is very similar to the episodic tremor and slip phenomenon reported for the Cascadia margin in northwest North America. The duration and recurrence interval of these episodes varies between tremor clusters even on the same subduction zone, attributable to regional difference in the frictional properties of the plate interface.

Keywords: Deep low-frequency tremor; Deep low-frequency earthquake; Slow slip; Dehydration; Subduction zone

1. Introduction

The Japan Islands are located in a complicated subduction zone, where the Philippine Sea plate and the Pacific plate subduct beneath the Eurasian plate (Fig. 1). This area exhibits high seismicity not only in association with the descending oceanic plate but also in the upper crust. On the mainland, deep low-frequency (DLF) microearthquakes often occur near Quaternary volcanoes at depths close to the Mohorovicic (Moho) discontinuity (Ukawa and Obara, 1993; Hasegawa and Yamamoto, 1994). DLF events are distinguished from other equivalently sized ordinary earthquakes by the depth of the hypocenter and the dominant frequency of the seismogram, and are considered to represent special phenomena that are specific to particular regions, such as volcanoes and faults. The medium at depths of approximately 30 km below the island arc is usually ductile and regarded as aseismic, except in the region of plate boundaries. The source process of DLF events is therefore expected to differ from that of ordinary earthquakes. According to Nishidomi and Takeo (1996) and Okada and Hasegawa (2000), DLF events in volcanic areas occur due to the upwelling of magma from the mantle to the lower crust, suggesting that the

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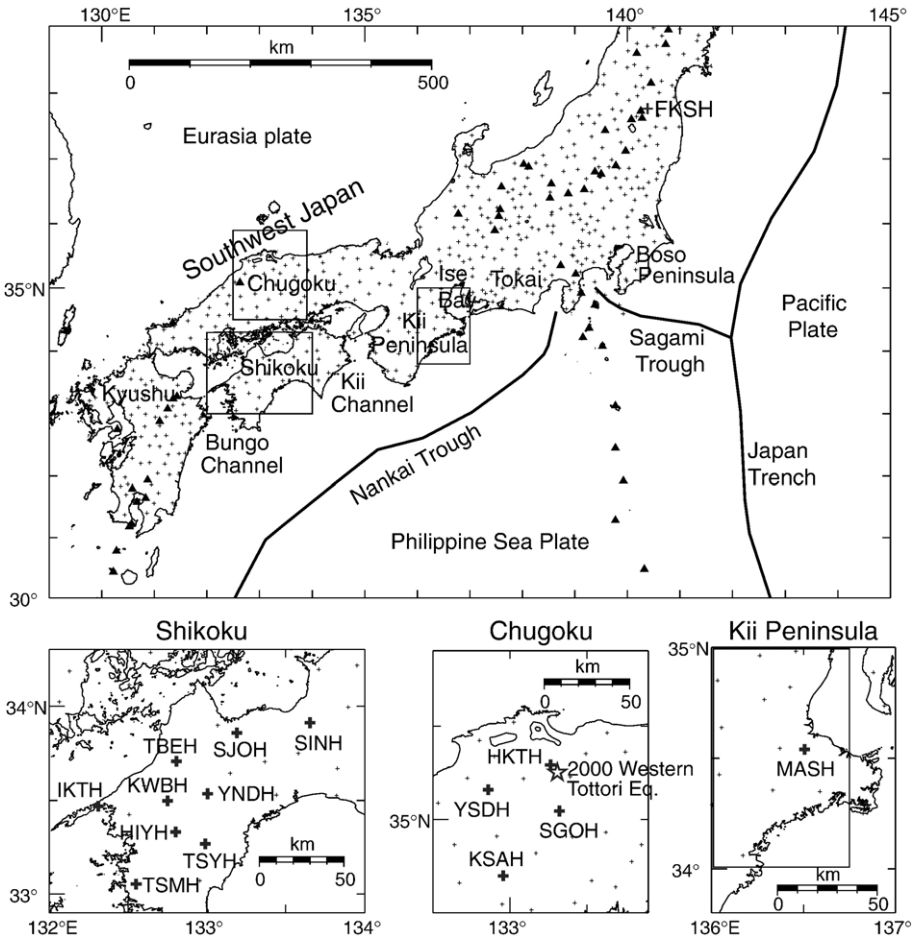

Fig. 1. Study area. Bold lines represent trench axes (plate boundaries). Both the Philippine Sea plate and Pacific plate subduct beneath the Eurasian plate in this area. Solid triangles denote Quaternary active volcanoes defined by JMA and thin crosses indicate NIED Hi-net seismic stations. Bold crosses with station code denote stations involved in the present analyses. The star in the lower middle panel indicates the epicenter of the 2000 Western Tottori Earthquake, which is very close to the epicenter of the Tottori DLF event (see Fig. 5).

migration of fluid is a possible mechanism of DLF events. Similar DLF events have also been observed near active fault systems (Ohmi, 2002). For example, DLF events occur frequently at depths close to the Moho discontinuity near the source area of the 2000 Western Tottori Earthquake (M_w 6.7). Based on an analysis of the amplitude ratio of P waves to S waves and the particle motion of S waves, Ohmi and Obara (2002) estimated the source mechanism of the DLF event to be a single-force process. This result supports the model of fluid flow as the mechanism driving DLF events.

Following the disastrous 1995 Great Hanshin-Awaji (Kobe) Earthquake, a very dense high-sensitivity seismograph network was deployed throughout Japan by the National Research Institute for Earth Science and Disaster Prevention (NIED) to promote basic research on seismic activity and further investigation of future large earthquakes (Obara, 2002a; Okada et al., 2004). This seismic network, called "NIED Hi-net" offers improved detection capability for microearthquakes, and the high-quality dense network of stations has lead to new seismological observations. One of the most unexpected results revealed by this network is the occurrence of non-volcanic deep tremors in southwest Japan (Obara, 2002b). The signals of these tremors are similar to those for DLF events, but the sequence of tremor activity lasts for much longer, from hours to weeks. The source area of the tremors extends for 600 km along the strike of the subducting Philippine Sea plate, suggesting that the tremors may reflect the subduction process.

Only very recently, crustal tilt monitoring has revealed the existence of slow slip events that occur in temporal and spatial coincidence with the active stages of tremors (Obara et al., 2004). The coupling between the tremors and slow slips is very similar to

that for the episodic tremor and slip (ETS) detected in the Cascadia margin of the North American continent (Rogers and Dragert, 2003). It has been clarified through global positioning system (GPS) monitoring that ETS events occur repeatedly at intervals of 13 to 16 months, with each episode lasting for a few weeks as

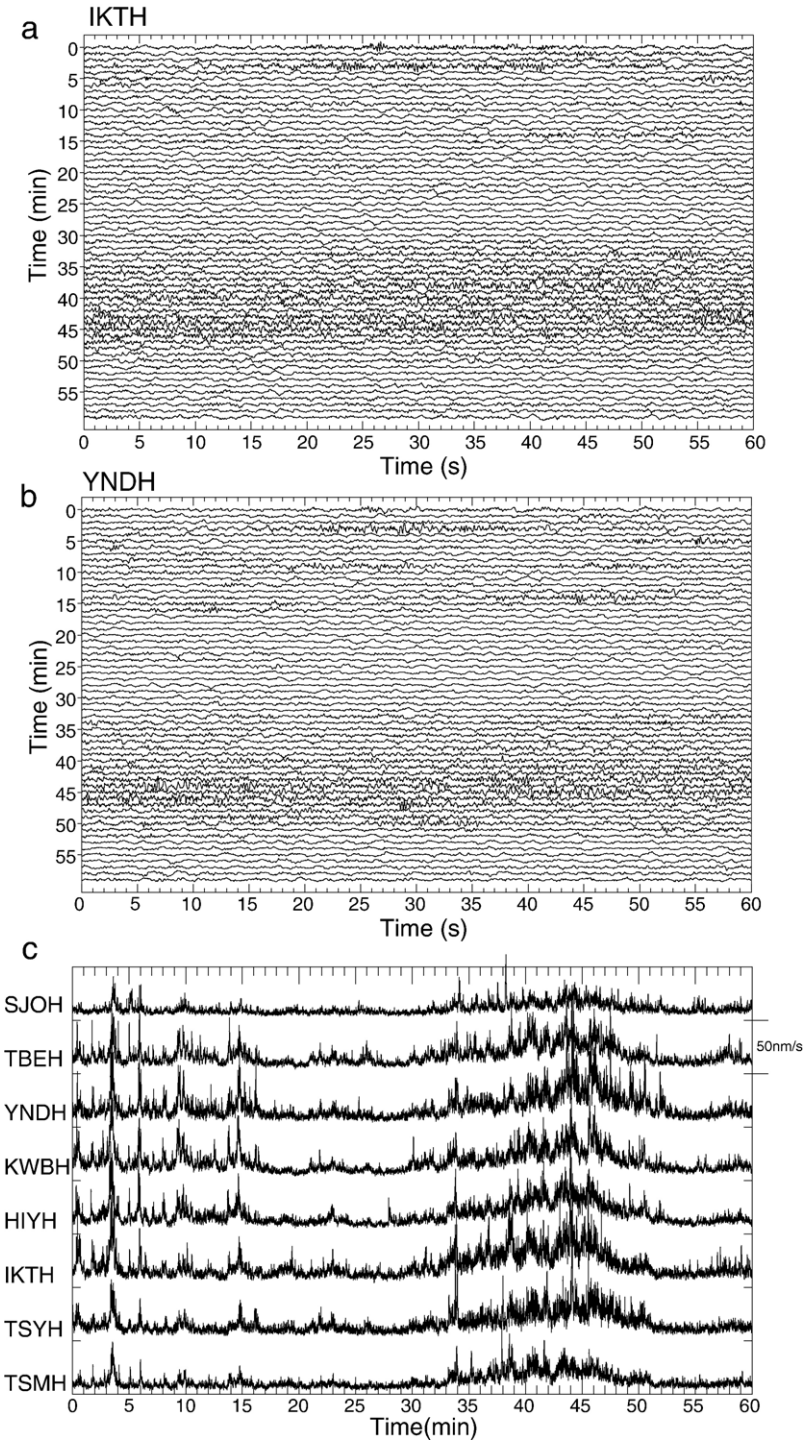

Fig. 2. Seismograms of deep low-frequency tremors. (a,b) 1-h continuous record of vertical component observed at stations IKTH and YNDH in western Shikoku at 04:00 on August 17, 2001. (c) 1-h envelope seismograms for a frequency range of 2–16 Hz observed at eight stations in western Shikoku.

it propagates along the strike of the subducting plate for up to 300 km. The migration velocity is ranging from 6 to 15 km/day (Dragert et al., 2001). These characteristics of ETS are very similar to the coupling of tremors and crustal tilting detected in western Shikoku.

It is important to clarify the activity and relationship between these tremors and slow slip in order to improve our understanding of the subduction process in this region. In this paper, the characteristics of non-volcanic deep tremors and slow slip events in southwest Japan are examined and compared with those of similar phenomena in other areas.

2. Characteristics of low-frequency tremor

An example of a 1-h continuous recording containing the signal of a deep tremor observed in western Shikoku (stations IKTH and YNDH) is shown in Fig. 2. The recording shows a relatively higher-frequency wave train lasting for 10 min from 04:35, which can be seen in the microseism noise whose dominant frequency is lower than 1 Hz caused by ocean swell and surf. Although the wave train appears similar to ground noise, such as that due to motor vehicles, similar wave trains were observed simultaneously at neighboring stations located over 10 km distance. The wave trains therefore cannot be attributed to human-made noise, and are considered to represent natural seismic events. The frequency content of the signal is similar to that of DLF microearthquakes, as shown in Fig. 3. This tremor event is referred to as a deep low-frequency tremor after the DLF microearthquakes, because the depth of the tremor is estimated to be around 30 km, similar to that for DLF earthquakes.

The tremor signal is composed of random wave trains with frequency content of 0.5–5 Hz. Within

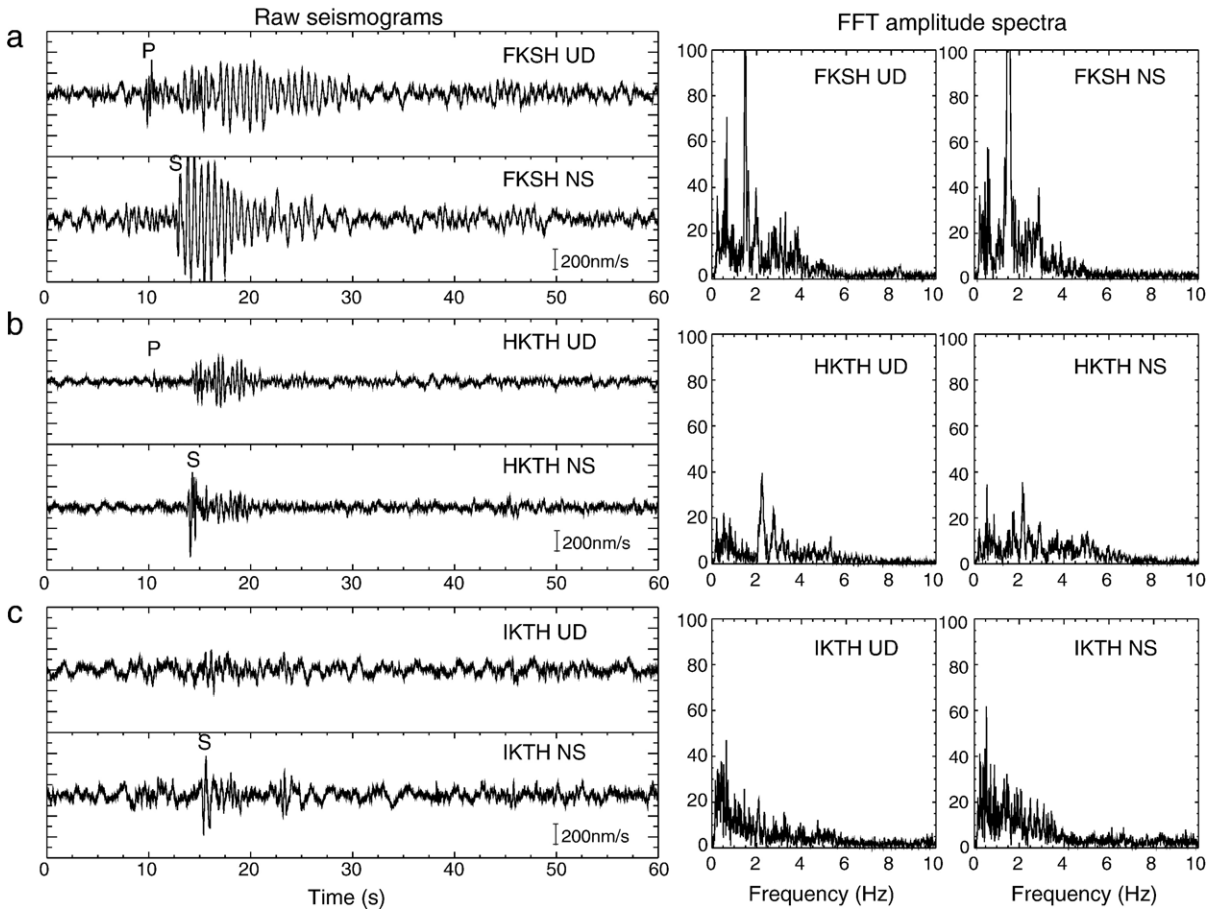

Fig. 3. Waveforms and fast Fourier transform (FFT) spectra of DLF earthquakes and low-frequency tremors in the subduction zone. (a) DLF earthquake near a volcano observed at station FKSH located in northeastern Japan. (b) DLF earthquake close to a fault system observed at station HKTH located in Chugoku, southwest Japan. (c) Deep low-frequency tremor observed at station IKTH in Shikoku, southwest Japan. P and S denote arrival of P and S waves.

these random wave trains, pulse-like signals with a frequency component of around 2 Hz occasionally appear. The amplitude of the tremor is usually very small, making it difficult to identify the tremor signal in the raw seismograms. However, the tremor is clearly identified in envelope seismograms which are root mean square (RMS) traces of bandpass-filtered outputs (2–16 Hz) as shown in Fig. 2(c). Envelope plotting aligns many traces in a compressed time window, enhancing the tremor signals. Coherent pattern of envelopes with small amplitude also appears in the envelope traces, indicating minor activity related to this tremor. Fig. 4 shows envelope seismograms for a period of three days. The envelope amplitude of the tremor signal changes very gradually, while the signal of ordinary earthquake exhibits a spiky shape. This feature indicates that the tremor is a continuous phenomenon with smoothly varying activity, in contrast to the instantaneous faulting of ordinary earthquakes.

The tremor is similar to DLF microearthquakes in terms of the frequency content, yet it exhibits a slightly different waveform pattern. The DLF microearthquakes that occur near volcanoes are usually characterized by relatively clear onsets of P and S waves, sometimes followed by monochromatic S coda waves with a predominant frequency of around 2 Hz (Fig. 3). The P phase sometimes includes relatively higher-frequency components compared with the S wave. These DLF events usually occur successively over a period of several minutes to hours, and sometimes persist continuously similar to a tremor. The waveforms of the DLF events that occur deep beneath the active faults are very similar to those of volcanic DLF events in terms of the relatively clear P and S phases and frequency content. The tremor waveform, on the other hand, does not exhibit distinct phases and reliable identification of P and S phases is difficult.

DLF events occur frequently in the vicinity of the source area of the 2000 Western Tottori Earthquake (M_w 6.7). The DLF activity is characterized by the successive occurrence of events. The most active DLF swarm occurred in this area on April 22, 2003. Fig. 5 shows envelope seismograms for a frequency range of 2–16 Hz prepared from seismograms observed at four stations in this area between April 22 and April 24. The DLF events occurred at around 9:40 on April 22, observed as a seismic vibration lasting for about 1 h with smoothly decreasing amplitude. Following this

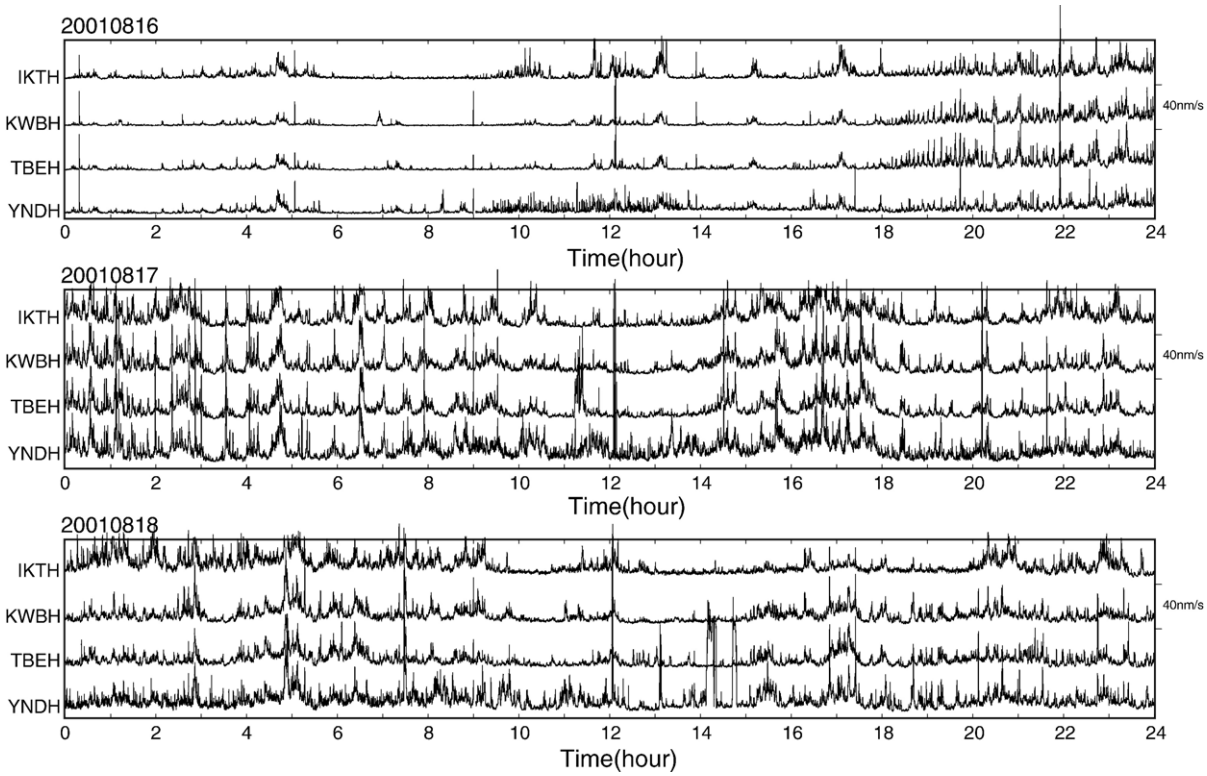

Fig. 4. Three-day envelope seismograms for 2–16 Hz observed at four stations in western Shikoku from August 16 to August 18, 2001. The tremor became active at around 18:00 on August 16 and lasted for several days. Pulse-like envelopes indicate normal earthquakes. Pulses at around 12:00 are due to nearby quarry blasting.

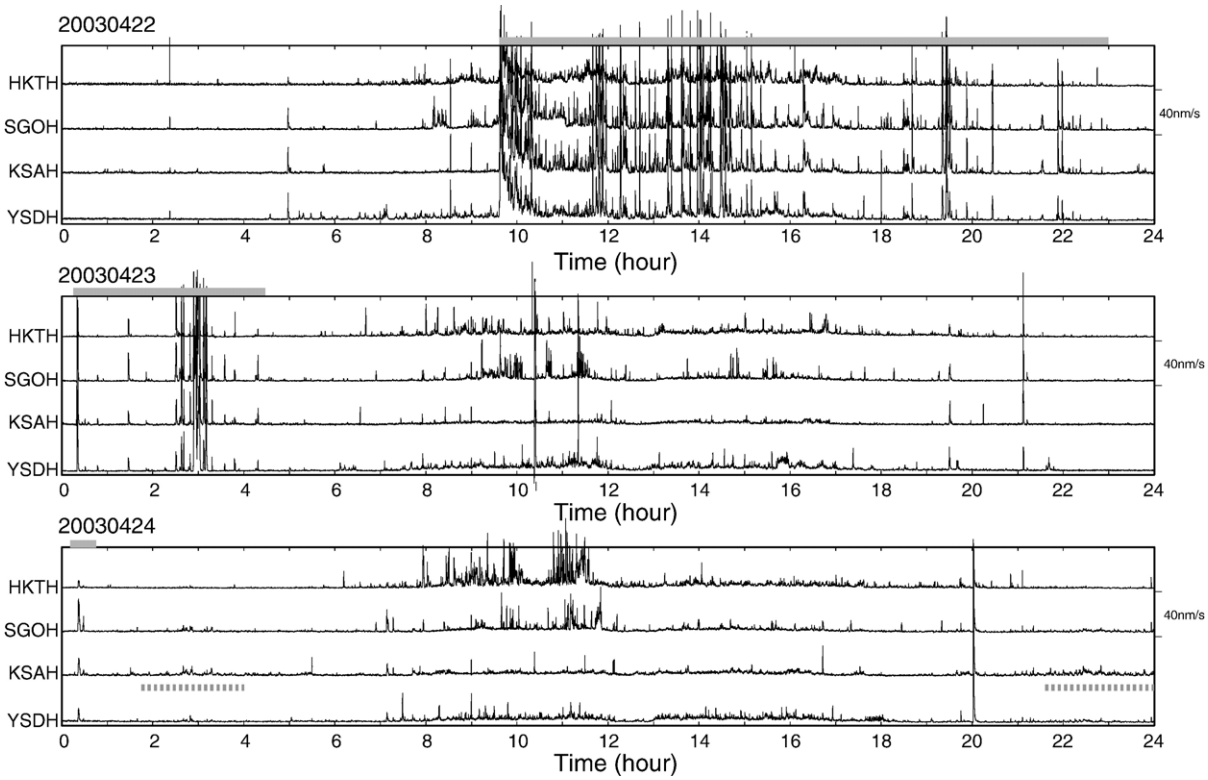

Fig. 5. Three-day envelope seismograms for 2–16 Hz observed at four stations around the source of the 2000 Western Tottori Earthquake from April 22 to April 24, 2003. DLF earthquakes occurred at around 9:40 on April 22, lasting for about 1 h with smoothly decreasing amplitude. Gray bold lines indicate the periods of DLF earthquake activity represented by pulse-like envelopes. Gray dashed lines denoted the correlated envelope pattern observed at 2:00–4:00 and 22:00–24:00 on April 24 due to a low-frequency deep tremor in Shikoku.

long vibration, a number of spike-like envelopes also due to DLF earthquakes appeared. This record, however, also includes two signals attributable to deep tremors that occurred in the Shikoku area, observed as weak signals recorded by multiple stations at 2:00–4:00 and 22:00–24:00 on April 24. The DLF events in Tottori appear as spiky noises in these records and more similar to ordinary earthquakes than the deep tremors.

3. Location of tremors

As no distinct P or S phases can be identified in the wave trains of these deep tremors, the onset times of these phases are very difficult to pick up and the source parameters cannot be calculated by the usual hypocenter determination method. However, on rare occasions, the random wave train includes a pulse-like phase with a predominant frequency of 1.5–2 Hz, as shown in Fig. 3(c). The Japan Meteorological Agency (JMA) measures the S onset time of such "low-frequency earthquake" (Nishide et al., 2000) based on the arrival of this pulse-like phase or the beginning of amplitude enhancement, allowing the source parameters to be calculated by the usual hypocenter determination method (Katsumata and Kamaya, 2003). However, during the most active stage of these tremors, it is impossible to locate the hypocenters by picking up the phases because the seismograms are too complicated.

As shown in Figs. 2 and 4, the envelope patterns recorded by neighboring stations are very similar. If the inhomogeneous structure has a scale length longer than the aperture of the stations, the seismic phases of distant events will commonly appear with the same pattern at neighboring stations. However, the shape of the coherent envelopes of tremors differs in each time period, and the amplitude of the coherent envelopes gradually decreases with distance among the local cluster of stations recording the highest amplitudes. This indicates that the tremor is a local seismic phenomenon, in which case the coherent amplitude change of envelopes will reflect the history of the tremor activity for a fixed source. Fig. 6 shows envelope seismograms observed at two stations in western Shikoku for a time window of 2 min. The envelopes are quite similar, with a time difference of only a few

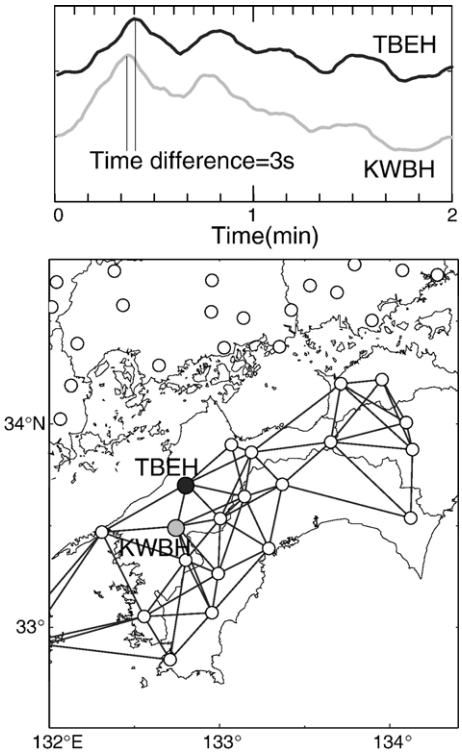

Fig. 6. Schematic of envelope correlation method used to determine the epicenter distribution of tremors. (Upper) Envelopes with duration of 2 min observed at stations TBEH and KWBH with a time difference of 3 s. (Lower) Pairs of stations (connected by lines) used in correlation analysis to calculate the time differences between stations.

seconds. This time difference is considered to represent the difference in distance from the source of the tremor. Using the correlation of envelope patterns, the time differences for many pairs of stations were calculated and averaged spatially based on the net-adjustment technique, which is often used in geodetic surveying. The apparent velocity of the envelope pattern is approximately 4 km/s in distant areas and faster than 4 km/s at stations near the tremor source. This suggests that the envelope pattern propagates as an S wave rather than a surface wave. The relative arrival time data are then used to locate the sources of tremors based on this assumption of S-wave propagation. This processing is carried out at 1 min intervals in the continuous record automatically. As this envelope correlation method for hypocenter determination yields a somewhat scattered pattern of hypocenters, the center of a distribution of tremors in each hour is selected as the epicenter of the tremor in that hour.

The epicenters of tremors determined by the envelope correlation method are plotted in Fig. 7. The tremors are distributed over a distance of 600 km from the Tokai area to the Bungo Channel, between the Kyushu and Shikoku islands, as pointed out previously (Obara, 2002b). The tremor epicenters are distributed parallel to the strike of the subducting Philippine Sea plate and are coincident with the depth contours of intra-slab earthquakes at depths of

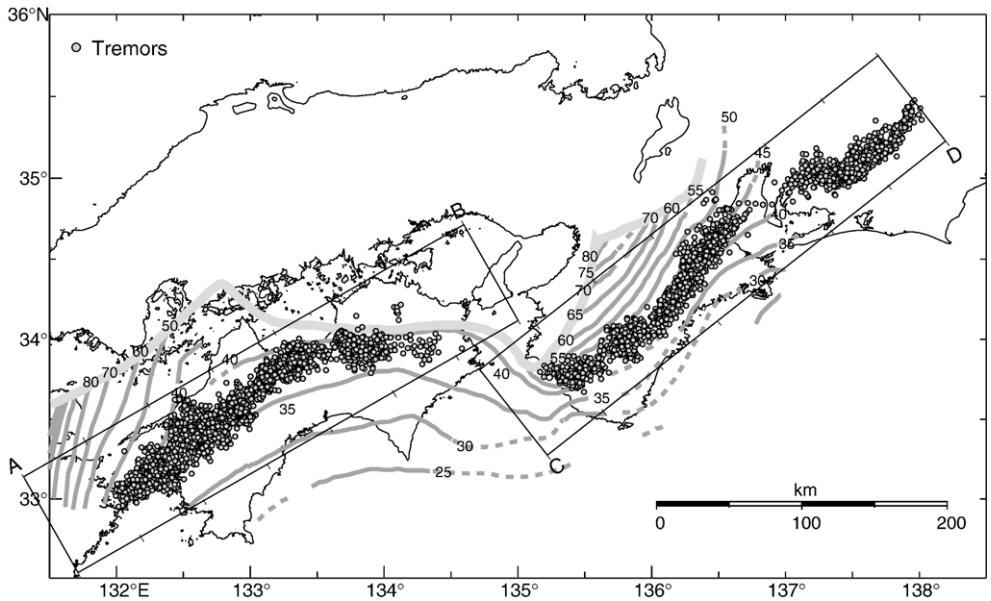

Fig. 7. Spatial distribution of tremors located by envelope correlation for the period from 2001 to 2003. Background depth contour lines indicate the depth of normal seismic activity inside the slab and the bold gray line indicates the leading edge of the slab (Nakamura et al., 1997). Boxes A–B and C–D with tick marks at 50 km intervals denote the areas used to count the number of tremors (see Fig. 8).

35–45 km (Nakamura et al., 1997). Various models have been proposed for the configuration of the Philippine Sea plate subducting beneath southwest Japan based on the reflection survey (Kurashimo et al., 2002), receiver function analysis (Shiomi et al., 2004), and seismicity in the subducting plate (Miyoshi and Ishibashi, 2004). All the models indicate that the plate boundary should be located 5–10 km above the depth contours of intra-slab seismicity. Considering the coupling state of the plate interface as estimated by Hyndman et al. (1997) based on the thermal structure, the distribution of tremors roughly corresponds to the transition zone at the deep extension of the locked zone on the plate interface. The source depth of the present deep tremors is approximately 30 km, which is above the intra-slab seismicity and roughly coincident with the subducting plate boundary. It should be noted that the depth determination includes a large degree of error as the analysis is based solely on S-wave arrivals. However, according to the JMA catalog, the low-frequency earthquakes corresponding to the tremors investigated in this paper are also located at depths of around 30 km.

When these deep tremors were first discovered (Obara, 2002b), no tremor activity was detected for the region of eastern Shikoku and the Kii Channel between Shikoku Island and Kii Peninsula. Recently, however, minor tremor activity has been detected in eastern Shikoku since new stations were deployed there although the amplitude of the tremor signal and duration of each event are much smaller than those detected in western Shikoku. Tremors have not been detected in the Kii Channel yet. Even if the tremors may exist there, they must be very small since major tremors are usually detected even in the stations with epicentral distances larger than 200 km. The belt-like distribution of tremor activity is approximately 30 km wide, expanding to 50 km in width in western Shikoku.

4. Time sequence of tremor

4.1. Periodic activity

These deep subduction-related tremors are distributed in clusters along the narrow belt parallel to the strike of the subducting Philippine Sea plate (Fig. 7). Fig. 8 shows a histogram of the tremor activity in the regions A–B and C–D indicated in Fig. 7, revealing regional peaks in tremor activity. In Shikoku Island, the tremor activity can be divided into eastern, central and western areas, and three areas of activity lie beneath the Kii Peninsula. Fig. 9 shows the sequence of tremor activity in the three areas of Shikoku. In western Shikoku, minor tremor activity occurs frequently, and major tremors lasting for longer than a few days occur at intervals of approximately six months. In central and eastern Shikoku, major tremors occur with a recurrence interval of only a few months. Neighboring clusters may become active simultaneously or successively. Thus, this belt-like active zone consists of a number of tremor clusters with individual periodicity.

4.2. Migration

The migration of tremors is an important feature. This phenomenon can be clearly observed in western Shikoku and the Bungo Channel area, which is the most active area in this belt-like distribution of tremors. Major tremor activity in this area occurs at intervals of approximately six months, and each episode of the major tremor activity is characterized by migration. Fig. 10 shows successive snapshots of the epicenter distribution of tremors for an episode that occurred in August and September, 2003. The epicenter distribution of tremors propagates from the Bungo Channel toward the northeast from August 27 to September 2 with an estimated migration velocity of 10 km/day. After September 3, tremors occurred intermittently over the en-

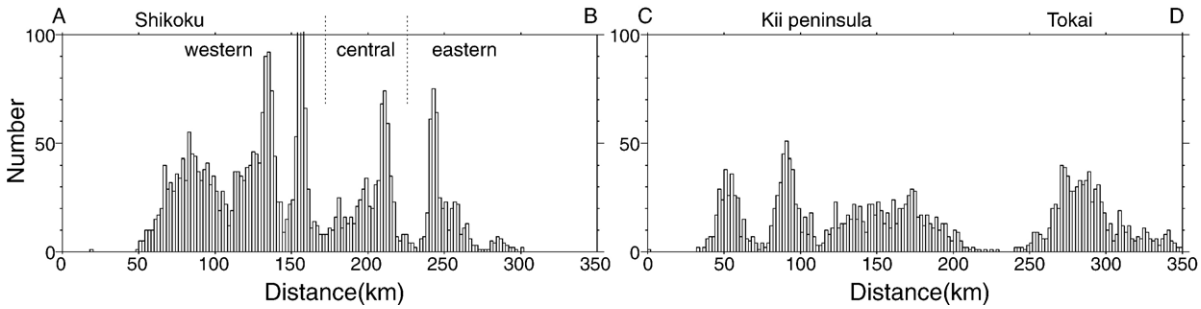

Fig. 8. Histogram of tremor activity in boxes A–B and C–D (see Fig. 7). The total number of located tremors is counted within a width of 2 km. The activity in Shikoku is divided into eastern, central and western parts, as indicated by dashed lines.

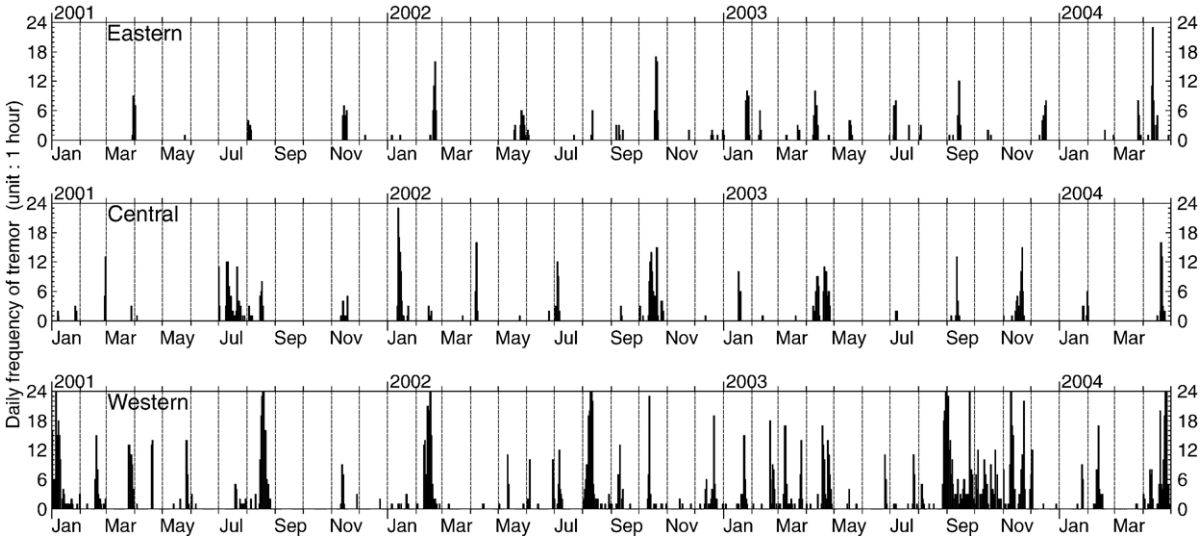

Fig. 9. Sequence of tremor activity in the, eastern, central and western parts of Shikoku. The daily frequency of located tremor activity counted with a unit of 1 h is plotted.

tire source area, 100 km along strike of the subducting plate. The migration velocity of this tremor episode is similar to that (8–13 km/day) reported by Obara (2002b), and comparable to the slow slip of 6–15 km/day detected in the Cascadia subduction zone (Dragert et al., 2001).

The migration phenomenon can be clearly observed in every episode of major tremor activity in western Shikoku. Fig. 11 shows the epicenter distribution of tremors detected in four major tremor episodes in 2001 and 2002, where the epicenter distribution for each episode is divided into two stages. In the episodes starting in January 2001 and February 2002, the tremor activity begins in the northeast of the tremor zone and migrates to the southwest. In the August 2001 and August 2002 episodes, on the other hand, tremors began near the western coastline of Shikoku Island and then propagated both northeast and southwest. However, the activity clearly increases in the northeast in the second stage, indicating that the major activity primarily migrates from southwest to northeast in the August episodes. Over this two-year period, the migration direction appears to alternate. This type of migration behavior is also recognized in the northern Kii Peninsula, as discussed later.

5. Slow slip and tremor activity

5.1. Western Shikoku

Each Hi-net station is equipped with a two-component tiltmeter besides a three-component velocity-type

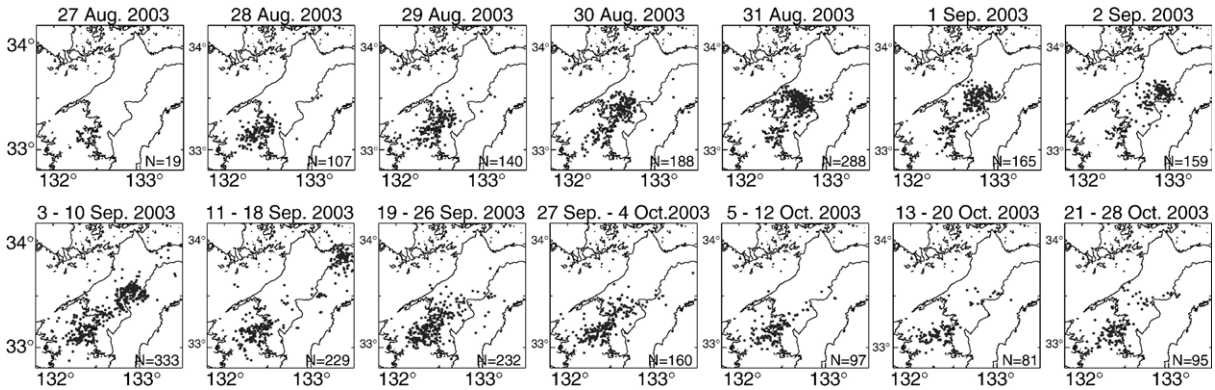

Fig. 10. Migration of tremor activity during an episode in the period from August 27 to September 2, 2003. (Upper) Epicenter distribution of tremors determined by automatic process of the envelope correlation method for each day. (Lower) Epicenter distribution of tremors in eight day intervals from September 3 to October 28. Only well-determined epicenters with errors of less than 1 km are plotted.

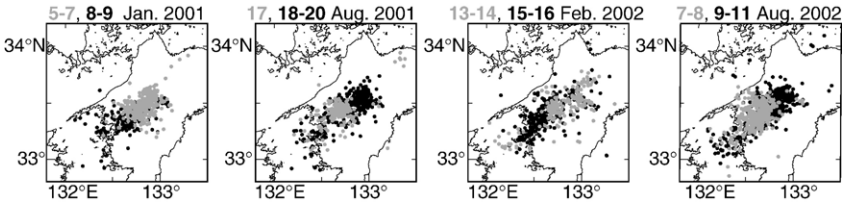

Fig. 11. Migration of tremor activity during four episodes: January and August, 2001, and February and August, 2002. The epicenters in the first half of each episode are plotted in gray, and those in the latter half are plotted in black.

seismometer in an anti-pressure capsule at the base of a borehole (Okada et al., 2004). The direct current component of the instrument can be treated as crustal tilt movement. The tilt data are processed with Baytap-G (Tamura et al., 1991) to remove the effect of the earth tide. The outputs are compared with the tremor activity in Fig. 12 for station HIYH, western Shikoku. Simultaneously with the four major tremor episodes in 2001 and 2002, step-like tilt changes were observed at some stations in this region, as indicated by arrows in Fig. 12. In fact these tilt changes have time constants of several days as shown in upper panels of Fig. 13, which shows the expanded view of tilt changes at the station HIYH with the daily frequency of the tremor activity. Such tilt steps were observed simultaneously at some stations, with a time constant of several days. No crustal deformation corresponding to these tilt changes were detected by the GEONET (Sagiya, 2004) GPS monitoring system of the Geographical Survey Institute (GSI) of Japan, indicating that the tilt change are caused by low-magnitude crustal deformation over a short period.

The four tilting episodes examined above can be classified into winter and summer groups according to the detailed history of tilt change. For the winter group, the episodes in January 2001 and February 2002, the N–S component of tilt change started three days in advance of the change in the E–W component. For the summer group, the episodes in August 2001 and August 2002, both components of tilt change started simultaneously, and the change in the N–S component persisted for a few days after the change in the E–W component has ceased. This difference in tilting pattern between the winter and summer groups is examined in the lower panels of Fig. 13, which shows the trajectory of tilt vectors for these four major episodes. The tilt movement progressed from south-down tilting to south-east-down tilting in the winter group, and from south-east-down to south-down in the summer group.

For analysis of crustal deformation data, the effects of atmospheric pressure and railfall need to be considered, as these effects sometimes contaminate the tilt output. Unfortunately, no meteorological measurements are made by the Hi-net stations. However, the fluctuation in the tilt signal due to the pressure change can be roughly estimated from other variations in the tilt. For example, a short-period fluctuation can be observed in the data over four days from August 19 to August 22, 2001, which can be correlated with a large typhoon that struck the Kii Peninsula in this period. The step-like tilt change due to slow slip, however, is much larger than that induced by this major depression of pressure,

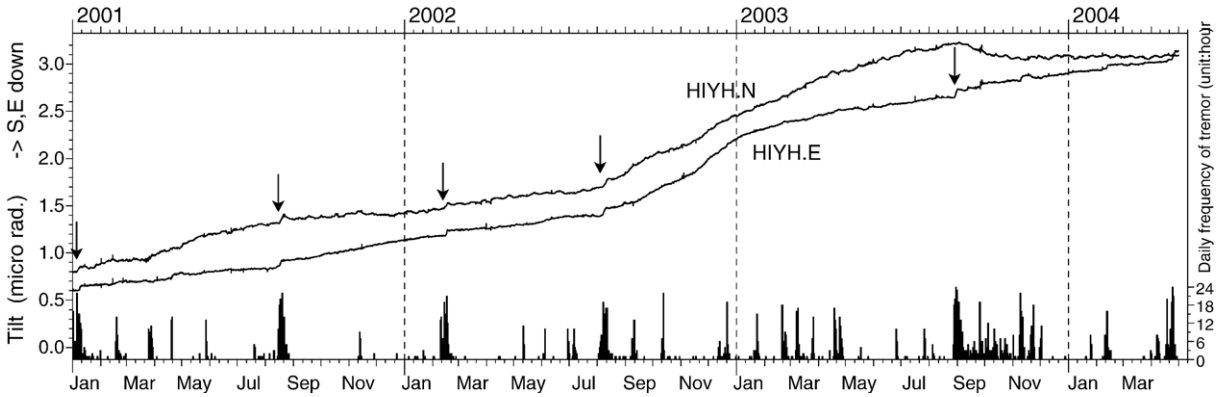

Fig. 12. Sequence of tremor activity around the Bungo Channel and western Shikoku, and tilt changes observed at station HIYH. Tilt data are corrected for the tidal component using Baytap-G. The daily frequency of tremors counted with a unit of 1 h for the western area of Shikoku is plotted for reference. The step-like tilt change is coincident with the major tremor activity indicated by arrows.

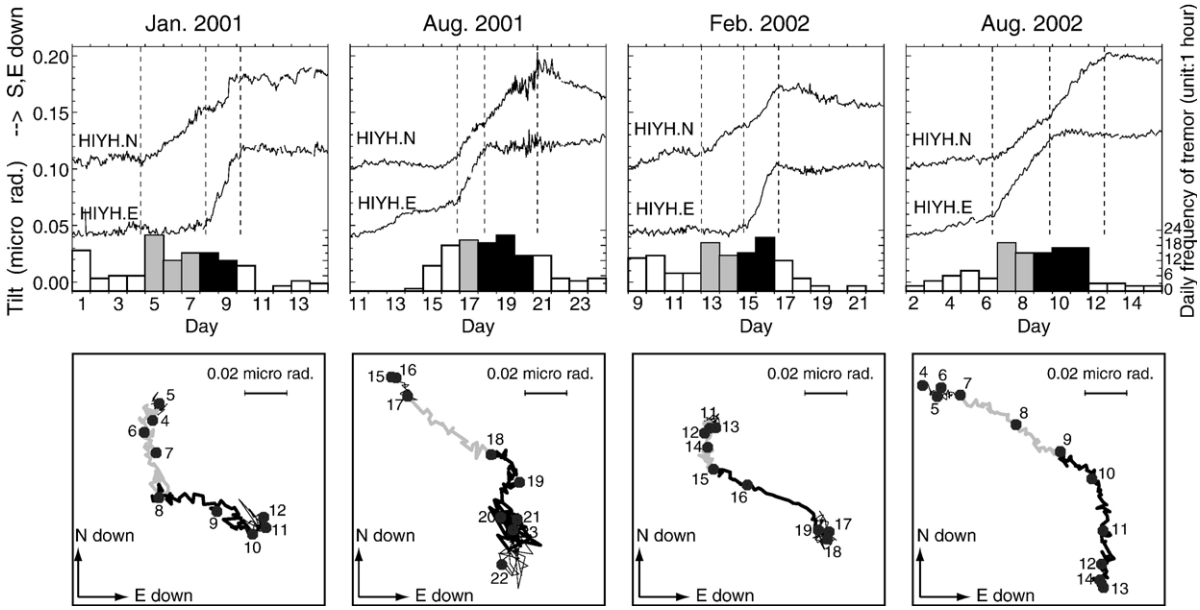

Fig. 13. (Upper) Detailed view of tilt change observed at station HIYH and tremor activity in western Shikoku over a period of 14 days. Dashed lines indicate the turning point of tilt change. The daily frequency of tremors is plotted at the bottom of the panel. Gray and black bars represent the first and second stages of each episode, respectively, corresponding to the period shown in Fig. 11. (Lower) Trajectory of tilt vectors for major episodes. The data measured at 0:00 of each day is plotted in black and labeled with the date. Gray and black lines denote the trajectory of tilt vectors in the first and second stages, respectively.

indicating that the effect of atmospheric pressure is negligible in this analysis.

The detailed patterns of tilt changes are closely related to the migration of tremor activity. As shown in Fig. 11, the tremor activity migrated from northeast to southwest in winter, and from southwest to northeast in summer. Comparing the tilt data observed at station HIYH with the migration of tremors, a south-down tilt occurs when the tremors are distributed in the northeast, and southeast-down tilt occurs when tremors are in the southwest. This indicates that the source location of the crustal deformation propagates with the migration of tremors. As shown in Fig. 13, the tremor activity usually precedes tilt change.

The most plausible source model is the occurrence of a slow slip event at the subducting plate boundary, similar to that at the Cascadia margin (Dragert et al., 2001). Using tilt data observed in western Shikoku (Obara et al., 2004), the slow slip fault model for the episode in August 2002 was estimated as follows. As the tilt vectors changed direction on August 10, 2002 (Fig. 13), the parameters of the slow slip model were estimated for two stages: August 6–9 and August 10–12. The strike and dip slips were calculated by a least-squares method using Okada's (1992) formula, and the fault geometry was optimized by genetic algorithm using a depth range of 30–45 km, a strike of 190–260°, and a dip of 0–45°. The estimated depth and bootstrap error is 40 ± 3 km, with strike of $226° \pm 7°$ and dip of $30° \pm 6°$.

The estimated fault geometries in the two stages are plotted in Fig. 14. The depth, strike and dip angles of the fault in the first stage were estimated considering the plate configuration, and the same parameters were used for the second stage. The dislocations in the first and second stages were estimated to be 3 and 0.7 cm, respectively. The fault is estimated to be a reverse fault, which is coincident with the plate subduction. The observed data can be explained by successive slips on two adjacent fault planes on reverse faults located in the tremor area over a horizontal range of 100 km, with a corresponding total moment magnitude (M_w) of approximately 6.0.

The estimated depth to the shallower edge of the fault is 40 km, lying just above the dipping intra-slab seismicity. Considering the depth error, the slow slip fault can be regarded as coincident with the plate boundary. The updip limit of the slip roughly corresponds to the source area of tremors, while the bottom edge corresponds to the transition zone on the subducting plate interface (Hyndman et al., 1997). The down-dip limit of the slow slip event may correspond to the triplication of the subducting oceanic crust, the continental lower crust and the wedge mantle according to

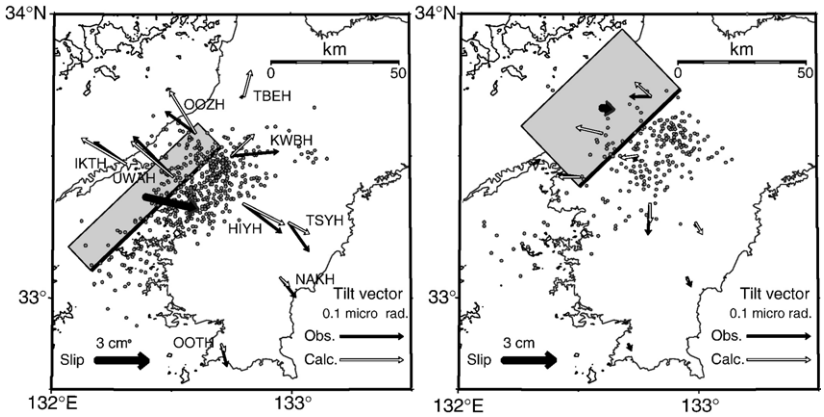

Fig. 14. Distribution of tilt vectors and fault geometry of slow slip events for two successive stages of the episode in August 2002. The epicenter distribution for these tremors is also plotted.

guided wave analysis (Ohkura, 2000). This would indicate that the slow slip event takes place on the interface between the subducting oceanic crust and the continental lower crust. The displacement expected from the estimated slow slip model is less than 2 mm on the ground surface, which would be very difficult to detect by the GPS observation network.

5.2. Eastern Shikoku

In eastern Shikoku, the major tremor activity repeats at intervals of two or three months, as shown in Fig. 9. The periodic tremor activity is sometimes accompanied by small tilt changes that are detected by a station close to the tremor source area. Fig. 15 shows the tilt change observed at station SINH for comparison with the tremor activity in eastern Shikoku (Fig. 8). Although the component due to earth tide is still remained in the data, a small E–W tilt can be seen to occur at the same time as the tremor activity. The tilt is always west-down movement of 0.02–0.05 μrad. The smaller tilt step and shorter duration of each episode compared to the phenomena in western Shikoku suggests that the slow slip events below eastern Shikoku are much smaller. The

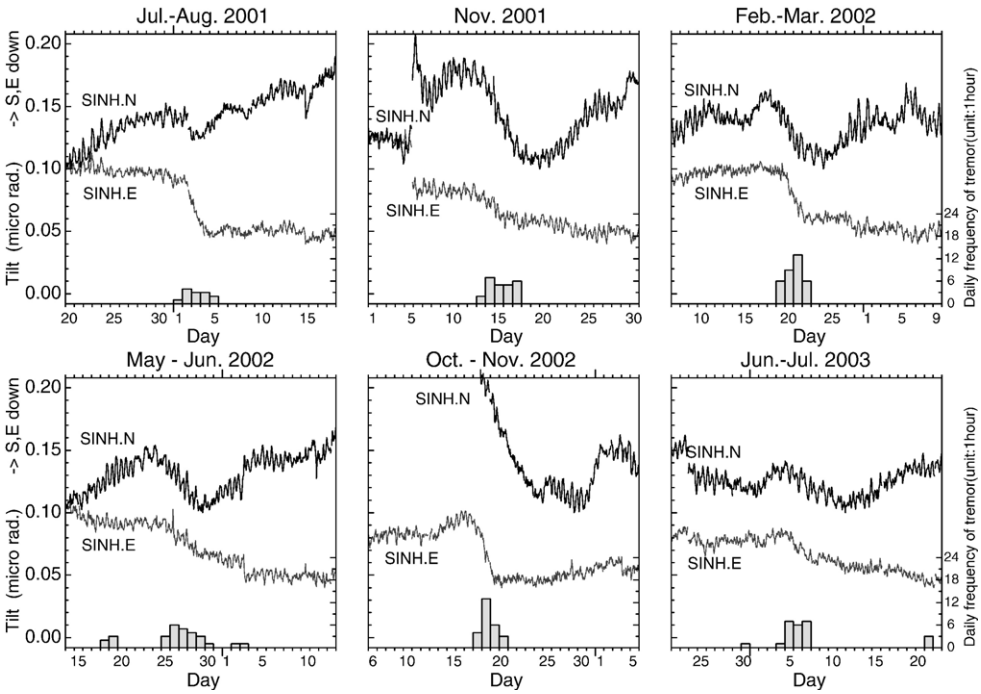

Fig. 15. Detailed view of tilt change at station SINH and tremor activity in eastern Shikoku over a period of 30 days.

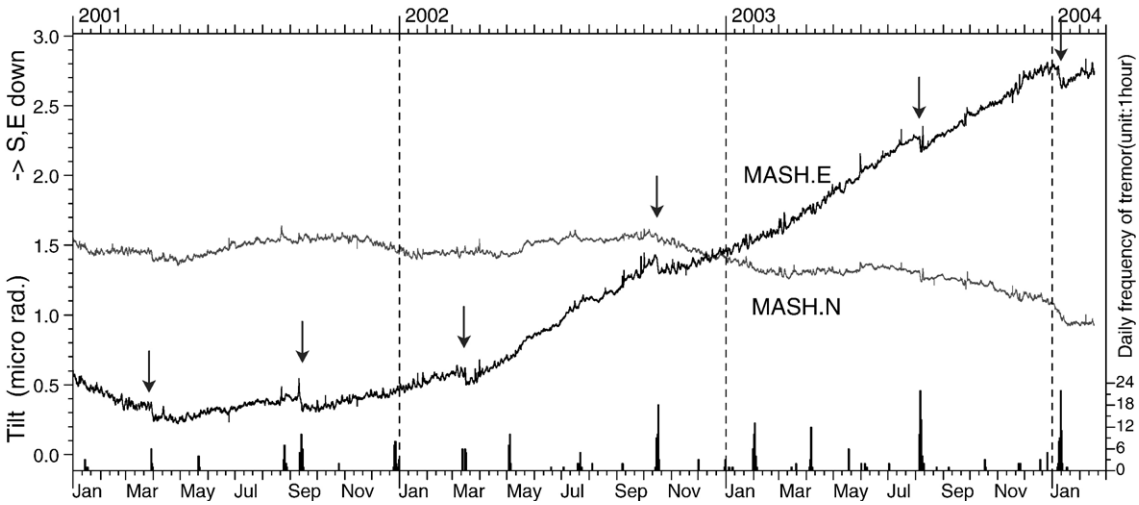

Fig. 16. Sequence of tremor activity in the northern Kii Peninsula and tilt change observed at station MASH. The daily frequency of tremors counted with a unit of 1 h located in the square region in lower-right of Fig. 1 is also shown.

recurrence time may therefore be proportional to the size of the slow slip event, which may indicate that the friction properties may differ between tremor clusters on the same subducting plate.

5.3. Northern Kii Peninsula

Three tremor clusters can be identified below the Kii Peninsula, as shown in Fig. 8. From a comparison of the sequence of tremor activity for each cluster with the tilt data recorded by neighboring stations, the tremor activity in the northern Kii Peninsula appears to be related to the crustal tilting observed at some Hi-net stations. Fig. 16 shows the tremor activity in the northern Kii Peninsula and crustal tilt change observed at station MASH for three years of 2001–2004. In this area, the tremor activity repeats at intervals of approximately six months, with some fluctuations. There are six step-like tilt changes accompanying tremor activity with a time constant of several days, as shown in Fig. 17. The tilt record is sometimes contaminated by atmospheric pressure changes, as indicated by the coherent patterns in both tilt components. For example, sharp coherent pulses appeared on August 8–9, 2003 and January 12–13, 2004, attributable to a typhoon and large storm with great low atmospheric pressure. These tilt changes due to atmospheric pressure variations quickly return to pre-storm levels. However, the tilt change associated with the tremor activity is large compared with such pressure-induced noise, and involves a permanent baseline shift (gray lines in Fig. 17). The tilt at station MASH coincident with major tremor activity is always of northwest-down sense, and tilting is always accompanied by major tremor activity. However, there are some tremor events that do not produce a tilt response, as shown in the lower-right corner of Fig. 17.

The tilt can also be observed at a few neighboring stations, indicating that the crustal tilt is caused by a deep, and slow slip event. The tremor activity associated with the episode in January 2004 migrated from northeast to southwest, as shown in Fig. 17. In this case, the migration velocity was roughly 10 km/day, the same as that observed in western Shikoku, suggesting that migrating tremor and slip phenomena are common features. As the crustal tilt changes in the northern Kii Peninsula and eastern Shikoku are detected at only one or a few stations, the fault model for the slow slip event cannot be estimated for these areas. However, the occurrence of episodic slow slip associated with major tremor activity remains an observational fact for these regions.

5.4. Long-term slow slip event in the Bungo Channel and Tokai

In the latter half of 2003, the slow slip and tremor activity below western Shikoku became very complex. As shown in Fig. 9, the tremor activity lasted for three months, from the end of August to the beginning of December in western Shikoku and the Bungo Channel. Major migrating tremor activity occurred initially, with an associated short-term slow slip, but thereafter appeared intermittently in the Bungo Channel, as shown in Fig. 10. The tremor activity from August to December 2003 is the longest recorded since the mon-

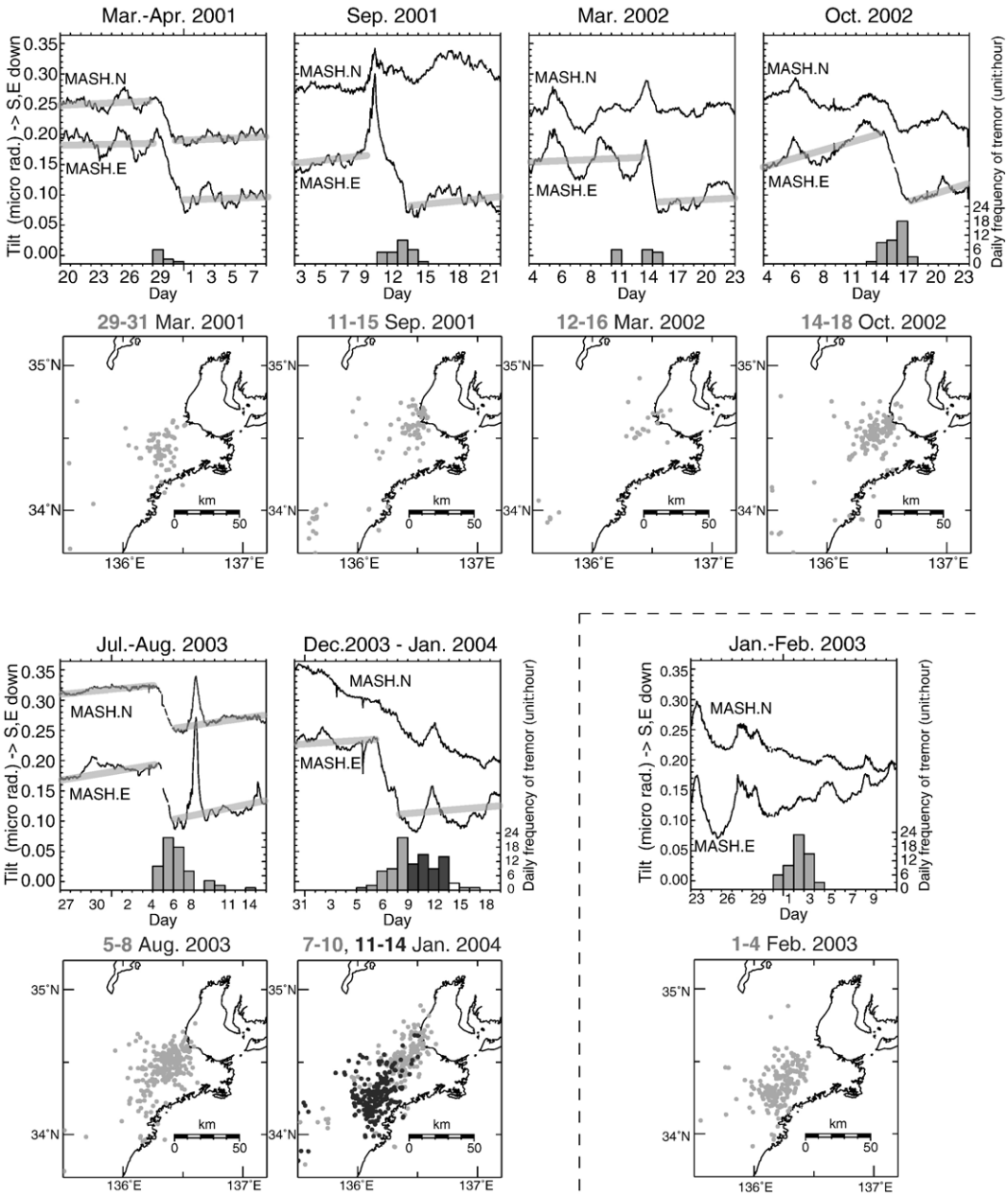

Fig. 17. Detailed view of tilt change and tremor activity for the six episodes shown in Fig. 16 and tremor activity without corresponding tilting events. (Upper) Tilt change plotted over a period of 20 days. (Lower) Epicenter distribution of tremors plotted in first (gray) and second (black) stages.

itoring of tremor activity in southwest Japan began in 2000 (Obara, 2002b), and is significantly longer than the usual duration of one week.

In the same period, a long-term crustal deformation was detected by a number of tiltmeters and GPS monitoring by GSI (Ozawa et al., 2004). The crustal movements continued from August to November, 2003, and are considered to be due to a slow slip event in the Bungo Channel area, as shown in Fig. 18. Hirose and Obara (2005) estimated the fault model for this slow slip event by combing the tilt and GPS data.

The short-term slow slip event is strongly correlated with the major migrating tremor activity, and the long-term slow slip event is similarly accompanied by tremor activity, which occurred intermittently over the period of the slip event. This long-term slow slip event is very similar to the slip recorded from 1996 to 1997 (Hirose

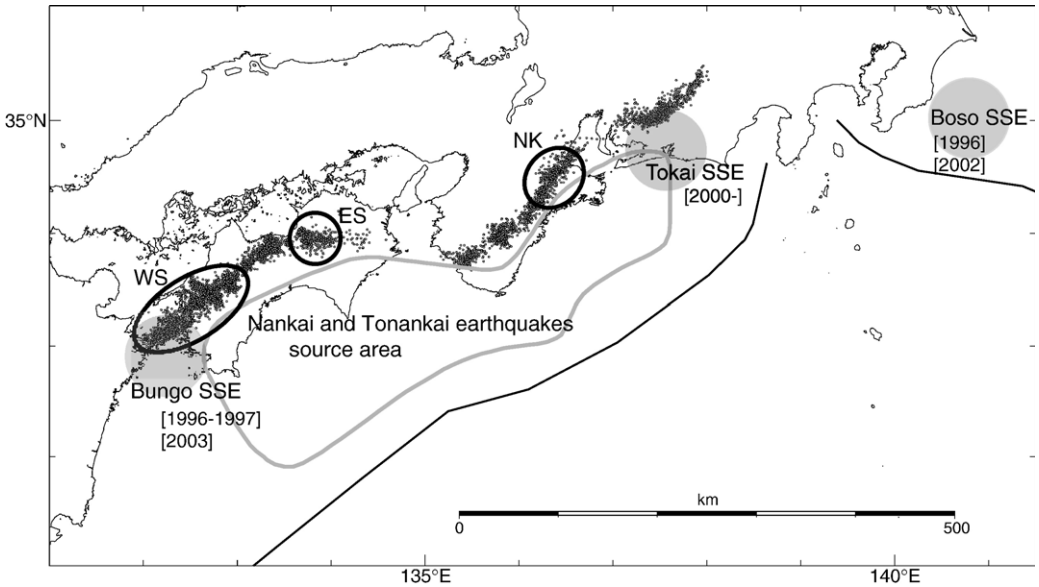

Fig. 18. Spatial distribution of slow slip patches in the tremor source area. Bold circles indicate source areas of short-term slow slip events associated with major tremor activity. The short-term slow slips in western Shikoku (WS), eastern Shikoku (ES) and northern Kii Peninsula (NK) are investigated in this study. Shaded circles represent slow slip events (SSE) detected by GPS monitoring. The estimated source areas of the next major Nankai and Tonankai earthquakes are also plotted (Headquarters for Earthquake Research Promotion, 2001). The Bungo SSE occurred in 1996 and 1997 (Hirose et al., 1999) and again in late 2003 (Ozawa et al., 2004; Hirose and Obara, 2005). The Tokai SSE is an ongoing event that began in 2000 (Ozawa et al., 2002). The Boso SSE occurred in 1996 and 2002 (Ozawa et al., 2003).

et al., 1999), although no information regarding the occurrence of tremor activity in that period is available because it was prior to the deployment of the dense seismic network.

In Tokai, central Japan, a slow slip event detected by GPS measurement is ongoing since 2000 (Ozawa et al., 2002). The estimated source area of the Tokai slow slip event is located on the updip side of the tremor zone on the plate boundary as shown in Fig. 18. At present, there is no clear correlation between the crustal deformation and tremor activity in the Tokai area, but based on the present discussion it would appear likely that the phenomena are related.

6. Discussion

6.1. Comparison between the tremor and DLF microearthquake

There are some clear differences and similarities between the deep low-frequency tremors that occur along the strike of the subducting Philippine Sea plate and inland DLF microearthquakes in the vicinity of volcanoes and active faults. Both events occur at a depth of around 30 km, indicating that both occur in similar pressure environments. Furthermore, the seismic signals have almost the same frequency content, with a predominant frequency of around 2 Hz, lower than that of ordinary earthquakes. If these signals are radiated by shear faulting, the low-frequency property may indicate a low stress drop or a low dislocation velocity on the crack surface due to the presence of gauge including fluid and other material. The basic mechanism for the low-frequency event is assumed to be crack dislocation in the presence of fluid under high temperature and pressure at a depth of around 30 km.

On the basis of this assumption, differences in the amount of fluid and the degree of crack development may be the key difference between these low-frequency events. At the Moho discontinuity on the backarc side, cracks and fluid are considered to exist only around volcanoes and fault systems as singular features. Therefore, such low-frequency events will occur only in limited areas around these singular points inside the island arc. In the tremor source area, on the other hand, it is expected that cracks will be well developed due to the underplating or underthrusting along the subducting plate interface and the metamorphism associated with the subducting oceanic crust (Kimura, 2002). Dehydration of the subducting oceanic plate will also release large volumes of fluid, which in northern Japan produces magma upwelling via volcanism, but in southwest Japan remains trapped due to differences in the thermal structure. The fluid liberated from the descend-

ing oceanic plate is therefore considered to accumulate at the plate interface without modifying the composition of the surrounding rock. This difference in the volume of available fluid and the crack density may be responsible for the different durations and periodicity of activity for tremors and DLF microearthquakes. The large volumes of fluid and high crack density are distributed along the entire plate interface, allowing tremors to occur over a large extent along the strike of the subducting Philippine Sea plate. The regions along the strike of the subducting plate without significant activity may represent areas of anomalous structure, as pointed out by Seno and Yamasaki (2003).

6.2. Relationship between deep tremors and slow slips

The major deep tremor activity and the slow slip events occur simultaneously in western Shikoku, whereas no crustal deformation is detected in association with minor tremor activity. This presents two possibilities for the relationship between the tremors and slow slip. The first is that the tremors and slow slip are identical geophysical phenomena, that is, each crack dislocation causes a micro slip and radiates low-frequency signals. In this case, superposition of seismic signals from each crack gives rise to the observed sequence of tremors, and the integral of micro slips produces the observed slow slip event. Thus, even minor tremor activity should be accompanied by small slow slip, which may not be detectable. The occurrence of minor tremor activity immediately prior to the onset of slow slip was observed for four major tremor and slip episodes. According to this model, the initial tremors would have been accompanied by micro slips that could not be detected with the present instrumentation.

The alternative possibility is that the tremor and slow slip are strongly related but occur by independent physical mechanism, in which case the source locations of the two event types will differ. It is still assumed that the tremor signal is radiated from micro cracks in an inhomogeneous layer close to the plate boundary, and that the slow slip occurs at the plate boundary. As shown in Figs. 12 and 16, tremors sometimes occur in otherwise quiescent periods of slow slip activity, triggered instead by local microearthquakes or teleseismic waves (Obara, 2003). Although the physical mechanism for this triggering remains unclear, the existence of such triggering phenomenon suggests that the tremors are very sensitive to the source environment. The slow slip event may represent one form of trigger: once a slip event occurs, instability in the tremor source zone increases and crack events occur continuously in a chain-like manner. In this case, the fluid liberated from the subducting oceanic crust will play an important role in promoting both slow slip and tremors through hydraulic fracturing. According to Nishiyama (1989), vein formation in metamorphic rocks results from the hydro-fracturing. The observed tremor may therefore represent chain-like hydro-fracturing in the ongoing metamorphic process at depth. The fluid-driven crack model suggested as the mechanism for volcanic tremors may also be applicable for the deep tremors.

Clarification of the relationship between tremors and slow slip will require more precise determination of hypocenter distribution of tremors and the fault geometry of slow slip events. However, the simultaneous occurrence of tremors and slow slip indicates that these phenomena are strongly related with each other and with the subduction of the oceanic plate.

6.3. Comparison of episodic tremor and slip in Cascadia and southwest Japan

In the Cascadia subduction zone, slow slip events occur repeatedly with a period of 13–16 months (Miller et al., 2002) and are always accompanied by deep seismic tremors (Rogers and Dragert, 2003). The characteristics of ETS in Cascadia are very similar to those detected in western Shikoku with differences in the recurrence interval and duration of each episode. These coupled phenomena occur at the deepest extent of the transition zone on the subducting plate boundary in both subduction zones, which are characterized by the subduction of young, warm oceanic plates. These observations indicate that the source area of the episodic tremor and slip is usually locked, but slides with a certain recurrence interval. The difference in the recurrence interval between Cascadia and southwest Japan may depend on specific conditions on the plate boundary.

In eastern Shikoku, major tremor activity repeats at intervals of two or three months. At a station close to the tremor source area, minor tilt changes lasting for a few days are sometimes observed in association with these tremor episodes. This suggests that the boundary conditions may also differ between tremor clusters on the same subducting plate. Another possibility is that difference in the fault size of the slow slip may control the recurrence interval even if the frictional properties are the same in both regions.

The slow slip events in southwest Japan, however, can last for much longer. In the Tokai area, near the eastern end of the belt-like tremor zone, a slow slip event first detected in 2000 continues today (Ozawa et

al., 2002). In the Bungo Channel, near the western end of the tremor zone, long-term slow slip events were recorded in 1996 and 1997 by GPS monitoring (Hirose et al., 1999) and at the end of 2003 by GPS and tiltmeters (Hirose and Obara, 2005). The source depth of both of these long-term slow slip events is slightly shallower than that of tremors or that of the short-term slow slip events accompanying tremor activities. As the coupling at the plate interface varies from a locked zone to free slip with increasing depth and temperature (Hyndman et al., 1997), the duration of the slip events may reflect differences in the friction properties at the plate interface.

Slow slip events are also detected off the Boso Peninsula, central Japan (Ozawa et al., 2003), as shown in Fig. 18. The recurrence interval of the Boso slow slip is about six or seven years, and each episode lasts for a few weeks. The slow slip events are always accompanied by a swarm of microearthquakes, and the source of the slow slip migrates (Kimura et al., submitted for publication). The time constant of this slow slip is similar to that for Cascadia and western Shikoku, although the accompanying seismic phenomena differ.

The tremor is thought to be related to fluid liberated from the subducting plate by dehydration, which would alter the friction properties at the plate interface, resulting in a slow slip event. These tremors and slow slip events are thus thought to represent the conditions at the plate boundary and stress accumulation in the locked zone directly. The continuous monitoring of these phenomena may therefore prove very useful for modeling and monitoring the loading process in these subduction zones. The present findings suggest that the episodic tremor and slow slip are due to instability and friction reduction at the interface caused by the steady accumulation of fluid in the inhomogeneous zone near the plate boundary.

6.4. Slow slip patch in the tremor zone

The tremor source area extends for 600 km along the strike of the subducting plate, with gaps in Ise Bay between Tokai and the Kii Peninsula, and in the Kii Channel. The tremors are also divided into clusters along this belt-like distribution, with each cluster having a characteristic periodicity. Some of the periodic clusters are accompanied by slow slip events. Although a fault model could only be estimated for the slow slip in western Shikoku, where tilt stations detected the crustal deformation, other short-term slow slip events also occur in eastern Shikoku and northern Kii Peninsula. The short-term slow slip events in these three areas correspond well to the tremor clusters shown in Fig. 18. Slow slip events have not yet to be detected in other areas exhibiting deep tremors, however, it is very important to keep monitoring tremors and slow slips in order to clarify common features and regional differences.

7. Summary

The deep low-frequency tremors detected along the strike of the subducting Philippine Sea plate in southwest Japan were investigated and compared with similar DLF events that occur near the Moho discontinuity in the vicinity of volcanoes and active fault systems in the Japan island arc. Although similarities were identified between the deep tremors and DLF events in terms of source depth and frequency content of the seismograms, the subduction tremor is considered to be a large-scale phenomenon, both temporally and spatially. The belt-like distribution of tremors consists of many tremor clusters, each with a characteristic periodicity of activity. Episodic slow slip events are well correlated with this periodic tremor activity in western Shikoku, southwest Japan. The recurrence period of activity in this area is approximately six months, and the duration of each episode ranges from a few days to a week. The source of the tremor and slow slip migrates along the strike of the subducting plate. The fault model determined for one slow slip event suggests that the slow slip occurred close to the tremor source area, adjacent to the plate interface. In eastern Shikoku and northern Kii Peninsula, slow slip events have also been detected during nearby tremor activity. These tremor and slow slip phenomena are very similar to the episodic tremor and slip detected in the Cascadia subduction zone. Along the subducting Philippine Sea plate, long-term slow slip events have been detected by GPS monitoring in Boso, Tokai and the Bungo Channel. However, while the long-term slow slip event in the Bungo Channel is strongly correlated with tremor activity, the relationship between these slow slip events and coupling seismic phenomena is different for each case. It is considered that differences in the condition on the plate interface may affect the behavior of plate slip and accompanying seismic phenomena.

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