Repeating deep tremors on the plate interface beneath Kyushu, southwest Japan

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In the subduction zone south of Kyushu Island, at the western extension of the Nankai subduction zone, southwest Japan, the age of the oceanic crust increases toward the south across the subducting Kyushu-Palau ridge. While tremor activity is very high in Nankai, tectonic tremors have only recently been discovered in Kyushu. In this study, we examined tremors beneath Kyushu using an improved version of the envelope correlation method. In doing so, we distinguished tremors from normal earthquakes and background noise using the criteria of source duration and the spectrum ratio between low and high frequencies. Accurate measurement of *S-P* times, using cross-correlation between vertical and horizontal seismograms, constrains the tremor depth precisely. Tremor activity is low and within a small region in southern Kyushu, where thick crust of the Kyushu-Palau ridge is being subducted, at depths between 35 and 45 km (i.e., shallower than intra-slab earthquakes by about 20 km), which is consistent with the location of the plate interface within uncertainties proposed in previous studies. Establishing precise depth estimates for tectonic tremors beneath Kyushu, which results from shear slip along the plate interface, is useful in defining the plate interface within the Nankai subduction zone.

Key words: Tectonic tremor, Kyushu, subduction, *S-P* times, cross-correlation, plate interface.

1. Introduction and Tectonic Setting

In the Nankai subduction zone, western Japan, the Philippine Sea (PHS) plate is subducting beneath the Eurasian (or Amurian) plate at a relative plate velocity of 6–7 cm/yr (Wei and Seno, 1998). At this location, the age of the subducting PHS plate is 30–17 Ma. Farther to the west, the Nankai subduction zone continues off to the south of Kyushu Island (Fig. 1(a)), where the age of the subducting plate is 60–40 Ma. The subducting Kyushu-Palau ridge (KPR) (Park *et al.*, 2009) (Fig. 1(a)) forms a tectonic boundary within the subducting plate that comprises a remnant arc, which is relatively buoyant and less dehydrating than the surrounding oceanic plate.

Tectonic tremors within the Nankai subduction zone were first documented by Obara (2002), and they are often accompanied by slow slip events (SSEs) and considered to represent shear slip within the transition zone between stick-slip and stable slip along the plate interface, based on their locations and focal mechanisms (e.g., Shelly *et al.*, 2006, 2007; Ide *et al.*, 2007). Fluids released by the dehydration of a subducting slab are thought to occur at high pore pressures in areas of tremor sources (Shelly *et al.*, 2006), thereby reducing the effective normal stress and enhancing the sensitivity of tremor activity to tidal stresses and large surface waves. For recent reviews of tremor characteristics, see Rubinstein *et al.* (2010), Obara (2011), and Beroza and Ide (2011).

The distribution of tremor in the Nankai subduction zone terminates at the Bungo channel, across which several characteristics of the subduction zone change dramatically, including seismicity, volcanic activity, and the geometry and age of the subducting plate. The relatively young and warm PHS plate within the Nankai subduction zone must exist at the required conditions to enable tremor activity, because both slip transition and active dehydration in the subducting slab occur at shallower regions of the plate interface. In contrast, the occurrence of tremor activity is not favored by the relatively old plate being subducting south of Kyushu. Nevertheless, the Japan Meteorological Agency (JMA) detected and located two low-frequency earthquakes (LFEs) at a forearc region in the southern Kyushu area (Fig. 1(b)) during a routine analysis, implying the existence of tectonic tremor activity in the area. Ide (2012) reported such tremor activity in this region. The main focus of the present study is to conduct a more detailed investigation of tremor activity in the southern Kyushu area and to accurately constrain the depth of these seismic events.

2. Detection and Hypocenter Determination of Tremor Events

2.1 Detection and location of tremors by automated analysis

To detect and locate tremor sources, we used the envelope correlation method (Obara, 2002; Wech and Creager, 2008; Ide, 2010, 2012). The original datasets that we acquired comprise continuous velocity seismograms in two horizontal components recorded at 24 seismic stations of the Hi-net network of the National Research Institute for Earth Science and Disaster Prevention, Japan (Fig. 1(a));

Fig. 1. Map views of Kyushu. (a) Tectonic setting of Kyushu. Triangles represent active volcanoes, and black crosses indicate the locations of Hi-net seismic stations used in this study. The thick dashed line demarcates the geographical extent of the KPR (Park *et al.*, 2009). The thin dashed line represents the position of a cross-section presented by Hirose *et al.* (2008), which is shown in Fig. 5(b). The black arrow indicates the velocity and movement direction of the subducting plate. The green rectangular area represents a horizontal projection of the source fault of SSEs, as proposed by Yarai and Ozawa (2010). (b) Distribution of detected seismic events (red dots) picked up by both the duration (>30 s) and the power spectrum ratio (>10). The two red stars are the LFEs detected by JMA. The blue dots are intra-slab earthquakes located by JMA and used in this study. Distribution of seismic events picked up based solely on duration are shown as a grey dot.

the data were recorded at 100 samples per second (sps). The data were bandpass filtered between 2 and 8 Hz, converted to envelope waveforms, low-pass filtered below 0.2 Hz, re-sampled at 2 sps, and then divided into half-overlapping 300 s time windows. For each time window, we calculated cross-correlation coefficients between stations, and if the maximum value of the coefficients exceeds 0.6 for more than 40 station pairs, we would then try to locate the source using a horizontally-layered velocity structure constrained by the results of *S*-wave tomography (Saiga *et al.*, 2010) (Table 1). If the standard deviation of misfits between the observed and calculated arrival time lags was larger than 1.0 s, we rejected the result. Otherwise, to distinguish tremor events from ordinary earthquakes or, alternatively, analytical artifacts due to background noise, we rejected all tremor events with a duration of less than 30 s, calculated as a half-value width of the stacked envelope aligned along the synthetic travel time. Additional details of the hypocenter detection and event rejection schemes can be found in Ide (2010, 2012). Throughout the data collection period, from April 2004 to September 2009, we detected and located 2097 tremor events (Fig. 1(b)).

As an extension of this method of tremor detection and location, we also compared the power spectrum ratios observed for each set of data—i.e., the ratio of the mean power at low frequencies (2–8 Hz) to the mean power at high frequencies (10–20 Hz) calculated from the original 100 sps data—based on the fact that tremors have a stronger power at low frequencies (Shelly *et al.*, 2006). If the average of the spectrum ratios for all stations within 50 km of each

Table 1. The 1D velocity structure used in this study, following Saiga *et al.* (2010).

Depth (km)	<i>S</i> -wave velocity (km/s)	<i>P</i> -wave velocity (km/s)
0–5	3.2	5.5
5–10	3.4	5.8
10–20	3.7	6.3
20–30	4.1	6.9
30–35	4.3	7.6
35–50	4.5	7.9
50–75	4.6	7.9
75–	4.7	8.1

epicenter is higher than 10, the signal is ultimately considered to be a tremor. As a result of employing this new criterion, most of the detected events were rejected as either earthquakes or artifacts of background noise. However, 488 ‘genuine’ tremor events were successfully picked up (Fig. 1(b)). By comparing the red dots and grey dots in Fig. 1(b), the scattered events that are usually considered to represent either normal earthquakes occurring outside of the station distribution or analytical artifacts originating from background noise are removed, whereas clustering events, which usually represent genuine tremor events, are not removed by this method.

Most small inland tremor events are clustered near active volcanoes, and many events are included in one large cluster in the Bungo channel, which is part of a well-known zone of tremor activity within the Nankai subduction zone.

Fig. 2. (a) Examples of tremor waveforms. (b) Power spectrum of noise (blue), ordinary earthquakes (black), and tremors (red). (c) Event times (dots) of tremor and SSE times (pink arrows) (Yarai and Ozawa, 2010) for the whole period (top) and a period of 3 days (bottom). The timing of ordinary earthquakes ($M_w > 5$) worldwide is compared in the lower layer.

Another notable tremor cluster, comprising 46 individual tremor events, exists in southeastern Kyushu, where two LFEs and the tremor activities were reported separately by JMA and Ide (2012), respectively. Figure 2 shows representative examples of waveforms, the power spectrum, and time sequences of tremor events. Although the amplitude of tremor signals (~ 100 nm/s) in Kyushu is not as large as those in Nankai, which makes tremor detection and location more difficult, the shape of the power spectrum is remarkably different from that of typical earthquakes near the tremor source area. The number of detected tremor events, 46 events in 66 months, is much smaller than that detected in Nankai. Although the tremor activity recorded shows some regularity, with a noticeable recurrence at almost every 8 months, evidence of tidal modulation is not apparent. Notably, the active periods of tremor activity identified in this study are not coincident with teleseismic large earthquakes ($M_w > 5$) in the Harvard CMT catalog (Fig. 2(c)).

2.2 Accurate estimation of tremor depths

Although the automated detection method described above can be used to estimate the depth of tremor events, it uses only S -waves, and, ultimately, these absolute depth determinations are relatively imprecise. A better constraint on the absolute depth of tremor events is provided by the S - P times for events occurring immediately beneath the

observation network, as demonstrated by La Rocca *et al.* (2009) in the Cascadia subduction zone. At some stations in Kyushu, P -waves are often visible in the vertical component, while S -waves are dominant in all components. Therefore, we can measure S - P times from cross-correlation functions between vertical and horizontal seismograms. At one station, two 40 s velocity seismograms, in both vertical and horizontal directions, are bandpass filtered between 2 and 8 Hz, converted into envelope waveforms, and smoothed by using a running average over 15 samples. The direction of the horizontal waveform, which is $N20^\circ W$ in this study, is chosen as P - S peaks, explained in the next paragraph, can be seen most clearly by testing some directions. In addition, we calculated a cross-correlation function for each dataset, changing the time lag between two envelope waveforms from -20 to 20 s.

Figure 3 shows representative examples of cross-correlation functions calculated for half-overlapping 40 s time windows of a 600 s record. When tremor signals are visible in velocity waveforms (left and center panels), two peaks appear in the cross-correlation function (right panel), corresponding to P - S and S - S combinations between the vertical and horizontal components. The amplitude of each correlation function is normalized and stacked for the 600 s period. The stacked function (bottom panel) shows more

Figure 4 summarizes the results of our depth estimations for tremor events. All of the S - P times estimated for a large number of tremor sequences are between 4.5 and 5.0 s (Fig. 4(a)), which suggests that the depths of all of these recorded tremor events are distributed over a narrow range. We also calculated S - P times for some intra-slab earthquakes that occurred near the recorded tremor events, and obtained cross-correlation functions with two distinct peaks separated by about 6.3–7.4 s (Fig. 4(b)). Notably, these calculated S - P times are consistent with the visual readings of

3. Discussion

One question that arises from this study is whether these tremor events occur on the plate interface, like those in Nankai. There are two plate interface models for the subduction zone beneath Kyushu (Yagi and Kikuchi, 2003; Hirose *et al.*, 2008), and the respective locations of these two contrasting model plate boundaries at depth are separated by about 10 km, as shown in Fig. 5. Figure 5 also depicts the locations of two LFEs at depths of around 50 km that are consistent with the plate boundary of Hirose *et al.* (2008), as estimated by Yoshioka *et al.* (2008) using *S*-wave arrival times and the double difference method. However,

Fig. 4. (a) Stacked cross-correlation functions of different tremor events. (b) Cross-correlation coefficients of intra-slab earthquakes shown in Fig. 1(b). (c) Relations between S - P time and the source depth calculated using the 1D velocity structure (Table 1) for different epicentral distances. The distributions of tremor events and intra-slab earthquakes are shown by pink and blue crosses, respectively, with error bars calculated as the half-value width of cross-correlation functions for tremors, and visual reading errors for intra-slab earthquakes.

the depth of one of these LFEs (July 18, 2005) and our estimated locations of tremor sources were both derived using our method which employs measured S - P times, which collectively yield a precise depth of about 40 km—i.e., about 20 km above intra-slab earthquakes and consistent with the shallower plate boundary proposed by Yagi and Kikuchi (2003).

Here, it is important to highlight that the seismic tomographic image of Hirose *et al.* (2008) shows a sharp boundary between high- and low- V_p/V_s regions (thin dashed line in Fig. 5(b)), located close to our estimated tremor sources and again about 20 km above intra-slab earthquakes. Therefore, we consider this boundary to represent the plate inter-

face beneath Kyushu Island, with all intra-slab earthquakes occurring on, or below, the oceanic Moho. The separation of up to 20 km between the plate interface and the oceanic Moho appears to be somewhat large, but this may reflect the subduction of the anomalously thick crust of the KPR. Another relevant observation is that a region with a high Poisson ratio occurs near the tremor sources identified in this study (see Fig. 5(a)), which is interpreted as evidence for the existence of fluids. In addition, Yoshioka *et al.* (2008) estimated the thermal structure of the Kyushu subduction zone, indicating that temperatures are about 350–400°C near the tremor sources, which is typical of the conditions required for tremor generation.

Fig. 5. (a) Cross-section of Poisson's ratio structure near the tremor source area, after Tahara *et al.* (2008). The gray rectangle at the top of the cross-section represents land area. The black dashed line represents the Moho discontinuity. The red and blue lines demarcate the plate boundary estimated by Yagi and Kikuchi (2003), showing co-seismic, and aseismic, slip, respectively. The white dots indicate the distribution of ordinary earthquakes. Tremor distribution is superimposed. The red and blue bars are the estimated depth of tremors based on S - P times, and based on the envelope correlation method, respectively. The two red stars indicate the hypocenters of LFEs relocated by Yoshioka *et al.* (2008). (b) Cross-section of V_p/V_s distribution near the tremor source area, after Hirose *et al.* (2008). The gray rectangle at the top of the cross-section represents land area. The red triangle indicates the location of an active volcano. The red line demarcates the plate interface estimated by Hirose *et al.* (2008). Black dots indicate the distribution of ordinary earthquakes. Red dots are LFEs, including volcanic ones, detected by JMA and relocated by Hirose *et al.* (2008). The thick dashed line represents the Moho discontinuity. The thin dashed line represents the boundary between low- and high- V_p/V_s zones. Tremor distribution is also superimposed as for (a). (Reprinted from *Phys. Earth Planet. Inter.*, **167**, M. Tahara *et al.*, Seismic velocity structure around the Hyuganada region, Southwest Japan, derived from seismic tomography using land and OBS data and its implications for interplate coupling and vertical crustal uplift, 19–33, Copyright (2008), with permission from Elsevier.)

In the Hyuga-nada region, offshore of the main location of Kyushu tremor activity, the afterslip of the 1996 (Yagi *et al.*, 1999) Hyuga-nada earthquakes and faint SSEs have both been observed, and their source region is estimated to be at slightly shallower depths than the tremor events beneath Kyushu (Figs. 1(a), 5(a)) (Yagi and Kikuchi, 2003; Yagai and Ozawa, 2010). The amplitude of these Hyuga-nada SSEs is smaller than the amplitude of the SSEs observed in the Nankai region; i.e., it is similar to the observed differences in tremor activities between the two regions. However, the time constants of SSEs, and the ~ 2 year recurrence interval and ~ 1 year duration of SSEs, are not consistent with comparable data on tremors detected in this study, which revealed an 8 month recurrence interval and a duration of only a few days. Nevertheless, based on all of these arguments and geophysical constraints, we conclude that the plate boundary beneath Kyushu exists at a depth of 40 km, and the tremor activity recorded in this study is interpreted to represent shear slip along the plate interface.

Also of note is the fact that tremor activity detected in Kyushu is spatially localized. The narrowness of the tremor source along the dip direction could be related to the steepness of the subducting PHS plate, which results in a rapid increase in temperature and a corresponding rapid transition from stick-slip to stable slip along the plate interface. The distribution of tremors along strike seems to coincide with the location of the KPR, which is a remnant arc that is considered to have undergone a comparatively small amount of dehydration during subduction compared with the surrounding oceanic plate. Within the Nankai subduction zone,

a relatively inactive tremor zone identified previously is interpreted as a result of ridge subduction (Matsubara *et al.*, 2009). Therefore, the coincidence between tremor activity and the KPR determined in this study appears to contradict some previous views and geophysical models of the region. However, additional studies are needed to examine the effects of ridge subduction on fluid distribution, thermal structure, and other important geological and geophysical factors linked with tremor generation.

4. Conclusions

In southern Kyushu, we recorded tectonic tremor events that exhibit a recurrence interval of about 8 months, spaced closely along the down-dip direction of small SSEs and their associated source fault (Yagai and Ozawa, 2010), and laterally distributed across the region where the KPR is being subducted (Park *et al.*, 2009). The depth of this tremor activity is estimated to be 35–45 km, based on tremor S - P times, which is consistent with the location of the plate boundary estimated by Yagi and Kikuchi (2003) and is coincident with a sharp boundary in the V_p/V_s structure reported by Hirose *et al.* (2008). The existence of fluids near our tremor sources has also been suggested (Hirose *et al.*, 2008; Tahara *et al.*, 2008), and the temperature in this region is 350–400°C (Yoshioka *et al.*, 2008); these conditions are considered to be essential for tremor generation. Based on these geophysical observations, the tremor activity in Kyushu probably represents shear slip along the plate interface, akin to tremor activity recorded in Nankai and other subduction zones. Ultimately, by determining the depth of such tremor activity accurately in this study, we were able

to constrain the location and geometry of the plate interface in the Kyushu region. In future geophysical studies of the area, it will be important to consider the effects of subduction of the KPR (a remnant arc) on the mechanisms of tremor activity. In this regard, it seems clear that suitable conditions for tremor activity in southern Kyushu would locally be realized due to ridge subduction, although the influence of subduction of the KPR on the process of tremor generation in the region has yet to be explored.

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