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Hypocenter Distribution of Deep Low-frequency Tremors in Nankai Subduction Zone, Japan

Takuto MAEDA* and Kazushige OBARA**

*Earthquake Research Department,
National Research Institute for Earth Science and Disaster Prevention, Japan
(Current affiliation : Center for Integrated Disaster Information Research,
Interfaculty Initiative in Information Studies, the University of Tokyo, Japan)
maeda@eri.u-tokyo.ac.jp
**Earthquake Research Department,
National Research Institute for Earth Science and Disaster Prevention, Japan
obara@bosai.go.jp

Abstract

Major clusters of nonvolcanic, deep low-frequency tremors distributed along the strike of the subducting Philippine Sea Plate with a length of 600 km from the Tokai area to the Shikoku area are analyzed, and the regional differences are discussed through the relocation of the tremor hypocenters. Reliable tremor sources relocated for each cluster are clearly coincident with the depth of the plate interface, which suggests that the tremors in western Japan occur mainly at the plate interface. In particular, the planar distribution of the relocated tremors is consistent with the fault geometry of the slow slip events and very low frequency earthquakes in western Shikoku, whereas the depth distribution of the eastern part of the belt-like zone is more scattered. The scattering of the relocated tremor particularly in the Tokai region may indicate the tectonic and/or geological situation around the plate interface.

Key words : Low frequency tremor, Slow slip events, Subduction zone, Plate boundary

1. Introduction

Recent observations of the deep low-frequency tremors (Obara, 2002 ; Kao *et al.*, 2005 ; Shelly *et al.*, 2006), slow slip events (Rogers and Dragert, 2003 ; Obara *et al.*, 2004; Hirose and Obara, 2006; Obara and Hirose, 2006), and very low frequency earthquakes (Ito *et al.*, 2007) in subduction zones mainly in southwestern Japan and Cascadia have aided the understanding of new types of slow earthquakes having various time scales (Ide *et al.*, 2007a).

Although deep low-frequency tremors are observed at many subduction zones (Schwartz and Rokosky, 2007), the Nankai subduction zone in Japan is one of the most suitable areas for studies because of the existence of a densely distributed high-sensitivity seismograph network (Hi-net) (Obara *et al.*, 2005). Using this network, many aspects of the tremors have been investigated, for example, source mechanisms (Ide *et al.*, 2007b), the duration-amplitude relation (Watanabe *et al.*, 2007), and dynamic triggering (Miyazawa and Mori, 2006). Obara and Hirose (2006) have shown that active deep low-frequency tremors show a strong temporal and spatial correlation with the slow slip detected by Hi-net tiltmeter records. Since the slow slip events are considered as stick-slip events at the deeper portion of the subducting plate boundary (Hirose and Obara, 2006), deep low-frequency tremors are also considered to occur on the plate boundary or its neighborhood.

Since the observed tremor signal is very complicated, tremors are located using an envelope correlation (Obara, 2002). Although the envelope correlation method successfully detects the large number of tremors, the accuracy of the location is expected to be low especially in depth. On the other hand, Japan Meteorological Agency

* Yayoi 1-1-1, Bunkyo, Tokyo, 113-0032, Japan
** Tennodai 3-1, Tsukuba, Ibaraki, 305-0006, Japan

(JMA) routinely locates tremors with the label "low-frequency earthquake (LFE)" when it recognizes the impulsive onset of the S waves that occasionally appear in the tremor signal.

Recently, Shelly *et al.* (2006) have discussed the fine-scale distribution of LFEs relocated using the cross correlation of tremor waveforms for the Shikoku region, western part of the Nankai subduction zone of the Philippine Sea Plate. They have shown that the tremors occur along the plate interface in the high-V_P/V_S region. Other recent tomographic results also support the fact that such tremors occur in the high- V_P/V_S region (Nugraha and Mori, 2006; Matsubara *et al.*, 2007). Further studies about the mechanism of LFEs (Ide *et al.*, 2007b) and hypocenter distribution in Tokai area (Ohta and Ide, 2008) also support that the tremor in Nankai subduction zone occurs at the plate interface.

On the other hand, in a Cascadian region, Kao *et al.* (2005) suggests the different interpretation of the mechanism of tremor. They have reported that the tremors occurs at the shallower extent from the plate interface. Based on the spatial correlation between locations of tremor locations and seismic reflectors, they interpret that tremor is fluid-related phenomena, rather than the slip on the plate interface. To reveal the mechanism of tremor and slow-slip occurrence, a fine-scale location of tremors, particularly with regard to the depth in many regions, is indispensable. In this paper, we report on the distribution and regional difference of hypocenter of tremors over the entire Nankai subduction zone in Japan, having a length of 600 km from the Tokai area to the Shikoku area.

2. Data and Method

We relocate the deep low-frequency tremors included in the JMA catalog occurring from October 2000 to November 2006 in the Nankai subduction zone, Japan, by using the double-difference algorithm (Waldhauser and Ellsworth, 2000) with accurate differential traveltime estimated by the cross-correlation method (e.g., Schaff *et al.*, 2004; Shelly *et al.*, 2006).

First, we calculate the differential traveltime of the *S* waves by cross correlation among different tremor signals observed at the same station. We use a vertical-component waveform of a lapse time 2.56 s around the catalog arrival time of the *S* waves recorded at the Hi-net borehole stations. Band-pass filters of 2–10 Hz are employed to enhance the tremor signals. **Fig.1** shows well-correlated waveforms observed at the KWBH station in western Shikoku as an example. The bars and shaded area denote the catalog traveltime of *S* waves and the time window for the cross-correlation analysis, respectively. Though the analysis window is limited, the two signals are very well correlated up to the coda portion of the *S* wave. This shows that the distance between these events is smaller than the characteristic length of the inhomogeneity around the source (Nakahara, 2004), and it implies that the relocation using the cross correlation is effective. However, we also find that the cross-correlation values are generally low as compared to the repeating earthquakes (e.g., Uchida *et al.*, 2003; Kimura *et al.*, 2006). Therefore, an appropriate tolerance is required to discard improper traveltime estimations.

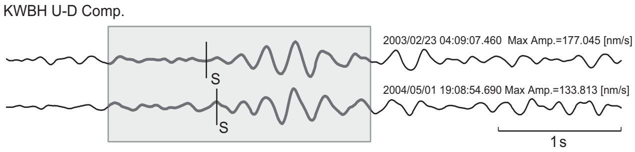

Fig. 1 Example of waveforms of different tremors observed at KWBH station in western Shikoku region. Vertical-component seismograms filtered with a corner frequency of 2–10 Hz are shown. The vertical lines indicate the catalog onset time of the *S* waves. A shaded area with a lapse time of 2.56 s is used for cross-correlation analysis.

To verify the adequacy of the use of the cross-correlation analysis having lower cross-correlation values, we compare the histograms of cross-correlation value of the real waveform and random noise, as shown in **Fig.2**. The same filters and cross-correlation technique are employed for the synthetic white noise. Because of the small length of the time window and band-pass filter, the cross correlation accidentally has a moderate value even if there is no real signal. However, we find that for random noise, the cross-correlation value very rarely exceeds 0.6. From this comparison, we decide to use a relative traveltime whose cross-correlation value is larger than 0.65 by trial and error. To achieve high-resolution relocation particularly with regard to depth, the use of *P* wave information is crucial. Although there are very few *P* wave catalog onsets, we estimate the differential traveltime of the *P* waves by using the predicted traveltime from the *S* wave arrival according to the approach by Shelly *et al.* (2006).

From 28.3 million possible pairs that satisfy the criteria of the cross-correlation coefficient, we select 724,345 cross-correlation pairs from the entire Nankai subduction zone. Thirty-three percent of this relative traveltime data is comprised of the *P*-waves. Absolute travel time measurements (1.96 million) that mainly consist of *S* waves are used together with the cross-correlation data for

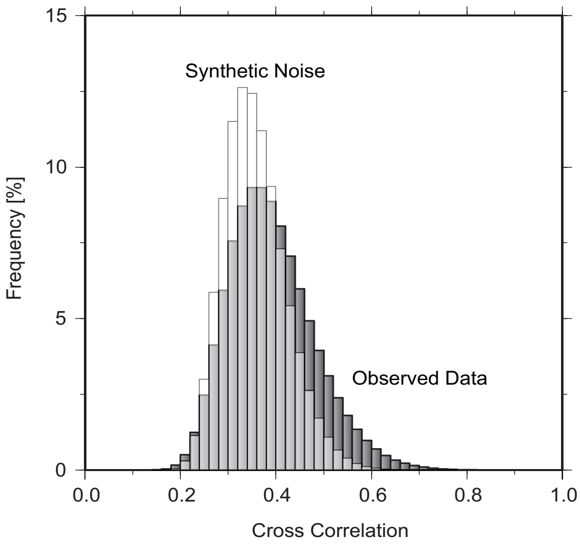

Fig. 2 Histogram of maximum cross-correlation values of real data observed at Hi-net stations (dark gray) and synthetic white noise data (white), respectively. We note that these two distributions is overlapped (light gray) for the cross-correlation values between 0.18-0.62.

relocation. To discuss a reliable depth distribution of the tremors, we further select the hypocenter that is located by at least 10 traveltime data that contains at least three *P*-wave traveltimes. Low-frequency earthquakes that are not related to the subduction zone are manually excluded. We finally relocate 635 tremors that satisfy the above criteria.

3. Results and Discussions

The circles and dots in **Fig.3** (**a**) show the spatial distribution of the relocated tremor and regular earthquakes whose depth is more than 20 km, respectively. The oceanic Moho depth estimated by receiver function analysis (Shiomi *et al.*, 2008) is shown by the contour lines. We divide the entire study area into five subregions: (I) western Shikoku, (II) eastern Shikoku, (III) western Kii, (IV) eastern Kii, and (V) Tokai. These regions are shown by boxes in the map.

Tremors mainly occur at the depths between 35 km and 40 km for all regions, although the distribution is scattered even after the relocation. Tremors are absent in the Kii channel and Ise Bay, where the oceanic Moho strongly bends. The relocated epicenters of the tremors are horizontally clustered. Especially in western Shikoku region, we can recognize the linear-cluster in NW-SE direction. Although the catalogs by the envelope correlation method of Obara (2002) show a belt-like distribution of tremor epicenters, a number of tremors along the belt parallel to the subducting Philippine Sea Plate have local peaks (Obara and Hirose, 2006). The locations of relocated clusters in **Fig.3** (**a**) show a good coincidence with the peaks of the tremor activities. In other words, the tremors relocated in this study are at the most active portion of whole tremor activities. Minor tremors that occur at the off-peak of the activity may have more incoherent signals; therefore, we cannot relocate them using the cross-correlation technique.

Fig.3 (**b**) shows the hypocenter depths of the relocated events along five profiles (A–E) as the representatives of the subregions shown in **Fig.3** (**a**). Tremors (circles) and normal earthquakes (dots) occurring within a distance of ± 20 km from the profile are shown in each panel. The oceanic Moho depths along the profile are shown by the thick black line. At the A-A' profile in the western Shikoku region, the tremors show a planar distribution clustered 5–10 km above the oceanic Moho, as reported by Shelly *et al.* (2006). This distribution is consistent with the fault geometry of slow slip events (Hirose and Obara, 2006) and the mechanism of very low frequency earthquakes (Ito *et al.*, 2007). On the other hand, the depth distribution of eastern part of the Nankai subduction zone, particularly the Tokai region (E-E' profile), is more scattered. This depth distribution is not caused by the initial hypocenter location. This scattered distribution is reconstructed even if we set the initial depth of the tremor artificially on the plate interface.

Fig.4 shows the histograms of relative depth of tremors measured from the oceanic Moho at each subregion. A negative relative depth means that a tremor that occurs at a shallower depth than the oceanic Moho. The average relative depth (μ) and standard deviation (σ) are shown in each panel. A histogram of the entire study area is also shown in the figure. The histogram has a clear peak, similar to the normal distribution. On average, tremors occur at a depth that is 4.9 km shallower than the oceanic Moho depth. The relative depth from oceanic Moho is in good agreement with the average thickness of the oceanic crust. Therefore, we interpret that deep low-frequency tremors in the Nankai subduction zone occur mainly at the plate boundary, which is different from the results of the observation of the Cascadia by Kao *et al.* (2005). Our result support that the interpretation by Ito *et al.* (2007) that the tremor occurs at micro cracks in the transition zone on the plate boundary with high pore-fluid pressure due to the dehydration of subducting oceanic crust associated with the migration of the slow slip.

In each subregion except for western Kii, the depth distribution shows clear peaks. In the western Kii subregion, it is difficult to discuss the distribution because of the very small number of estimated tremors. There are

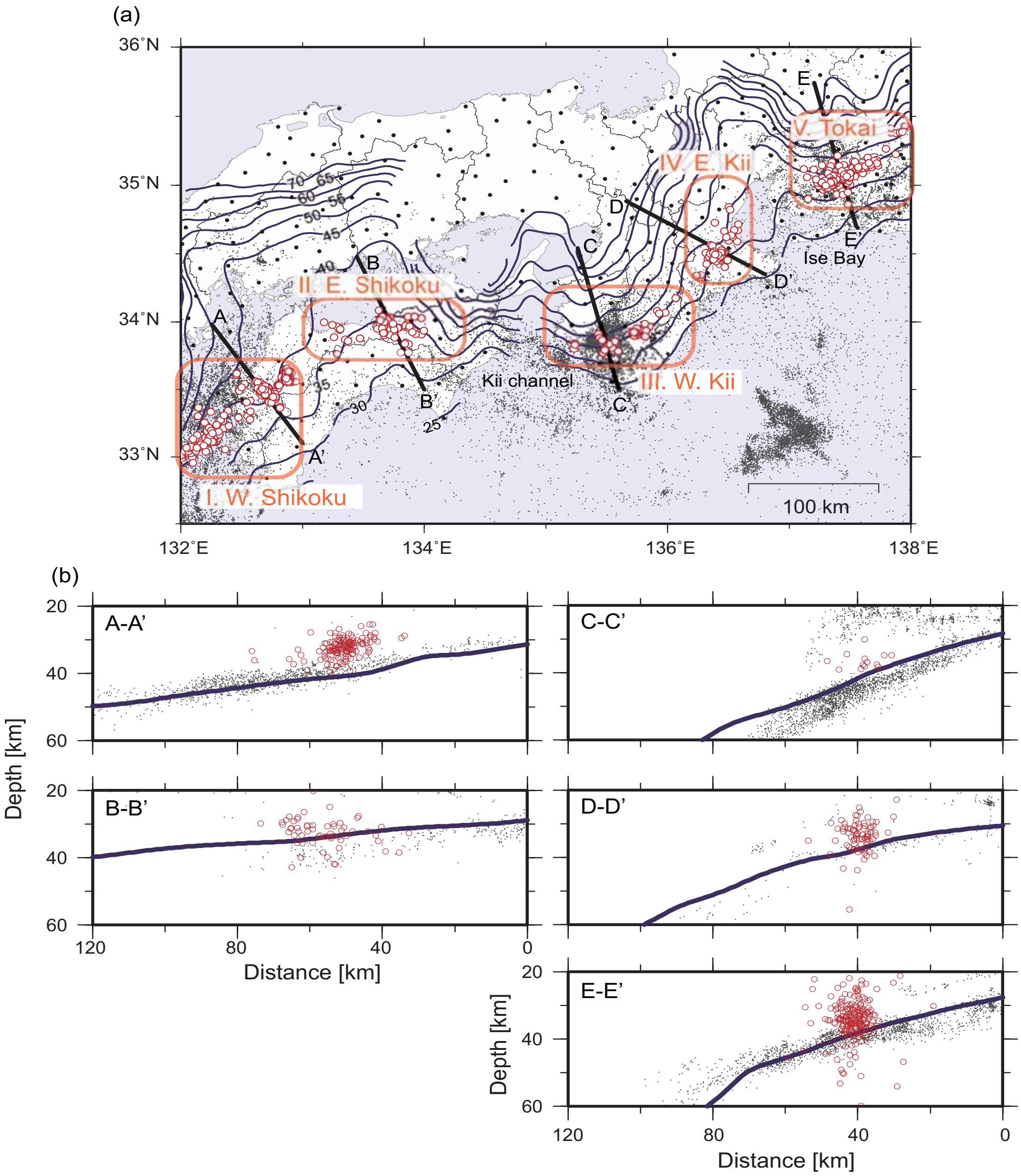

Fig. 3 (**a**) Spatial distribution of relocated deep low-frequency tremors (circles) for the period of 2001 to 2006. The depth contour lines indicate the depth of the oceanic Moho estimated by the receiver function analysis (Shiomi *et al.*, 2008). Normal earthquakes are indicated by dots. Hi-net stations are denoted by the filled black circles. The five boxes (I toV) show the subregions for the comparison of depth of tremors (refer to text). (**b**) Hypocenter distributions of relocated tremors indicated by five cross sections (A'-A to E'-E) shown in **Figure 3** (**a**). Tremors (circles) and normal earthquakes (dots) within 20 km from the cross section are plotted in each figure. The thick lines indicate the oceanic Moho depth.

slight differences between the average relative depths from the oceanic Moho among the regions from 1.9 km to 7.5 km. This difference may be owing to the difference in the thickness of the oceanic crust. Since the receiver function analysis (Shiomi *et al.*, 2008) reveals a strong impedance contrast associated with the Moho discontinuity, it is difficult to determine the exact plate interface. Additionally, the depth distribution of western Shikoku has a significantly smaller deviation than other regions. In particular, in the Tokai area, the standard deviation of the depth is twice that in western Shikoku. This difference may be caused by the regional difference in the crack distribution off the plate interface. Since the micro-crack distribution of tremor should be related to the

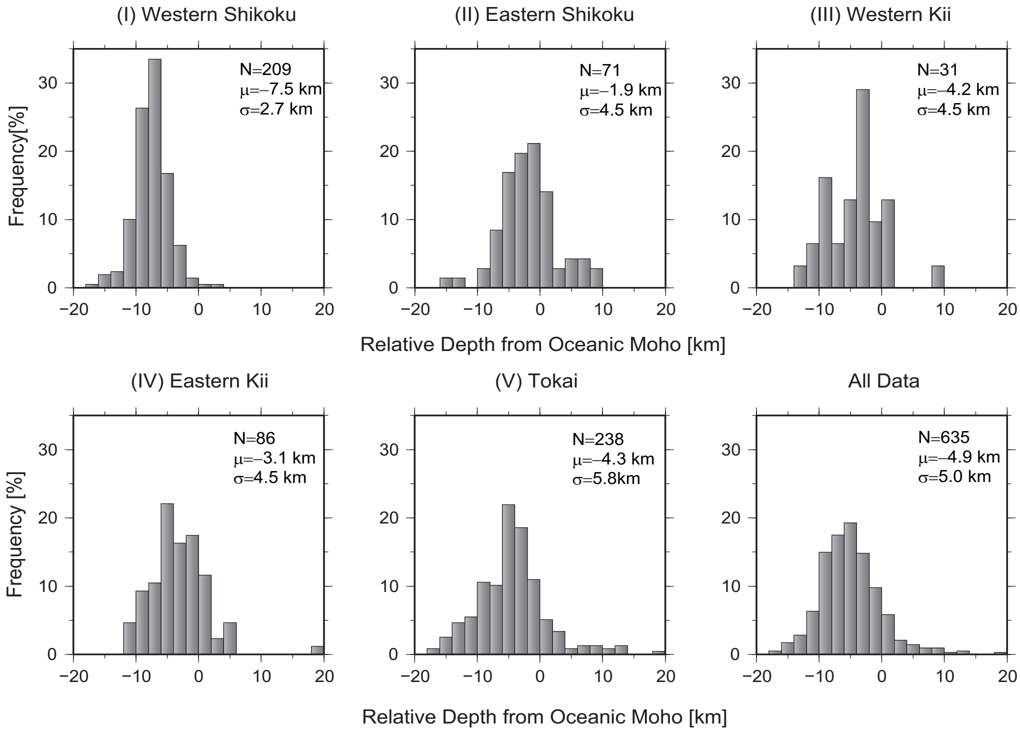

Fig.4 Histogram of relocated tremor depth relative to the oceanic Moho, which is estimated by receiver function analysis, for the five subregions shown in **Figure 3** and entire western Japan region. A negative relative depth implies that the tremor is located at a shallower depth than the oceanic Moho. The average deviation (μ) and standard deviation (σ) of each distribution are shown in each panel.

dehydration process of subducting oceanic crust under the certain temperature-pressure condition, our result may suggest that the scattered distribution of tremor in depth is owing to the inhomogeneity of crack distribution caused by the lateral variation of the temperature on the plate interface of Nankai subduction zone.

4. Summary

We systematically relocated the major clusters of deep low-frequency tremor in southwestern Japan by the measurements of differential traveltimes by cross-correlation technique of *P*- and *S*-waves between events. Relocated clusters of tremor show planner distribution at the depths several kilometers above the oceanic Moho, which suggests that the tremor occurs on the plate boundary. The scattered distribution of tremor and the rationality of average relative depth of tremors from the oceanic Moho may reflects the inhomogeneity of micro-crack distribution and thickness of oceanic crust.

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西南日本沈み込み帯における深部低周波微動の震源分布

前田拓人*・小原一成**

*独立行政法人 防災科学技術研究所 地震研究部
（現：東京大学大学院 情報学環 総合防災情報研究センター）
**独立行政法人 防災科学技術研究所 地震研究部

要　旨

西南日本に沈み込むフィリピン海プレートにおいて，幅約 600 km にわたって発生する深部低周波微動の震源分布の再検討とその地域性を検討した．波形相関を用いた相対走時推定に基づく精密震源再決定の結果，東海地方から四国西部までの広域にわたって，深部低周波微動が海洋性モホ面の約 5 km 上方で発生していることが明らかになった．このことは，深部低周波微動は沈み込むプレート境界面で発生していることを示唆している．四国西部地域においては再決定された微動が面的に分布しているが，東海地方においては相対的に分布が深さ方向に広がりを持っている．また，再決定された微動の深さとレシーバー関数解析により推定された海洋性モホ面の相対深度は地域ごとに異なっている．これらの特徴はプレート境界面の水平方向の不均質構造を表しているのかもしれない．

キーワード：深部低周波微動，スロースリップ，沈み込み帯，プレート境界