Resistivity image of the Philippine Sea Plate around the 1944 Tonankai earthquake zone deduced by Marine and Land MT surveys

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(Received November 30, 2004; Revised February 17, 2005; Accepted February 21, 2005)

The Nankai Trough is an active convergent region in southwest Japan and mega-thrust earthquakes have repeatedly occurred in some areas of its plate-boundary interface. Generation of mega-thrust earthquakes is inferred to be related to the existence of water. The resistivity structure is very sensitive to the existence of water. For that reason, it is important to obtain the resistivity image around the rupture area of mega-thrust earthquakes. We carried out land and marine magnetotelluric surveys in the Kii Peninsula and the offshore Kii Peninsula where the 1944 Tonankai earthquake occurred. We constructed a 2D resistivity model using an inversion technique. The modeled resistivity structure portrayed the Philippine Sea Plate as a resistive region. However, its resistivity becomes more conductive as the plate subducts, showing 10 Ω-m around the down-dip limit. These characteristics are considered to relate to the water. Therefore, we infer that water might control the generation of mega-thrust earthquakes.

Key words: Nankai trough, 1944 Tonankai earthquake, Magneto-telluric, OBEM, Resistivity, Conductor, Fluid, Kii Peninsula.

1. Introduction

The Nankai Trough is located offshore of the Kii Peninsula and is an active convergent region in southwest Japan. Around this region, mega-thrust earthquakes with magnitudes greater than 8 have repeatedly occurred on some portions of the plate-boundary interface (e.g. Ando, 1975). The 1944 Tonankai earthquake was generated at the plate-boundary interface in this area. Large earthquakes occurred around the trough axis 100 km offshore southeast of the Kii Peninsula on 5 September 2004.

Multi-channel seismic (MCS) and ocean bottom seismometer (OBS) reflection surveys were carried out around the Nankai trough region to elucidate what occurred on the thrust earthquake rupture zone. Park *et al.* (2002) reported the existence of splay faulting that branches upward from the plate-boundary interface at a depth of 10 km. Moreover, the existence of fluid is indicated because of reverse polarity reflections. Nakanishi *et al.* (2002) performed a wide-angle seismic survey and constructed a crust and uppermost mantle P-wave velocity model across the Nankai Trough around the Tonankai earthquake rupture zone. According to Nakanishi *et al.* (2002), the thickness of the oceanic layer is 7–8 km and dip angle across the rupture zone of the 1944 event is estimated at 11°. That study also determined that the rupture zone of the 1944 event does not reach the fore-arc mantle and concluded that this depth agrees with the locked zone defined from a thermal model by Hyndman *et al.* (1995).

Magneto-telluric (MT) surveys have been carried out to obtain electrical resistivity images of the subduction zone. These help us to investigate the existence of water which plays an important role in the occurrence of earthquakes. Therefore, the resistivity image is very important data. For Shikoku, southwest Japan, Yamaguchi *et al.* (1999) showed a highly resistive Philippine Sea Plate overlaid with a thin conductive layer by using the Network-MT method (Uyeshima *et al.*, 2001). In the ‘Electro Magnetic Study of the Lithosphere and Asthenosphere Beneath’ the Juan de Fuca Plate (EMSLAB) project, seafloor and land electromagnetic (EM) surveys were carried out in the area of the Juan de Fuca subduction system. (Wannamaker *et al.*, 1989). The subducting plate image from ocean to land was imaged by forward modeling.

In the Kii Peninsula, Fuji-ta *et al.* (1997) found the top of a conductor at a depth of 20 km. Kasaya *et al.* (2003) also detected a conductor at a depth of 20–50 km by forward modeling of land ULF-MT data. However, their results retain the ambiguity of structural effects below the seafloor. In spite of the importance of a deep conductor in discussing the generation of mega-thrust earthquakes, there have been no previous studies imaging the resistivity structure around a whole area (from up-dip to down-dip limits) of a rupture zone of a mega-thrust earthquake.

The object of this study is to estimate a complete resistivity image of the subducting Philippine Sea plate causing mega-thrust earthquakes and to compare with other geophysical evidence. We made observations on both the

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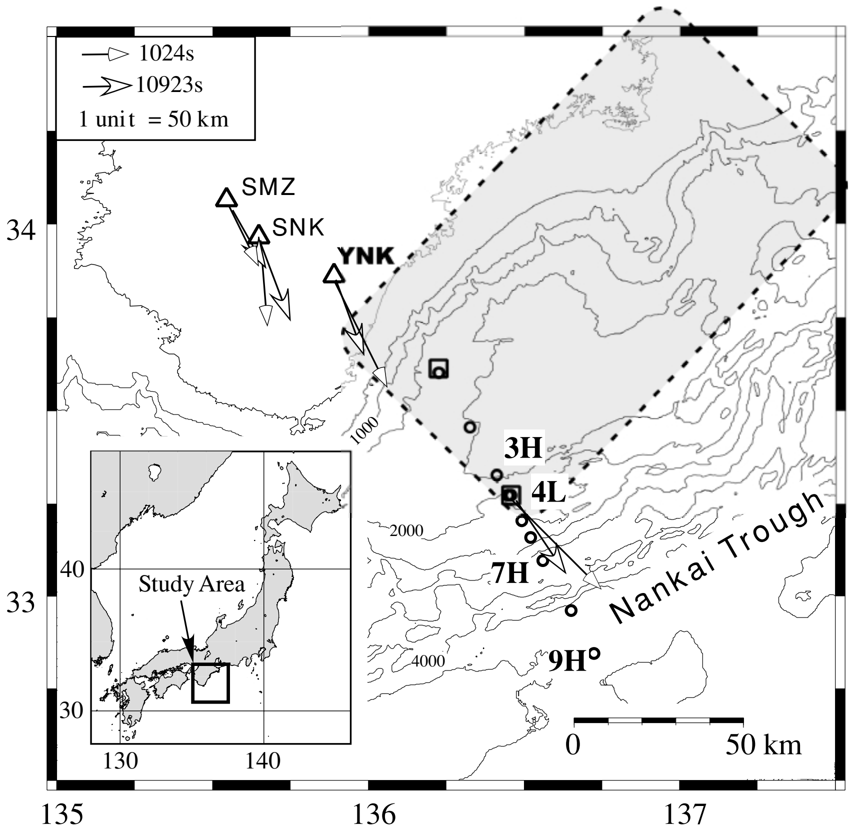

Fig. 1. Destribution of the observation sites around the Kii peninsula and Nankai trough. Triangles, small circles, and small squares denote land electromagnetic observation sites, high frequency type OBEM (HF-OBEM), and long-term OBEM (LT-OBEM), respectively. A shadow zone with dashed line indicates the 1944 Tonankai coseismic slip area with a displacement greater than 0.5 m (Kikuchi *et al.*, 2003). Arrows show the real induction vector at 1024 and 10923 second.

seafloor and land and estimated a resistivity model by applying an inversion technique to the data. The down-dip limit of the rupture area of the 1944 Tonankai earthquake extends below the Kii peninsula (Kikuchi *et al.*, 2003). For that reason, it is important to observe both on the sea floor and on land. Our observation and analysis will contribute to obtaining a complete resistivity image of this subduction zone.

2. Data Acquisition and Time Series Processing

2.1 Marine Magneto-telluric survey

We used two types of the Ocean Bottom Electro-Magnetometer (OBEM) for marine MT survey in KY02-12 cruise, by JAMSTEC R/V Kaiyo. One is denoted as "HF-OBEM" in this paper, and has been developed by EMI Inc. with the capability of making high-frequency EM recordings with a 6.25 Hz sampling rate. This instrument measured two horizontal magnetic components with induction coil sensors and two electric components with silver-silver chloride electrodes. The advantage of this instrument is that it can obtain information of a shallower resistivity structure with a high sampling rate. Acquired data were stored on a compact flash memory. We deployed seven HF-OBEMs and two HF-OBEs (similar to the HF-OBEMs), on a profile (Fig. 1) between 22–23 December 2002. We recovered all HF-OBEMs at the end of this cruise (31 December 2003–1 January 2003).

Another type of OBEM is designated as Long-Term OBEM (LT-OBEM) in this paper, and this can measure three components of magnetic-field variations with fluxgate magnetometers and two horizontal components of electric field variation with electrodes. The maximum lifetime of this instrument is about a year on the seafloor. We deployed LT-OBEMs at two sites during the KY02-12 cruise; these are shown as 1L and 4L in Fig. 1. The recovery operation was carried out on 20 May 2003 during the KR03-05 cruise using JAMSTEC R/V Kairei. In our survey, continuous electromagnetic data sets for three and six months were recorded with a 30 sec sampling rate.

We estimated a MT response (apparent resistivity and phase) using the robust remote reference method (RRRMT) of Chave *et al.* (1987). HF-OBEM data have unknown noises that were independently recorded at each site data. Almost all the noise is high frequency (higher than 0.1 Hz) as well as spikes. Therefore the time series of HF-OBEM were resampled at 4 sec. Then we chose sections of time series data with high coherency between the MT sites by visible selection. Finally, the selected time series were analyzed using the RRRMT method to obtain enough quality MT responses (Goto *et al.*, 2003). The LT-OBEM time series is free from noise and cross-reference processing for each time series was carried out.

2.2 Long-term Magneto-telluric observation in Kii peninsula

On land, MT measurements were started from December 2002 at the three sites shown Fig. 1. Measurements at

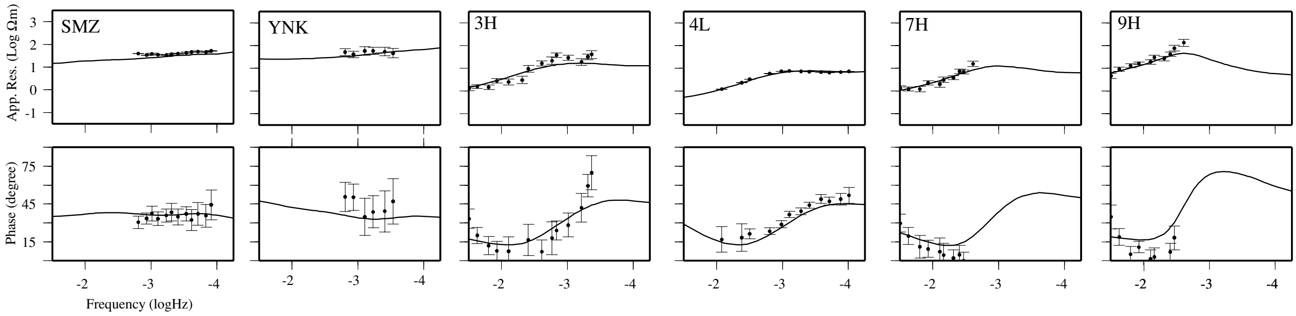

Fig. 2. Apparent resistivity (upper panel) and impedance phase (lower panel) at sites SMZ, YNK, 3H, 4L, 7H and 9H. Lines indicate the calculated MT responses from the resistivity model in Fig. 3.

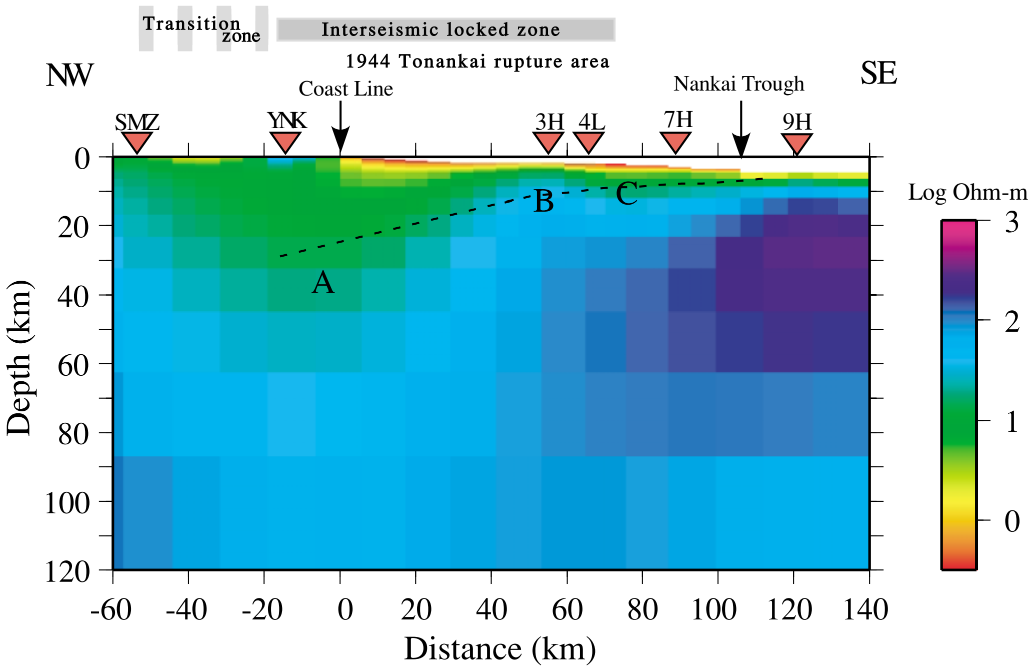

Fig. 3. Best-fitting regional resistivity model across the Nankai Trough and Kii Peninsula. Triangles indicate the horizontal position of observation sites used for the inversion procedure. The 1944 Tonankai earthquake rupture zone are estimated by Kikuchi *et al.* (2003). Interseismic locked and transition zone estimated geothermal data (Hyndman *et al.*, 1995). A dashed line shows the plate boundary estimated by Nakanishi *et al.* (2002).

YNK ended in February 2003. However, synchronous observation between SMZ and SNK continued until the end of May 2003. We used three MT systems with a fluxgate magnetometer (Tierra Tecnica U43 system). This system is continuously synchronized by a GPS clock signal. Consequently, we were able to carry out high-frequency sampling observations and remote-reference processing (Gamble *et al.*, 1979). Electromagnetic data with three magnetic and two horizontal electric components were stored at 4 Hz sampling rate on 384 MB compact flash memory. Some troubles occurred with the flash memory (Kasaya *et al.*, 2003), but sufficient data quality and quantity were obtainable. The RRRMT method was also adopted for land MT data. The remote reference site was used by having another site observed simultaneously.

3. 2D Inversion and Resistivity Structure

Marine data acquired by HF-OBEMs included some noise. Moreover, non-diagonal components at sites located near a coastline were shown to be very large. Therefore, we used only four marine sites for 2D inversion analysis (Fig. 1). The induction arrow from the land stations suddenly changes at the period range of from 50 to 400 sec. However, induction arrows show an almost constant angle (about S30°E) at a longer period range. Therefore, we only used data sets of SMZ and YNK at the long period range from 500 to 10,000 sec, but not the SNK data set because the data quality at long period range was poor. Furthermore, seismic structures around the Nankai Trough show a 2D structure along the trough axis (Park *et al.*, 2002; Nakanishi *et al.*, 1998; Nakanishi *et al.*, 2002). We therefore chose N61.2°E as the strike direction of the 2D analysis.

We performed a 2D analysis for selected TM-mode data to construct the 2D resistivity model because the TM-mode is too robust against 3D structures (Ting and Hohmann, 1981). We used the 2D inversion code with the smoothness constraint based on Akaike's Bayesian Information Criterion (ABIC) developed by Uchida and Ogawa (1993). The original program code did not consider marine MT analysis. For that reason, we used an improved code to be able to treat the MT response on the seafloor (Goto *et al.*, 2002). The initial model was a 100 Ω-m half-space with a fixed 0.25 Ω-m ocean of known bathymetry at both sides of the half-space. The calculation area extends 3000 km deep and about 2000 km on either side of the body to avoid edge effects. We adopt the final model with the smallest ABIC and RMS misfit after 20 iterations as the final model. Figure 3 shows the best-fit model with the position of observation. Observed responses and calculated sounding curves deduced from Fig. 3 are shown in Fig. 2.

4. Discussion

The salient result of this study is the resistivity change of the subducting Philippine Sea Plate (Fig. 3). The uppermost mantle below the ocean in particular is shown as a resistive region (up to 500 Ω-m at a depth of 40 km), and its resistivity decreases to about 10 Ω-m with the plate subducting beneath the island arc crust (A in Fig. 3). We carried out a sensitivity check by forward calculation to confirm the reliability of this transition. Forward tests were performed for models to which the portion below the plate boundary was changed, respectively, to 20, 50, 100, 200, and 500 Ω-m. Consequently, MT responses deduced by the modified model could not explain the observed responses if the modified resistivity was over 100 Ω-m. Resistivity of the subducted plate must therefore include such a decrease of resistivity with subduction. The locked zone (Hyndmann *et al.*, 1995), except for the up-dip limit, is also included in the detected conductive portion. This conductive locked zone consists of the dehydration area of the crust (Yamasaki and Seno, 2003). Moreover, the bottom of the conductor, at depths 40–60 km, also agrees with the surpentinized mantle as estimated by Ulmer and Trommsdorff (1995). Therefore, the possibility exists that the conductive zone below the coastline is related to dehydration and that the water of dehydration may control mega-thrust earthquake generation.

The oceanic crust is portrayed as a layer with a resistivity of about 10 Ω-m (C in Fig. 3). The upper part of the oceanic crust is formed by sediment filled with the pore water. However, resistivity of the subducting oceanic crust at a depth of 10–15 km increases around 40–60 km offshore (B in Fig. 3). This resistivity transition of the oceanic crust may be interpreted as the diagenesis of the sediment, mechanical compaction and the smectite-illite phase transition. Therefore, it is estimated that the amount of water generated by the diagenesis raised up to the sea floor and formed a spray fault and decollement beyond the plate boundary.

Moreover, region "B" in Fig. 3 coincides with the position where the dipping angle changes and the up-dip limit of the rupture area (Park *et al.*, 2002; Nakanishi *et al.*, 2002). Thus, the less fluid condition may be related to the locked plate boundary.

5. Conclusion

We obtained a regional resistivity image of the subducting Philippine Sea Plate along the 1944 Tonankai event. Resistivity model characteristics are as follows:

(1) Oceanic crust shows a conductive layer with some undulation.

(2) The subducting Philippine Sea Plate is imaged as a resistive region and its resistivity decreases with subduction.

(3) Resistivity around the up-dip limit of the 1944 event is relatively high, while the area around the down-dip limit shows low resistivity (high conductivity).

The 1944 Tonankai rupture area is located almost in the conductor. Dehydration from oceanic crust and mantle was estimated around the detected conductor. Therefore, it is suggested that the dehydration water controls the generation of a mega-thrust earthquake.

Acknowledgments. We gratefully thank Captain Hasegawa of the R/V Kaiyo, Captain Tanaka of the R/V Kairei and the crew of both research vessels for performing our surveys. Marine technicians of Nippon Marine Enterprise Co. Ltd. helped onboard operations. Dr. T. Mogi (Hokkaido Univ.) and Dr. A. White (Flinders Univ.) helped clarify the manuscript. We thank Dr. Yamaguchi (Kobe Univ.), Mr. Hoso, Dr. T. Kagiyama and Dr. N. Oshiman (Kyoto Univ.) for helping us land MT surveys. We are grateful to Dr. Y. Ashida (Kyoto Univ.), Dr. T. Watanabe (Nagoya Univ.), Dr. N. Seama (Kobe Univ.) and their students who supported our research cruise. Dr. K. Yamane and AOA Geophysics helped with onboard operations of HF-OBEMs.

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