Centroid moment tensor inversions of offshore earthquakes using a three-dimensional velocity structure model: Slip distributions on the plate boundary along the Nankai Trough

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Running Title
3D CMT inversion along the Nankai Trough

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Summary

Due to complex three-dimensional (3D) heterogeneous structures, conventional one-dimensional (1D) analysis techniques using onshore seismograms can yield incorrect estimation of earthquake source parameters, especially dip angles and centroid depths of offshore earthquakes. Combining long-term onshore seismic observations and numerical simulations of seismic wave propagation in a 3D model, we conducted centroid moment tensor (CMT) inversions of earthquakes along the Nankai Trough between April 2004 and August 2019 to evaluate decade-scale seismicity. Green’s functions for CMT inversions of earthquakes with moment magnitudes of 4.3–6.5 were evaluated using finite-difference method simulations of seismic wave propagation in the regional 3D velocity structure model. Significant differences of focal mechanisms and centroid depths between previous 1D and our 3D catalogues were found in the solutions of offshore earthquakes. By introducing the 3D structures of the low-velocity accretionary prism and the Philippine Sea Plate, dip angles and centroid depths for offshore earthquakes were well-constrained. Teleseismic CMT also provides robust solutions, but our regional 3D CMT could provide better constraints of dip angles. Our 3D CMT catalogue and published slow earthquake catalogues depicted spatial distributions of slip behaviours on the plate boundary along the Nankai Trough. The regular and slow interplate earthquakes were separately distributed, with these distributions reflecting the heterogeneous distribution of effective strengths along the Nankai Trough plate boundary. By comparing the spatial distribution of seismic slip on the plate boundary with the slip-deficit rate distribution, regions with strong coupling were clearly identified.

Keywords:
Computational seismology, earthquake ground motions, earthquake source observations, seismicity and tectonics, wave propagation
1. Introduction

Focal mechanisms of earthquakes and their spatial distributions are important for evaluating tectonic/local stress and strain fields (e.g. Saito et al. 2018, Terakawa & Matsu’ura 2010, Townend & Zoback 2006). To determine focal mechanisms, first-P or S polarisation inversion (e.g. Hardebeck & Shearer 2002, Shelly et al. 2016) and waveform-based centroid moment tensor (CMT) inversion (e.g. Dziewonski et al. 1981, Ekström et al. 2012, Kanamori & Rivera 2008) techniques have been widely used around the world. One-dimensional (1D) Earth models are assumed in typical focal mechanism determination methods. In regions with complex three-dimensional (3D) heterogeneous structures, first-motion solutions using the 1D Earth model systematically show mis-estimations (e.g. Takemura et al. 2016). Although CMT methods based on long-period (> 10 s) waveforms can be applied only for moderate-to-large earthquakes due to signal-to-noise problems for long-period components, their evaluations of source parameters are generally robust against structural heterogeneities in comparison to first-motion solutions.

Along the Nankai Trough, megathrust earthquakes have repeatedly occurred at intervals of 100–150 years (e.g. Ando 1975). Evaluating seismicity around this region is important for contributing to the understanding of megathrust earthquakes, such as evaluating stress accumulation/release processes on plate boundaries. In Japan, regular and slow earthquakes have been systematically monitored by the seismic networks of the Monitoring of Waves on Land and Seafloor (MOWLAS; https://doi.org/10.17598/NIED.0009) operated by the National Research Institute for Earth Science and Disaster Resilience (NIED; Okada et al. 2004). According to the combined earthquake catalogues of the International Seismological Centre-GLOBAL Earthquake Model (ISC-GEM; Storchak et al. 2013), the Japan Meteorological Agency (JMA), and the NIED F-net (Fukuyama et al. 1998, Kubo et al. 2002), the seismicity of regular earthquakes along the Nankai Trough, especially interplate earthquakes, is quite low. Figure 1 shows the spatial distribution of regular earthquakes with moment magnitudes ($M_w$) of 4.3–6.5 that occurred from April 2004 to August 2019, as listed in the F-net moment tensor (F-net MT) catalogue. The regional moment tensor inversion can be applied to earthquakes with $M_w$ > about 4, which is smaller than a lower limit of teleseismic moment tensor inversion (e.g., Figure 5 of Ekström et al. 2012). This is an advantage for discussing detail seismicity in a certain region. A few shallow offshore earthquakes occurred in the Tonankai and Nankai regions, and their focal mechanisms in the F-net catalogue were not characterised by low-angle thrust faulting. In other words, no earthquakes suggesting faulting
on the plate boundary around the Tonankai and Nankai regions are listed in the F-net MT
catalogue.

On 1 April 2016, the $M_w$ 5.8 earthquake, called “2016 southeast off the Kii Peninsula
earthquake”, occurred in the Tonankai region (marked A in Figure 1). The F-net MT solution
of this earthquake was characterised by high-angle ($38^\circ$) reverse faulting below the upper
surface of the Philippine Sea Plate, indicating it was an intraslab earthquake. However, a
detailed analysis of this earthquake revealed that it could be modelled by low-angle thrust
faulting at a depth of approximately 10 km, suggesting seismic slip along the plate boundary
(e.g. Nakano et al. 2018a, Takemura et al. 2018a, Wallace et al. 2016). Source models
suggested in these studies were also consistent with a model based on observed tsunami data
(Kubota et al. 2018). In regions with a thick accretionary prism, characteristics of surface
wave propagation are significantly affected by a low-velocity accretionary prism (e.g.
Gomberg 2018, Kaneko et al. 2019, Shapiro et al. 1998). Thus, the focal mechanisms of other
offshore earthquakes along the Nankai Trough could be incorrectly estimated using
conventional 1D regional MT inversion, even for long-period displacements. Indeed, shallow
very low frequency earthquakes along the Nankai Trough have been interpreted as low-angle
thrust faulting on the plate boundary by using offshore seismic observations (e.g. Nakano et
al. 2018b), but their focal mechanisms based on 1D analysis of onshore observations were
high-angle reverse faulting mechanisms within the accretionary prism (e.g., Ito & Obara
2006). To evaluate seismic activity along the Nankai Trough more precisely, offshore
earthquakes listed in the previous 1D catalogues require re-analysis.

Parallel simulation codes of seismic wave propagation (e.g. Gokhberg & Fichtner 2016,
Maeda et al. 2017) and 3D seismic velocity structure models (e.g. Eberhart-Phillips et al.
2010, Koketsu et al. 2012) enable the simulation of Green’s functions propagating through
realistic 3D Earth models (hereafter called ‘3D Green’s functions’), which have been used to
develop CMT inversions (e.g. Hejrani et al. 2017, Lee et al. 2013, Okamoto et al. 2018,
Although the resolution of detailed source characteristics for offshore earthquakes derived
using the 3D CMT method and onshore seismograms is limited compared to those using
offshore observations, these methods provide similar focal mechanisms and centroid
locations (see Figure 2 of Takemura et al. 2018b). Thus, offshore seismic activity, including
earthquakes before offshore seismic observations, can be effectively evaluated.

To investigate the decade-scale seismicity of offshore earthquakes along the Nankai
Through, we re-evaluated focal mechanisms based on CMT inversion using 3D Green’s function datasets, which were evaluated by numerical simulations of seismic wave propagation in a regional 3D velocity structure model. Then, to investigate spatial variation in slip behaviours on the plate boundary along the Nankai Trough, we compared the spatial distribution of focal mechanisms based on the 3D CMT technique with the spatial distribution of slip-deficit rates (Noda et al. 2018), slow slip events (SSEs; Kobayashi 2014, Miyazaki et al. 2006, Nishimura et al. 2013, Takagi et al. 2016, 2019, Yokota & Ishikawa 2020), shallow low-frequency tremors (LFTs; Yamashita et al. 2015), shallow very low-frequency earthquakes (VLFEs; Takemura et al. 2019b), and the 1968 Hyuga-nada earthquake (Yagi et al. 1998).

2. Data and Methods

We used three-component (NS, EW, and UD) velocity seismograms from F-net (NIED 2019), for which the performance of the sensors have been systematically monitored (Kimura et al. 2015). To conduct CMT inversion of the target earthquakes, we applied a band-pass filter with passed periods of 25–100 s. We selected a 25–100 s period band because ground motions for periods of 8–20 s are significantly affected by internal structures of the accretionary prism along the Nankai Trough (e.g. Takemura et al. 2019a). The selected period band is enough longer than corner periods of source spectra for target earthquakes. In our CMT inversions, we used 10-min F-net velocity seismograms from three minutes before the initial origin minute to conduct pre-processing (filter and integration) stably. We obtained displacement waveforms by calculating time integration of each filtered velocity record. The target earthquakes occurred within the region of assumed source grids (grey crosses in Figure 2) between April 2004 and August 2019, and values of Mw in the F-net catalogue ranging from 4.3 to 6.5. According to the signal-to-noise ratios for the target period band, the magnitude range of the analysed earthquake was determined by trial and error. Source grids were uniformly distributed at horizontal intervals of 0.1°. Depths of source grids ranged from 6 to 50 km at an interval of 2 km. The total number of source grids was 61,433.

Green’s functions were evaluated by solving equations of motion in the 3D viscoelastic medium model based on the finite-difference method (FDM) simulations. The 3D simulation model covered an area of $900 \times 1,000 \times 100$ km$^3$, which was discretised by grid intervals of 0.5 km in the horizontal direction and 0.2 km in the vertical direction. We used a parallel simulation code of OpenSWPC (Maeda et al. 2017), which includes the reciprocal calculation.
mode for effectively evaluating Green’s functions. The reciprocal calculation has proved very
useful in the case that the number of seismic source grids is significantly larger than the
number of seismic stations (e.g. Eisner & Clayton 2001, Hejrani et al. 2017, Okamoto et al.
2018). We obtained a total of approximately 35,000,000 Green’s function SAC files from
61,433 source grids to 32 F-net stations (black and blue filled triangles in Figure 2) via 96
reciprocal FDM calculations. The source time function of each Green’s function was the
Küpper wavelet with a duration of 1 s.

The 3D velocity model of Koketsu et al. (2012) was used, as it has been widely applied in
studies of seismic ground motions across Japan. The configurations of the subducting oceanic
plate and the Moho discontinuity are consistent with other models (e.g., Hirose et al. 2008,
Shiomi et al. 2006). The oceanic crust of the model of Koketsu et al. (2012) has
approximately 7 km thickness, which corresponds to those by seismic surveys (e.g.,
Nakanishi et al. 2002). The topography model in our simulations was the ETOPO1 model
(Amante & Eakins 2009). The P- and S-wave velocities and density ($V_P$, $V_S$ and $\rho$) in the
seawater layer were 1.5 km/s, 0.0 km/s and 1.04 g/cm$^3$, respectively. The air column was
modelled as a vacuum with $V_P$ of 0.0 km/s, $V_S$ of 0.0 km/s and $\rho$ of 0.001 g/cm$^3$. The
minimum $V_S$ in the solid column of 1.5 km/s was assumed. The accretionary prism is
important for constraining centroid depth, but detail velocity structure within the accretionary
prism has limited effects on long-period (> 20 s) seismograms (Figures 5 and 6 of Takemura
et al. 2019a).

Simulations were conducted using the computer system of the Earthquake and Volcano
Information Center at the Earthquake Research Institute, the University of Tokyo. Each
simulation required 385 GBytes of computer memory and a wall-clock time of 2.5 hours and
was performed using parallel computing with 432 cores to evaluate seismic wave propagation
of 200 s with 20,000 time-step calculations. According to our grid and model settings, our
FDM simulation can precisely evaluate long-period (> 10 s) seismic wave propagation.

Examples of Green’s functions are illustrated in the right panels of Figure 2. The source
(red star) was located at a depth of 10 km, near the plate boundary. We employed the
Cartesian coordinate system of Aki & Richards (2002), where $x$, $y$, and $z$ are taken as north,
east, and down, respectively. Due to the low-velocity accretionary prism and seawater,
durations of surface waves were amplified and elongated. In particular, for $M_{sw} = 1.0$ (i.e. a
pure strike-slip with strike angle of 0°, dip angle of 90°, and rake angle of 0°), Love waves on
horizontal components were strong and long. We assumed six-element moment tensors for
the CMT inversions, which includes five double couple and isotropic moment tensors (e.g. Kikuchi & Kanamori 1991).

In the CMT inversions, we basically used Green’s functions at F-net stations within epicentral distances of 100–400 km from the initial epicentre. The initial epicentre was obtained from the F-net MT catalogue. In cases where earthquake $M_w < 4.5$, we selected a distance range of 100–350 km due to the signal-to-noise ratio of the observed waveforms for the analysed period. We visually checked the filtered displacement waveforms and discarded noisy ones. Centroid location and time of the analysed earthquake were determined using grid search inversion. Because the analysis period range was longer than the source durations of target earthquakes with $M_w = 4.3–6.5$, we did not estimate source durations of these events. A set of Green’s functions at the source grids, which were located in a $\pm 0.4^\circ$ region from the initial epicentre and were distributed at depths of 6–50 km, was selected for the grid search inversion.

The CMT inversions were conducted for each selected source grid every 1 s from three minutes before the origin minute as recorded in the F-net catalogue. We used a 200-s time window for each CMT inversion. During grid search CMT inversion, we did not allow time shifts between synthetic and observed seismograms. After CMT inversion at all of the selected source grids, we obtained seismic moments and focal mechanisms at all locations and times. To identify the optimal solution, we evaluated variance reductions (VRs) between the observed and synthetic displacement seismograms for periods of 25—100 s. The VR could then be evaluated using the following equation:

$$
VR = \left[ 1 - \frac{\sum_{i=1}^{N_S} \int \left( u_{i}^{obs.}(t) - u_{i}^{syn.}(t) \right)^2 dt}{\sum_{i=1}^{N_S} \int \left( u_{i}^{obs.}(t) \right)^2 dt} \right] \times 100 \% \tag{1}
$$

where $N_S$ is the number of stations and $u_{i}^{obs.}$ and $u_{i}^{syn.}$ are the time-series of observed and synthetic displacements, respectively. If observed and synthetic seismograms are perfectly matched, VR is 100 %. The solution with the maximum VR was considered the optimal solution, providing the optimal centroid location, depth, time, focal mechanism, and seismic moment of each earthquake. In the case that the optimal solution was located at the edges of the initial source grids, we performed the CMT inversion again using Green’s functions for a broader source grid dataset. In the cases of regions around the edges of all source grids (all crosses in Figure 2), such as southern Kyushu and eastern Izu, we could not extend the grid set, and then the optimal solution was located at the grid edge. These events may include possibilities of some shifts outside the edges of the assumed source grids. Our grid search
CMT inversion required approximately 15–20 minutes using a typical, single-core desktop machine.

3. Results

We obtained a total of 215 CMT solutions for moderate earthquakes that occurred between April 2004 and August 2019. We discarded the solutions with a maximum VR of less than 20%. Our 3D CMT catalogue is listed in Global CMT (GCMT) format in the Supplementary data (Table S1) and the CSV format full catalogue data is available from https://doi.org/10.5281/zenodo.3674161. The size distribution and magnitude-time diagram of our 3D CMT catalogue are shown in Figures S1 and S2. The estimated moment magnitudes were slightly changed from the original F-net catalogue. The VRs of earthquakes with small magnitudes tended to be low (Figure S2) due to the signal-to-noise ratio for the analysed period range. We also compared our results with the GCMT catalogue (Figure S1).

Teleseismic CMT inversion is robust, but our regional CMT catalogue contains more earthquakes, whose $M_w$ values are less than about 5.

Figures 3 and 4 show examples of CMT solutions for the southeast off the Kii Peninsula earthquake (1 April 2016) and the Hyuga-nada earthquake (9 May 2019), respectively. In our previous study (Takemura et al. 2018a), the 2016 southeast off the Kii Peninsula earthquake was also analysed. The epicentre location and origin time were fixed in the previous study. We re-analysed this earthquake via full 3D CMT inversion, which estimates centroid location, depth, time, and moment tensor. The F-net MT solution of this earthquake was a high-angle (38°) reverse faulting mechanism (grey focal sphere in Figure 3). Its optimal solution is an $M_w$ 5.9 low-angle (10°) thrust faulting at a depth of 10 km (Figure 3), which is close to the plate boundary (e.g. Kamei et al., 2012; Park et al., 2010). The synthetic seismograms of the optimal solution corresponded well with the observations. The depth variation of VRs illustrated a clear peak around the optimal depth. The centroid depth of this earthquake was well constrained by our CMT inversion. Takemura et al. (2018a) numerically demonstrated that the low-velocity accretionary prism just above the seismic source—which controls long-period surface wave propagation—provides a better constraint on the centroid depth. They also demonstrated that the 3D oceanic plate has an important role in constraining focal mechanism (Figures 7 and 8 of Takemura et al. 2018a).

The centroid location was also close to that estimated by ocean-bottom seismometers deployed just above the source region (Nakano et al. 2018a, Wallace et al. 2016), while the
GCMT solution was slightly (0.2°) shifted to the south (Figure S3). The CMT result was consistent with models estimated by offshore observations (Kubota et al. 2018, Nakano et al. 2018a, Wallace et al. 2016). Especially, by using travel times, tsunami, and afterslip records, Wallace et al. (2016) and Nakano et al. (2018a) concluded that this earthquake could be interpreted as an interplate earthquake. Our CMT solution based on 3D Green’s functions and onshore seismograms also suggests this earthquake was due to faulting on the plate boundary.

Figure 4 shows the results of the CMT inversion and waveform fitting for the Hyuga-nada earthquake on 9 May 2019. The F-net MT solution was also high-angle (33°) reverse faulting. The optimal CMT solution indicated an Mw 6.2 low-angle (16°) thrust mechanism. The dip angle from the CMT solution agreed well with that of the Philippine Sea Plate around this earthquake. The synthetic waveforms also corresponded well to observed ones. Although the optimal depth (26 km) was determined to be close to the upper surface of the Philippine Sea Plate (approximately 27 km), a high VR (> 80%) area was found within a wider depth range (16–32 km). Because the depth of this earthquake was deeper than the 2016 southeast off the Kii Peninsula earthquake, the effects of the low-velocity accretionary prism might not have been so strong. Thus, the depth resolution of the CMT solutions might not be good when compared to the case of the 2016 southeast off the Kii Peninsula earthquake. To constrain the hypocentre depth more sharply, additional data, such as shorter-period (~ 4 s) first-arrival P-wave waveforms, would need to be considered (e.g. Okamoto et al. 2018, Takemura et al. 2018a, Wang & Zhan 2020).

Figure 5 shows a comparison of the estimated focal mechanisms for the F-net and our 3D CMT catalogues. Our CMT solutions of onshore earthquakes did not differ significantly from the F-net solutions. However, our offshore CMT solutions differ from those based on the F-net 1D analysis. In particular, dip angles and centroid depths of offshore earthquakes—which are important for distinguishing interplate and intraslab earthquakes—were different. Differences in dip angles and depths were clearly illustrated in detailed comparisons of seismicity southeast off the Kii Peninsula and the Hyuga-nada (Figures 6, 7, S4, and S5). The conventional 1D CMT inversion poorly estimated the dip angles of offshore earthquakes that occurred outside of onshore seismic arrays due to the lack of the 3D subducting oceanic plate and the accretionary prism (e.g. Takemura et al. 2018ab). The comparisons and error estimations of dip angles are illustrated in Figures 9 and 10. The comparison of spatial distributions of CMT solutions with the GCMT catalogue is also illustrated in Figure S3.

We focused our attention on seismicity southeast off the Kii Peninsula and the Hyuga-nada
(local names are illustrated in Figure 1), where seismic activities are relatively high in the Nankai subduction zone. Figure 6 shows spatial distributions of the CMT solutions southeast off the Kii Peninsula. We also plotted shallow VLFEs in the catalogue of Takemura et al. (2019b) as grey focal spheres. Shallow VLFEs, which were characterised by low-angle thrust faulting, were concentrated near the trench. In the region with shallow VLFE active, low-angle thrust type CMT solution at depths of 5-10 km, which suggests seismic slip on the plate boundary, was not estimated. On the down-dip side of the shallow VLFE region, a low-angle thrust faulting mechanism was estimated at a depth near the plate boundary (along with profile A in Figure 6). This earthquake is the 2016 southeast off the Kii Peninsula earthquake (Figure 3). Almost all of the other earthquakes plotted in Figure 6 are aftershocks of the 2004 Mw 7.5 intraslab earthquake that occurred on 5 September 2004 southeast off the Kii Peninsula. Our CMT solutions of these aftershocks were separately distributed at two depths within the oceanic crust and mantle (10–15 and 20–30 km depths). This separation corresponded well to the hypocentre depth distributions of the aftershocks of the Mw 7.5 earthquake as determined using ocean-bottom seismometers (e.g., Nakano et al. 2015, Sakai et al. 2005). On the other hand, almost all the centroid depths of the F-net solutions were concentrated within the accretionary prism, crust and oceanic crust (5–15 km depths; Figure S3). According to comparisons with the detail hypocentre distributions in this region, even for earthquakes near the trough axis, our CMT method provided better constraints for centroid depths, compared to the 1D F-net MT solutions.

Figure 7 shows the spatial distribution of CMT solutions around the Hyuga-nada region. Our CMT solutions characterised by low-angle thrust faulting mechanisms were distributed across the region with average slip rates of approximately 20–40 mm/yr as inferred from small repeating earthquakes (e.g., Yamashita et al. 2012). The optimal centroid depths of such thrust solutions were concentrated around the plate boundary (profiles B and C in Figure 7). The distribution of our CMT solutions agreed with that derived from onshore and temporal offshore seismometers (Tahara et al. 2008). The 3D CMT solutions, especially for depths and low-angle thrust faulting mechanisms, corresponded to the areas of detected repeating earthquakes (Yamashita et al. 2012). Our 3D CMT also worked well in this region. The dip angles of the F-net MT solutions at depths around the plate boundary were slightly higher than those of the plate boundary, as shown in Figure S4. The centroid depths of the F-net catalogue were also slightly deeper than the depths of the plate boundary. These also might have been due to a lack of 3D geometry of the subducting oceanic plate in the 1D analysis.
To evaluate differences between 3D CMT and F-net MT solutions, we calculated correlation coefficients of P-wave radiation patterns (e.g., Helffrich 1997, Kuge & Kawakatsu 1993) between two catalogues. We also calculated differences of estimated centroid depths from corresponding F-net solutions. Figure 8 shows the spatial distribution of correlation coefficients of P-wave radiation patterns and depth differences between our CMT and F-net MT catalogues. The values of correlation coefficients of offshore earthquakes (enclosed by dashed lines in Figure 8a) were widely distributed. Centroid depths of offshore earthquakes were also different from those of the F-net catalogue. Other earthquakes, which occurred in onshore regions or had good station coverages, have high similarities and small depth differences. The rigidity of the 3D model complicatedly depends on the centroid location, and consequently, depth shifts could cause shifts of moment magnitudes from the F-net catalogue (Figure S1). We also plotted comparisons of estimated depths with corresponding GCMT solutions in Figure 3c. Almost of our centroid depths were 0-5 km shallower than those in the GCMT catalogue.

The parameter of the dip angle is important for distinguishing earthquake types. In the Nankai subduction zone, because megathrust earthquakes have repeatedly occurred, seismicity of interplate earthquakes is important. We selected low-angle thrust faulting solutions at depths around the plate boundary from our 3D CMT catalogue. These selected events could be interpreted as seismic slips on the plate boundary. Figure 9 shows a comparison of dip angles between the Philippine Sea plate and suggestive interplate earthquakes from the 3D CMT catalogue. We also compared dip angles of corresponding earthquakes in the F-net MT and GCMT catalogues. Although dip angles of F-net catalogues were higher angles compared to the Philippine Sea plate, our CMT solutions well correlated with dip angles of the plate boundary. Dip angles of the GCMT catalogue roughly corresponded to dip angles of the Philippine Sea plate, but our solutions showed better agreements with the plate dip angles. The teleseismic CMT solutions are generally robust, but regional 3D CMT could provide better constraints of dip angles.

We also calculated the VRs between observed and synthetic displacement waveforms to discuss estimation errors of dip angles for offshore earthquakes. Synthetic displacement waveforms were calculated from 3D Green’s functions, assuming double-couple point sources and fixing hypocentre locations and seismic moments. Figure 10 shows spatial distributions of VRs for the 2016 southeast off Kii Peninsula earthquake and the 2019 Hyuganada earthquake. Clear trade-offs between strike and rake angles appeared in the strike-rake
plane (upper panels). We confirmed that higher VR values (> 75 %) only appeared in the regions with dip angles of 5-15° and 10-20° for both earthquakes. Thus, our 3D CMT provides constraints of dip angles with uncertainties of approximately ±5°.

4. Discussion

4.1. Slip behaviours on the plate boundary along the Nankai Trough

In order to discuss slip behaviour on the plate boundary, Figure 11 shows the spatial distribution of slow earthquakes and earthquakes with low-angle (< 25°) thrust faulting solutions at depths around the plate boundary along the Nankai Trough. The large coseismic slip area of the 1968 Mw 7.5 earthquake (Yagi et al. 1998) is indicated by the blue area in Figure 11. The cumulative deep SSE slips in each grid were determined by summing slip of each SSE in each catalogue (Nishimura et al. 2013, Takagi et al. 2016, 2019). We then evaluated the SSE slip rates by dividing the cumulative SSE slip at each grid by the analysis period of each catalogue. The SSE slip rate indicates the activity of deep slow earthquakes. We did not calculate SSE slip rates for shallow SSEs reported by Yokota & Ishikawa (2020) because the number of detected events was still too low at each region. For similar reasons, the long-term SSEs off the Kii Channel (Kobayashi 2014) and Tokai (Miyazaki et al. 2006) regions were also excluded from the SSE slip rate calculation. Thus, we plotted the fault configurations or large slip areas of long-term SSEs and shallow SSEs. We also plotted shallow LFTs (Yamashita et al. 2015) and shallow VLFEs (Takemura et al. 2019c) as indicators of shallow slow earthquake activity. The spatial distribution of slip-deficit rates from GNSS and GNSS-A observations by Noda et al. (2018), is plotted using blue contour lines in Figure 11.

At deeper depths (30–40 km), deep slow earthquakes were active, especially in areas with high SSE slip rates, but no interplate regular earthquakes were found. Although SSEs were not removed in the slip-deficit rate estimation of Noda et al. (2018)—except for long-term SSEs at the Bungo Channel—the regions with deep SSEs were characterised by low (20–40 mm/y) slip-deficit rates. At shallower depths (< 30 km) in the offshore region, regular earthquakes, slow earthquakes, and high (> 60 mm/y) slip-deficit zones were separated from each other. Similar separations of the repeating earthquakes, slow earthquakes, and large coseismic slip areas of megathrust earthquakes at shallower depths were observed in the regions of Tohoku (e.g., Nishikawa et al. 2019), Central Ecuador (e.g., Vaca et al. 2018) and Costa Rica (e.g., Dixon et al. 2014). In particular, Nishikawa et al. (2019) pointed out that
slow earthquakes were complementarily distributed in the regions surrounding the large
coseismic slip area of the 2011 Mw 9.0 Tohoku earthquake. Takemura et al. (2019c) pointed
out that shallow, slow earthquakes cluster or migrate due to the existence of pore fluid in the
transitional regions between high-strength and low-strength zones of the plate boundary.
According to these previous studies and our observations, we suggest that the observed
separation between slip behaviours on the plate boundary along the Nankai Trough are
related to the heterogeneous distribution of effective strengths on the plate boundary, which is
controlled by the frictional coefficient, pore fluid pressure, and normal stress.

4.2. Regional 3D CMT inversions for the Mw 7.2 and 7.5 earthquakes southeast off the
Kii Peninsula

We conducted 3D CMT inversions of offshore earthquakes with Mw of 4.3–6.5. During
the analysis period (April 2004 to August 2019), Mw 7.2 and 7.5 intraslab earthquakes occurred
southeast off the Kii Peninsula on 5 September 2004. Because typical Mw 7 class earthquakes
have rupture durations of 30–50 s and fault areas of 1000–5000 km² (e.g. Kanamori & Brodsky
2004), precise source parameter estimation for such earthquakes is difficult based on our
assumptions of the CMT inversion. Despite these disadvantages, the rapid estimation of the
CMT solution for these large earthquakes is important for disaster mitigation, such as a CMT-
based tsunami warning system. We, therefore, tested our simple CMT inversion for the Mw 7.2
and 7.5 southeast off the Kii Peninsula earthquakes. Because amplitude saturation of F-net
broadband seismometers occurs for regional large earthquakes, we used F-net strong motion
seismometers, which have a large clip level and similar frequency response to STS-2
seismometers for periods less than 100 s. We selected F-net stations with distances of 200–500
km from the initial epicentre, which were slightly farther than for the original CMT settings
(100–400 km).

Figures 12 and 13 show the results of CMT inversions for the Mw 7.2 and 7.5 earthquakes
southeast off the Kii Peninsula, respectively. Detailed estimated parameters are also listed in
Table S2. Signal-to-noise ratios were high enough compared to smaller (Mw < 4.5)
eartquakes in this study but the VRs were low compared to those of moderate earthquakes.
The synthetic waveforms roughly corresponded to the observed ones (Figures 12b and 13b).
Due to the assumptions of a point source and simple-source time function, detailed
characteristics of the observed waveforms were not successfully reproduced. Furthermore,
the high (> 66%) VR areas were wider than the CMT results for moderate earthquakes within
the same region (Figure 3). The estimated deviatoric components were very similar to those
in the GCMT catalogue, but, especially in the result of the Mw 7.2 earthquake, a large
isotropic component appeared. Waveform fitting and large non-double couple components
suggest the likely complexity of the rupture processes and the source extents for the Mw 7.2
and 7.5 earthquakes. Estimated moment magnitudes were slightly smaller than those of the
GCMT catalogue as a result of the analysed period and the deeper centroid depths. Our
analysis periods were not significantly larger than the rupture durations of Mw 7 earthquakes,
leading to size underestimation. However, the regional 3D CMT method provides better
constraints of dips and depths for offshore earthquakes compared to 1D CMT systems
(Figures 3, 4, 5, 6, and 9), and our 3D grid search required only 15-20 minutes. These points
show the advantages for CMT-based tsunami prediction systems (e.g. Inazu et al. 2016,
Reymond et al. 2012). To obtain more accurate solutions, the CMT method with various
durations (e.g., Takemura et al. 2019b) or deconvolution method (e.g., Vallée et al. 2011)
should be implemented. Such sophisticated methods require more time to obtain solutions.

We compared our CMT result for the Mw 7.2 earthquake with the finite-fault model
(Okuwaki & Yagi 2018) conducted using teleseismic records based on Yagi & Fukahata
(2011). Our horizontal centroid location was very close to an area with large (> 3 m)
coseismic slips (Figure 14). The horizontal locations of the dominant slip and centroid
locations of the 3D CMT solution were shared. The centroid location also agreed with that
estimated by tsunami record (Satake et al. 2005). Thus, we think that the centroid location of
the Mw 7.2 earthquake was well constrained by our 3D CMT method. The depths of large
coseismic slips in the finite fault model ranged from 9 to 18 km, but the optimal centroid
depth of the 3D CMT inversion was 26 km. The depth difference could be due to the regional
3D heterogeneities (accretionary prism, bathymetry change, and subducting plate). According
to the hypocentre determinations derived using ocean-bottom seismometers (Nakano et al.
2015, Sakai et al. 2005), the hypocentres of aftershocks due to the Mw 7.5 earthquake were
distributed at depths of approximately 10–30 km. We also tested the centroid depth and large
isotropic components by our 3D CMT inversion. By using simulated seismograms of the
finite-fault model (Okuwaki & Yagi 2018) as observed seismograms, we conducted CMT
inversion of the simulated Mw 7.2 intraslab earthquake (Figure S6). The centroid location and
depth well corresponded to the large slip area of the finite-fault model. The large isotropic
component was also estimated. Thus, large non-double couple components suggest the likely
complexity of the rupture processes and the source extents for the finite-fault model of the
Mw 7.2 earthquake. Based on the hypocentre distribution of aftershocks, the fault dimensions
of the Mw 7.2 earthquake, and synthetic test, we considered that the extension of seismic slips
at depths of approximately 26 km might be possible.

The detailed rupture processes of the $M_w$ 7.2 and 7.5 earthquakes remain unclear. The regional seismic data and 3D Green’s functions may provide additional constraints for large offshore earthquakes. The finite fault modelling based on the 3D Green’s functions is an important but challenging issue that requires particular attention in future studies.

5. Conclusion

We conducted 3D CMT inversions of moderate earthquakes along the Nankai Trough using the regional 3D Green’s function dataset. By comparing 3D CMT solutions with those in the F-net catalogue, large differences in focal mechanisms and centroid depths were found for offshore earthquakes. These differences could be caused by 3D offshore heterogeneities, such as the low-velocity accretionary prism and subducting Philippine Sea plate. Onshore MT inversion using a simple 1D Earth model could provide incorrect estimations due to offshore heterogeneities and station coverage. By introducing the effects of such 3D heterogeneities, the 3D CMT solutions for offshore earthquakes practically agreed with hypocentre distributions determined by ocean-bottom seismometers. Furthermore, our CMT method based on onshore seismograms provided better constrained focal mechanisms and centroid depths compared to the F-net MT catalogue. We also compared our CMT solutions with those of the GCMT catalogue. The teleseismic CMT solutions are generally robust but regional 3D CMT could provide better constraints of dip angles. The regional 3D CMT catalogue contains more earthquakes compared to the GCMT catalogue, where only earthquakes with about $M_w > 5$ are listed. To investigate detailed decade-scale seismicity in a certain region, CMT inversion incorporating regional 3D velocity model should be required.

Although no suggestive interplate earthquakes are listed in the 1D catalogue, some low-angle thrust faulting solutions at depths around the plate boundary were confirmed by our 3D CMT catalogue. These earthquakes could be interpreted as interplate earthquakes. By using our 3D CMT catalogue and previously published slow earthquake models, we illustrated the spatial distribution of slip behaviours on the plate boundary along the Nankai Trough.

Regular interplate earthquakes and slow earthquakes occur within different segments on the plate boundary. These separated distributions might reflect the heterogeneous distribution of effective strength on the plate boundary. The gap zones, where no regular interplate and slow earthquakes occurred, were found in the Nankai, Tonankai, and Tokai regions. These were the regions with large (> 60 mm/y) slip-deficit rates, where the plate boundary can be strongly
The regional CMT inversion of earthquakes with $M_w > 7$ was generally difficult due to their fault size and the amplitude saturation of the broadband sensors. CMT inversions for the 2004 $M_w 7.2$ and $7.5$ intraslab earthquakes southeast of the Kii Peninsula were performed using the regional broadband strong motion sensors of F-net. Although signal-to-noise ratios of the observed displacements were good enough, the waveform fittings of the $M_w 7.2$ and $7.5$ intraslab earthquakes were not good compared to those of typical moderate earthquakes due to fault sizes and the rupture complexity. However, the centroid location agreed with that estimated by tsunami record, and the focal mechanism could be constrained. These points and the rapid availability of a solution could be the advantages for CMT-based tsunami warning systems.

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F-net waveform data and the F-net MT catalogue are available via the website of the National Research Institute for Earth Science and Disaster Resilience ([https://doi.org/10.17598/NIED.0005](https://doi.org/10.17598/NIED.0005)). Bathymetric depth data was obtained from ETOPO1 (Amante & Eakins 2009). OpenSWPC software (Maeda et al. 2017) and the 3D model of Koketsu et al. (2012) were obtained from [https://github.com/takuto-maeda/OpenSWPC](https://github.com/takuto-maeda/OpenSWPC) and [https://www.jishin.go.jp/evaluation/seismic_hazard_map/lpshm/12_choshuki_dat/](https://www.jishin.go.jp/evaluation/seismic_hazard_map/lpshm/12_choshuki_dat/), respectively. Generic Mapping Tools (Wessel et al. 2013) and Seismic Analysis Code (SAC; Helffrich et al. 2013) were used to make the figures and when conducting the signal processing work, respectively. The catalogue of slow earthquakes (Nishimura et al. 2013, Takagi et al. 2016, Yamashita et al. 2015) was downloaded from the Slow Earthquake Database website ([Kano et al. 2018; http://www-solid.eps.s.u-tokyo.ac.jp/~sloweq/](http://www-solid.eps.s.u-tokyo.ac.jp/~sloweq/)). Our CMT catalogue and CMT results of assumed source grids for each earthquake are available from [https://doi.org/10.5281/zenodo.3674161](https://doi.org/10.5281/zenodo.3674161). The FDM simulations of seismic wave propagation were conducted on the computer system of the Earthquake and Volcano Information Center at the Earthquake Research Institute, the University of Tokyo. This study was supported by the Japan Society for the Promotion of Science (JSPS) KAKENHI Grant Numbers 17K14382 under the Grant-in-Aid for Young Scientists (B) and 19H04626 in Scientific Research under Innovative Areas ‘Science of Slow Earthquakes’. We also thank Dr. J. Braunmiller, an anonymous reviewer and Editor Prof. J. Virieux for careful reviewing and constructive comments, which have helped to improve the manuscript.

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**Figures**

Figure 1. Map of the study region. The black contour lines are the iso-depth contour lines of the upper surface of the Philippine Sea Plate of Koketsu *et al.* (2012). Focal mechanisms are the moment tensor (MT) solutions of regular earthquakes with *M*$_w$ of 4.3–6.5 in the F-net catalogue (Fukuyama *et al.* 1998, Kubo *et al.* 2002) that occurred in the area with latitudes less than 34.8°N, longitudes greater than 131°E, and at depths of less than 50 km. The plotted MT solutions range from April 2004 to August 2019. The blue hatched areas represent the expected source region of the Nankai and Tonankai earthquakes (Earthquake Research Committee, 2001, available at: [http://www.jishin.go.jp/main/chousa/01sep_nankai/index.htm](http://www.jishin.go.jp/main/chousa/01sep_nankai/index.htm)). The earthquake marked A is the *M*$_w$ 5.8 southeast off Kii Peninsula earthquake that occurred on 1 April 2016.
Figure 2. Calculation settings used in this study were the blue dashed line represents the horizontal coverage of the simulation model region. The triangles and crosses in the map denote the locations of the F-net stations and the assumed source grids, respectively. Green’s functions from the source grids to the black-fill and blue-fill triangles were evaluated via reciprocal calculations using OpenSWPC code (Maeda et al. 2017). The right-hand panels show examples of filtered displacement Green’s functions from a certain hypocentre (red star, at a depth of 10 km) to the N.UMJF station (blue triangle), whose epicentral distance is 263 km. The filter passband was 25—100 s.
Figure 3. CMT results for the southeast off Kii Peninsula earthquake that occurred on 1 April 2016. (a) Locations of the optimal solutions, used stations, and depth variations of optimal solutions at each source grid. Colours of the focal mechanisms reflect values of variance reduction between observed and synthetic displacements in the 25–100 s period band. The numbers above the optimal solutions in (a) are the optimal centroid depths. The grey focal mechanism in (a) is the F-net MT solution of this earthquake. (b) Comparisons of observed and synthetic displacements in the 25–100 s period band. Grey solid and blue dotted lines are the observed and synthetic seismograms, respectively. Synthetic seismograms were evaluated by assuming the optimal solution. Amplitudes at each station were normalised by the maximum amplitude of both observed and synthetic three-component displacement waveforms. Detailed source parameters are listed in Table S1.
Figure 4. CMT results for the Hyuga-nada earthquake that occurred on 9 May 2019. (a) Locations of the optimal solutions, used stations, and depth variations of optimal solutions at each source grid. Colours of the focal mechanisms reflect values of variance reduction between observed and synthetic displacements in the 25–100 s period band. The numbers above the optimal solutions in (a) are the optimal centroid depths. The grey focal mechanism in (a) is the F-net MT solution of this earthquake. (b) Comparisons of observed and synthetic displacements in the 25–100 s period band. Grey solid and blue dotted lines are the observed and synthetic seismograms, respectively. Synthetic seismograms were evaluated by assuming the optimal solution. Amplitudes at each station were normalised by the maximum amplitude of both observed and synthetic three-component displacement waveforms. Detailed source parameters are listed in Table S1.
Figure 5. Comparisons of estimated CMT solutions between the (a) 3D CMT and (b) F-net MT catalogues. Colours of focal mechanisms represent the centroid depths of each solution. Detailed source parameters of our 3D CMT solutions are listed in Table S1. The regions enclosed by the dashed lines in (a) are enlarged in Figures 6 and 7.
Figure 6. Spatial distribution of the CMT solutions southeast of the Kii Peninsula. Coloured focal mechanisms are our CMT solutions. Grey focal mechanisms are the CMT solutions of shallow VLFEs (Takemura et al. 2019b). The right-hand panels show cross-sections along profiles A-A’, B-B’ and C-C’. The bathymetry of ETOPO1 (Amante & Eakins 2009), the upper surface, and oceanic Moho of the Philippine Sea Plate (Koketsu et al. 2012) along each profile are plotted in the right-hand panels.

Figure 7. CMT results for the Hyuga-nada region. Coloured focal mechanisms are our CMT solutions. The right-hand panels show cross-sections along profiles A-A’, B-B’ and C-C’.
The bathymetry of ETOPO1 (Amante & Eakins 2009), the upper surface, and oceanic Moho of the Philippine Sea Plate (Koketsu et al. 2012) along each profile are plotted in the right-hand panels.

Figure 8. Spatial distributions of (a) correlation coefficients (CCs) of P-wave radiation patterns between 3D CMT and F-net solutions and (b) depth differences of 3D CMT solutions from the F-net catalogue. Lower right panels in (a) and (b) show histograms of CCs and differences, respectively. Offshore earthquakes are defined as earthquakes that occurred within regions closed by dotted lines in (a).
Figure 9. Comparisons of dip angles between the Philippine Sea plate (PHS) and CMT solutions for suggestive interplate earthquakes. (a) Map of the region, comparisons of dip angles of the Philippine Sea plate with (b) CMT solutions of this study and (c) F-net MT solutions. The background colour in (a) represents the spatial distribution of dip angles of the Philippine Sea plate. The coloured circles denote dip angles of CMT solutions in this study. We compared dip angles between the Philippine Sea plate and (c) F-net MT and (d) GCMT solutions of corresponding earthquakes.
Figure 10. Distributions of variance reductions (VR) in the strike-rake and strike-dip planes for (a) the southeast off Kii Peninsula earthquake on 1 April 2016 and (b) the Hyuga-nada earthquake on 9 May 2019. In synthetics of displacement seismograms with various strike, dip, and rake, we assumed pure double-couple point sources and fixed hypocenter locations and seismic moments from CMT results (Figures 3 and 4).
Figure 11. Spatial distribution of slip behaviours on the plate boundary along the Nankai Trough. Plotted focal mechanisms are low-angle thrust faulting solutions at depths around the plate boundary. The coseismic slip area of the 1968 $M_w$ 7.5 Hyuga-nada earthquake (Yagi et al. 1998) is shaded in dark blue. SSE slip rates were evaluated from the combined SSE catalogues (Nishimura et al. 2013, Takagi et al. 2016, 2019). The pink circles indicate the epicentres of the shallow LFTs of the Hyuga-nada and the shallow VLFEs in the Tonankai region referred from Yamashita et al. (2015) and Takemura, Noda, et al. (2019). The pink shaded areas and pink rectangles indicate the large slip areas of long-term SSEs (Kobayashi 2014, Miyazaki et al. 2006) and shallow SSEs (Yokota & Ishikawa 2020), respectively. The blue contour lines indicate the slip-deficit rates [mm/yr] on the plate boundary by Noda et al. (2018).
Figure 12. CMT results for the $M_w$ 7.2 southeast off the Kii Peninsula earthquake that occurred on 5 September 2004. Grey focal mechanisms are the solutions of the F-net MT and GCMT catalogues. (a) Locations of the optimal solutions, used stations, and depth variations of optimal solutions at each source grid. Colours of the focal mechanisms reflect values of variance reduction between observed and synthetic displacements in the 25–100 s period band. The numbers above the optimal solutions in (a) are the optimal centroid depths. The grey focal mechanism in (a) is the F-net MT solution of this earthquake; (b) Comparisons of observed and synthetic displacements in the 25–100 s period bands. Grey solid and blue dotted lines are the observed and synthetic seismograms, respectively. Synthetic seismograms were evaluated by assuming the optimal solution. Amplitudes at each station were normalised by the maximum amplitude of both observed and synthetic three-component displacement waveforms. Detailed source parameters are listed in Table S2.
Figure 13. CMT results for the Mw 7.5 southeast off the Kii Peninsula earthquake that occurred on 5 September 2004. (a) Locations of the optimal solutions, used stations, and depth variations of optimal solutions at each source grid. Colours of the focal mechanisms reflect values of variance reduction between observed and synthetic displacements in the 25–100 s period bands. The numbers above the optimal solutions in (a) are the optimal centroid depths. The grey focal mechanism in (a) is the F-net MT solution of this earthquake; (b) Comparisons of observed and synthetic displacements in the 25–100 s period band. Grey solid and blue dotted lines are the observed and synthetic seismograms, respectively. Synthetic seismograms were evaluated by assuming the optimal solution. Amplitudes at each station were normalised by the maximum amplitude of both observed and synthetic three-component displacement waveforms. Detailed source parameters are listed in Table S2.
Figure 14. Comparison of the CMT results for the Mw 7.2 southeast off the Kii Peninsula earthquake and other CMT catalogues (Ekström et al. 2012, Fukuyama et al. 1998, Kubo et al. 2002) and finite fault modelling (Okuwaki & Yagi 2018) solutions. The bottom panel is the slip distribution of the finite fault model in the strike-depth plane.