Estimation of S-wave anisotropy in the Nankai Trough using active and passive seismic dataset observed by DONET

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In the Nankai Trough seismogenic zone, cabled real-time observation system, DONET and IODP C0002G borehole observatory, have been operating to monitor seismic activity, crustal deformation and tsunami propagation since Aug. 2011 and Jan. 2013, respectively. For elucidating dynamic processes from preseismic state to the generation of mega-thrust earthquake, which occurs repeatedly in subduction zones, it is important to observe and monitor the stress state, i.e., a key parameter governing its fault dynamics in the vicinity of seismogenic fault. In this study, we performed active and passive seismic data processing to obtain seismic anisotropy, as a proxy of stress state, by using dataset acquired by three-component seismometers installed in the DONET and IODP C0002G observatories. In the passive data processing, we applied a seismic interferometry method to ambient noise records acquired by horizontal components of each seismometer. After the application of cross-dipole analysis to the acquired records, several coherent events have become visible. These events were perceived to be reflected S-wave from each layer below seafloor, and S-wave splitting caused by seismic anisotropy was observed. We then estimated anisotropy direction and amplitude beneath each seismometer in shallow sediment layer. For the active seismic dataset that was acquired in azimuthally aligned airgun shot locations for each seismometer performed in November 2013, we steered horizontal records for each pair of shot and seismometer to form radial and transverse components to each shot location at a distance of ca. 3km from the seismometer. In the steered records, P-S converted waves from bottom of shallow sediments were clearly visible. The horizontal axis of symmetry to fast S-wave direction was estimated through the fitting of a simple sinusoidal curve to the axial amplitude distribution in radial and transverse components for S-wave anisotropy in the shallow sediments in the Nankai Trough. We finally compared the obtained S-wave anisotropy from the passive data with the active one. The comparison in the order of anisotropy and the orientation of the horizontal axis of symmetry showed good agreement with each other, especially in landward area. Some differences in the complicated structure zone were probably caused by the signal-to-noise ratio deterioration due to the influence of the dimensionality of the sub-seafloor structure and the seafloor topography around the observatory in the active dataset, and to the contamination of seafloor microseisms in the passive data. We now plan to develop a new scheme including layer stripping method, 3-D rotation method, etc., to improve the quality of our analysis.

1. INTRODUCTION

In the Nankai Trough subduction zone, Japan, Philippine-sea plate is subducting beneath the Eurasian plate with the rate of approximately 4.5 cm/year, and M 8.0 class huge mega-thrust earthquake occurred repeatedly with interval of 100 to 150 years (Seno, 1993). In this area, we are operating cabled seismic observation system, named DONET (Dense Oceanfloor Network systems for Earthquake and Tsunamis), which includes twenty three-components seismometers deployed on the seafloor (Kaneda et al., 2010). We also have a borehole observatory IODP (Integrated Ocean Drilling Program) C0002G, which have borehole seismometers, including three-components seismometers at the bottom of the borehole with depth of 900 mbsf (Kopf et al., 2011, Kimura et al., 2013) (Fig. 1). These seismometers were mainly distributed for passively monitoring natural earthquake related phenomena, e. g. regional microearthquakes, VLF (Very Low Frequency) events, seismic microtremors, etc. in the seismogenic zone.

For elucidating preparation and generation process of mega-thrust earthquake, which occurs repeatedly in subduction zones, it is important to observe and monitor the stress state, which is a key parameter governing its fault dynamics in the vicinity of seismogenic fault. In-situ stress analysis such as borehole breakout analysis may provide the
orientation and the order of differential stress around the borehole, but it is still challenging to drill seismogenic fault, and is even more difficult to monitor temporal change of stress state, especially in wide area. Therefore, we have to consider another method to estimate stress state. In this study, we performed active and passive seismic data processing to obtain seismic anisotropy using dataset acquired by three-component seismometers installed in the DONET and IODP C0002G observatories. Seismic anisotropy can be a proxy of stress state, and furthermore, its temporal change is expected to identify change of stress around the seismogenic fault.

2. METHOD AND DATA PROCESSING

We performed (1) passive and (2) active seismic data processing using dataset observed by DONET and IODP C0002G seismometers to obtain seismic anisotropy. Fig.2 shows the schematic of the passive and active survey. The details of each processing are described as follows.

(1) Passive seismic data processing

Seismic interferometry can retrieve the impulse response by the cross-correlation of seismic records simultaneously acquired by the two seismometers (Schuster, et al., 2004; Wapenaar and Fokkema, 2006). In this study, we applied seismic interferometry to ambient noise records acquired by horizontal components of each seismometer. Because the horizontal components are dominated by S-wave energy, we expected that obtained auto-correlation function (ACF) and cross-correlation function (CCF) would provide us the knowledge of S-wave velocity and anisotropy beneath seafloor. In obtained ambient noise records, predominant peak of microseisms ranges from 0.15 Hz to 2 Hz is clearly visible. Microseisms are major ambient noise which is generated by ocean swell loading the seafloor (Longuet-Higgins, 1950) and propagate in the seafloor as Stoneley wave (Kedar et al., 2008). In this study, we focused on body wave component. Therefore, band-pass filter with pass band of 2-10 Hz was used to suppress the amplitude of microseisms for the following processing.

Then we obtained zero offset 4-C ACF and CCFs calculated from horizontal components of each seismometer. In the x-y plane of the Cartesian coordinates, the 4-C records can be written as the following:

\[
V = \begin{pmatrix} v_{11} & v_{12} \\ v_{21} & v_{22} \end{pmatrix},
\]

where \( v_{ij} \) represents virtual shot records with \( i \)-direction source and \( j \)-direction receiver component. For example, \( v_{12} \) represents the virtual shot record with x-direction source and y-direction receiver obtained from cross-correlation between x- and y-component of the seismometer. We obtained ACF and CCFs calculated from each 1 hour dataset for more than one year, and stacked all results. We finally obtained stacked ACF and CCFs for each observatory.

To estimate anisotropy information from the obtained 4-C ACF and CCFs, we applied Alford rotation to the 4-C dataset. The Alford rotation is a widely used method to determine first and slow directions of S-wave anisotropy. The rotation can be performed as following equation (Alford, 1986).

\[
U = \begin{pmatrix} \cos^2 \theta v_{11} + \sin^2 \theta v_{22} & \cos \theta \sin \theta (v_{12} + v_{21}) \\ .5 \sin 2\theta (v_{12} + v_{21}) & .5 \sin 2\theta (v_{22} - v_{11}) \\ \cos \theta \sin \theta (v_{12} - v_{21}) & \cos^2 \theta v_{12} + \sin^2 \theta v_{22} \\ .5 \sin 2\theta (v_{22} - v_{11}) & -.5 \sin 2\theta (v_{12} - v_{21}) \end{pmatrix}
\]
We can obtain counter clockwise rotated 4-C data from equation (2). If the rotated angle $\theta$ is agreed with the direction of the S-wave anisotropy, the off-diagonal element of matrix (2) can be minimized. In practical, we calculated power of off-diagonal elements of the rotated 4-C matrix with the rotated angle changing and find the optimum value of the angle using time windows that include target reflection wave. The amplitude of the S-wave anisotropy can be calculated the time difference of the target reflection wave between V11 and V22.

Fig. 3 shows an example of 4-C ACF and CCFs, which were calculated from ambient noise records obtained by IODP C0002G borehole seismometer. In hourly 4-C records (Fig. 3(a)), coherent events are visible mainly in 2.0 to 5.0 s. A simple travel...
time calculation using the simple layered velocity model shown in Fig. 2, confirmed that these events should include S-wave, which are reflected from the seafloor and propagate in the shallow sediment layer. Fig. 3 (b) shows the stacked 4-C data calculated from the hourly 4-C records. We apply Alford rotation to the stacked 4-C data of all observatories, and then obtain S-wave anisotropy, azimuth and amplitude in shallow sediment layer at each observatory.

(2) Active seismic data processing

As active seismic dataset, we used airgun survey data observed during KR13-17 cruise, including airgun circular shooting around seismometers of DONET and IODP borehole observatories, by R/V Kairei with a tuned airgun array system of 7800 cubic inch. The airgun shootings at circle survey lines were started from North end of each survey line, and were conducted with clockwise direction.
at every 2.0 and 0.5 degree for R3 (radius: 3km) and R10 (radius: 10km) survey lines, respectively. In this study, we used R3 survey line dataset observed by three component seismometers deployed at DONET and IODP C0002G observatories. For more details about data acquisition, see Kimura et al. (2015).

We computed radial and transverse components for all R3 survey lines, comprising C0002G-R3, KMA04-R3, KMB07-R3, KMC09-R3, KMC12-R3, KMD13-R3, KMD15-R3, and KME17-R3. Fig. 4 shows computed radial and transverse components of KMB07-R3 survey line. A simple 2-D simulation by the rotated staggered grid finite difference method (Saenger, 2000) was applied to the horizontal layered model to confirm P-S reflection waves (Fig.2 (c)). From the result of the 2-D simulation, P-S converted wave from the bottom of shallow sediment layers with depth of 1 km is visible at 3.8 s. In obtained radial and transverse dataset of all observatories, we found clear events around 3.0 to 5.0 s, and assumed that these events were P-S converted waves from the bottom of shallow sediments. However, the P-S converted waves reflected from the top of oceanic crust, which were identified around 7.4 s in the result of the 2-D simulation, were not clearly identified in the computed radial and transverse records. Therefore, in this study, we focus on the P-S converted waves from the bottom of shallow sediment layer. We applied a simple method to the radial and transverse dataset, to obtain S-wave anisotropy parameter. For transverse component, we calculated RMS amplitude of target events with time window of around 0.3 to 0.5 s. Fig. 4 (b) shows the result of the KMB07 transverse component. We then obtained 30, 110 degrees from North as S-wave anisotropy azimuth. For radial component, we calculated “anisotropy semblance” to obtain fast S-wave azimuth and anisotropy

Figure 4 (a) Radial and transverse records observed by horizontal components of DONET KMB07 seismometer. (b) Anisotropy parameters calculated form radial and transverse records.
amplitude. The anisotropy semblance was calculated by the following equation:

\[ A(\tau, \phi) = \sum_{i=1}^{n} \sum_{j=1}^{m} d(t_j - (\tau \cdot \cos(2(\theta_i - \phi)) - \tau)), \]  

where \(d\) is observed radial data, \(l\) and \(m\) represent the start and end samples of the time window used for calculation, \(n\) represents number of traces, \(\tau\) represents anisotropy amplitude in second, \(\theta\) represents shot-receiver azimuth for each trace, and \(\phi\) represents fast azimuth of S-wave anisotropy. We obtained optimized \(\tau\) and \(\phi\) by searching maximum value of \(A(\tau, \phi)\). In the KMB07 records, we obtained optimized values as \(\tau = 0.015\) s and \(\phi = 30^\circ\) (Fig.4(b)), and these results were consistent with the result obtained from the transverse record. We applied this method to all dataset, and then obtained S-wave anisotropy parameters in the shallow sediment layer below each observatory.

3. RESULT AND DISCUSSION

We finally obtained S-wave anisotropy information below DONET and IODP C0002G observatories in shallow sediment layer by passive and active seismic dataset. Fig. 5 shows the obtained result of S-wave anisotropy. Yellow arrows and green bars indicate fast S-wave azimuth obtained from passive and active seismic dataset, respectively. S-wave anisotropy amplitudes calculated from passive dataset are also shown in Fig. 5 as interpolated color map. Magenta arrows indicate directions of crustal deformation observed by seafloor acoustic GPS stations (Tadokoro et al., 2012). Blue bar indicates S-wave fast azimuth, as existing result by Tsuji et al. (2011) from borehole walk around VSP survey. In Kumano basin, landward of mega splay fault, the results of passive and active methods have good agreement with difference of 15 deg. maximum, and these results also good agreement with the existing results obtained from the other methods. In KMC09, KMC12 observatories, located in the vicinity of the trough axis, obtained fast S-wave azimuths are parallel to subducting direction. These results imply that the results obtained by our methods can estimate principal horizontal stress in the Kumano basin. However, in the imbricate thrust zone (e. g. Kinoshita et al., 2011), KMB07 and KMD15, the difference between results of active and passive methods are large, more than 45 degree. The differences may be caused by complicated 3-D structure below observatories including seafloor topography and dip reflectors.

4. CONCLUSION

In this study, we estimated S-wave anisotropy using both passive and active seismic dataset acquired by DONET and IODP C0002G seismometers. Obtained results from the datasets have good agreement with each other. There are, however, some differences in the results in complicated structure zone. Since we assumed a simple layered transverse isotropic medium with the horizontal axis of symmetry (HTI) for the model, the dimentionality due to complicated 3D velocity structure, seafloor topography, etc. should attribute to the discrepancy in the survey area. We now plan to develop a new scheme that could accommodate the above-mentioned dimensionality effects, using layer stripping method, 3-D rotation method, etc. The inclusion of the dimensionality in

Figure 5 Obtained S-wave anisotropy in shallow sediment layer from passive and active dataset. Yellow arrows and green bars indicate fast S-wave azimuth obtained from passive and active method, respectively. Magenta arrows indicate directions of crustal deformation observed by seafloor acoustic GPS stations (Tadokoro et al., 2012). Blue bar indicates S-wave fast azimuth, as existing result by Tsuji et al. (2011) from borehole walk around VSP survey.
our method would probably permit the discussion of S-wave anisotropy in deeper parts and of the possibility to detect the stress state in time-series in the subduction zone.

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REFERENCES