Time-lapse change in anisotropy in Japan’s near surface caused by the 2011 Tohoku-Oki earthquake

Nori Nakata & Roel Snieder
Center for Wave Phenomena, Geophysics Department, Colorado School of Mines, Golden, Colorado 80401

ABSTRACT
We apply seismic interferometry to strong-motion records to detect the change in anisotropy caused by the M$_W$ 9.0 Tohoku-Oki earthquake on 11 March 2011. It is known that large earthquakes change fluid condition in cracks, and thereby the polarization-anisotropy condition. The Tohoku-Oki earthquake increased the difference between fast and slow shear-wave velocities arising from shear-wave splitting in most parts of northeastern Japan, but it did not significantly change polarization directions of the fast shear waves in the upper few hundred meters. The difference between fast and slow velocities increased gradually for more than one month after the main shock, whereas the decrease in velocity occurred suddenly after the event. Through monitoring of anisotropy and shear-wave velocity, we find that the changes in anisotropy and velocity recover with time. The change in anisotropy are correlated with the change in velocity in the northeastern Honshu. Also, the change in the largest principal stress direction weakly correlates with that in anisotropy.

1 INTRODUCTION
The change in near-surface shear-wave velocity caused by the M$_W$ 9.0 Tohoku-Oki earthquake on 11 March 2011 is noted by Nakata and Snieder (2011). The earthquake, among the largest in recent history, resulted in a reduction in the near-surface velocity averaged over two months following the earthquake of about 5% throughout northeastern Japan (a region 1,200-km wide). In this study, we estimate the change in near-surface polarization anisotropy by applying seismic interferometry to seismograms recorded by KiK-net, a strong-motion recording network operated by the National Research Institute for Earth Science and Disaster Prevention (NIED).

At a depth greater than a few kilometers, the polarization anisotropy in northeastern half of Japan is related to the mantle wedge; the fast shear-wave polarization directions on the fore-arc and back-arc sides are approximately along the north-south and east-west directions, respectively (Okada et al., 1995; Nakajima and Hasegawa, 2004). Local geology, however, alters the shear-wave polarization in the near surface, a few hundred meters deep, and even at hard-rock sites the fast shear-wave polarization directions on the fore-arc side do not agree with the deep anisotropy (Nakata and Snieder, 2012).

Conventionally, shear-wave splitting is estimated by the cross-correlation method (e.g., Fukao, 1984). One can estimate the fast and slow shear-wave polarization directions and the delay time between the fast and slow shear waves, which are a mean value along a ray path. Moreover, using a cluster of earthquakes, one can estimate the vertical variation of anisotropy (e.g., Okada et al., 1995). The polarization directions and the delay time are changed by large earthquakes (e.g., Cochran et al., 2006). Some previous studies have found the delay time to increase after intermediate or large earthquakes, but the directions do not change (e.g., Saiga et al., 2003; Liu et al., 2004) because the delay time is more sensitive to change in stress than the polarization direction is (Peacock et al., 1988). Since changes in delay times have been observed prior to major earthquakes (e.g., Peacock et al., 1988; Crampin et al., 1990; Crampin and Gao, 2005), monitoring the delay time has been proposed as a diagnostic for earthquake prediction (Crampin et al., 1984).

We present the change in anisotropy based on shear-wave splitting caused by the Tohoku-Oki earthquake inferred from seismic interferometry. First, we show shear-wave splitting of individual earthquake signals at one KiK-net station. Then we compute changes in polarization anisotropy after the main event for all available stations and compare with the changes in shear-wave velocity and the largest principal stress direction caused by the main shock.
Figure 1. Variation in shear-wave velocity and anisotropy coefficient from 1 May 2010 to 31 December 2011 at station FKSH12. The top panel depicts the isotropic velocity of each earthquake (black dot) and its nine-point moving average (blue line). The bottom panel indicates the anisotropy coefficient computed from fast and slow shear-wave velocities (black cross) and its nine-point moving average (blue line). The velocity and the anisotropy coefficient estimated from the main event are illustrated by magenta symbols. Green vertical lines denote the origin time of the event. Red horizontal lines and gray shaded areas are the mean values and the mean values ± the standard deviations of the measurements of all used earthquakes during each period. We show mean values of each period in Table 1.

2 EARTHQUAKE RECORDS AND THE ANALYZING METHOD

2.1 KiK-net

KiK-net, which includes about 700 stations all over Japan, has recorded strong motions continuously since the end of the 1990s (Okada et al., 2004). Each KiK-net station has a borehole a few hundred meters deep and two three-component seismographs, with a 0.01-s sampling interval, at the top and bottom of the borehole. All the earthquakes used in this study are at a depth greater than 7 km, which is large compared to the depth of the boreholes. The velocity in the near surface is much slower than at greater depths. Since we consider events much deeper than the borehole, and because of the slow velocities at the near surface, we assume the waves propagate from the borehole receiver to the surface receiver as plane waves in the vertical direction at each station.

To confirm this assumption, we compute the angle of incidence $\theta$ by employing the procedure proposed by Nakata and Snieder (2012) using ray tracing. All earthquake data used have $\cos \theta > 0.975$, which means the maximum of the estimated velocity bias is 2.5%. The bias is, in practice, much smaller because of averaging over many earthquakes. This inaccuracy is not sensitive to the analysis of shear-wave splitting because $\cos \theta$ is the same for the waves in all polarization directions.

2.2 Seismic interferometry

We apply deconvolution-based seismic interferometry to the seismograms of each station to retrieve the wave that propagates from the borehole receiver to the surface receiver. We deconvolve the seismogram of the borehole sensor from that of the surface sensor in the frequency domain (Snieder and Şafak, 2006):

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D(\omega) \approx \frac{u(r_s, s, \omega)}{u(r_b, s, \omega)} \approx \frac{u(r_s, s, \omega)}{|u(r_b, s, \omega)|^2 + \epsilon},
$$

(1)

where $r_s$ and $r_b$ are the locations of the surface and borehole receivers respectively, $s$ is the position of the hypocenter of an earthquake, $u(r, s, \omega)$ is the wavefield recorded at $r$, and $\epsilon$ is a regularization parameter to prevent instability of the deconvolution. In this study, we use $\epsilon$ as 1% of the average of power spectrum $|u(r_b, s, \omega)|^2$, which is the smallest value necessary to obtain stable deconvolved waveforms (determined by a visual inspection), and we apply a bandpass filter from 1 to 13 Hz after computing the deconvolved waveforms.

To determine the fast and slow polarization directions caused by shear-wave splitting, we follow the
The moving average of the anisotropy coefficient (the blue line in the bottom panel of Figure 1), the anisotropy coefficient continues to increase for more than one month after the main shock, which we might attribute to several large aftershocks during that period. Saiga et al. (2003) observe similar phenomena. In contrast, the velocity decreases suddenly after the main shock (see the blue line in the top panel of Figure 1).
mediate earthquakes that occur before the main shock; such events may change the anisotropy as well. For example, the M6.2 earthquake on 13 June 2010 (a distance of 100 km and a depth of 40 km from station FKSH12) and the M5.7 earthquake on 29 September 2010 (a distance of 40 km and a depth of 8 km from the station) might both have been sources of elevated the anisotropy coefficient. The absence of such intermediate events in the nine weeks before the main earthquake near the station could have caused the observed reduction of the anisotropy coefficient in that period.

We compute the fast and slow polarization directions and estimate the anisotropy coefficient for all available stations throughout Japan for a period before the major earthquake (1 January 2011–10 March 2011) and a period afterward (12 March 2011–26 May 2011) (Figure 2). To reduce uncertainty, we use only stations that have 1) more than three earthquake records during both time intervals, 2) travel times of interferometric waves greater than 0.1 s, 3) anisotropy coefficients greater than 1%, and 4) a standard deviation of velocity measurements smaller than 5%. The average change in the angles of the fast shear-wave polarization directions before and after the main event over all used stations is 17 degrees; this is close to the uncertainty, 15 degrees, computed from data over 11 years (Nakata and Snieder, 2012). We conclude that the fast shear-wave polarization direction does not change significantly as a result of the main shock.

In contrast, the anisotropy coefficient in most parts of northeastern Japan increases after the earthquake. To evaluate the change in the anisotropy coefficient caused by the event, we define the change in anisotropy as \( \Delta C = \frac{C_{\text{after}} - C_{\text{before}}}{C_{\text{before}}} \), where \( C_{\text{before}} \) and

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**Figure 2.** Changes in shear-wave velocity and anisotropy coefficient after the Tohoku-Oki earthquake. Each map has a label at upper-right: anisotropy coefficients before (Before: 1 January 2011 to 10 March 2011) and after (After: 12 March 2011 to 26 May 2011) the Tohoku-Oki earthquake, its change, defined as \( \Delta C = \frac{C_{\text{after}} - C_{\text{before}}}{C_{\text{before}}} \) (Anisotropy change), and the change in the isotropic shear-wave velocity (Velocity change). Dark blue (before) and light blue (after) arrows on the Before, After, and Anisotropy-change maps represent the direction of fast shear-wave polarization. The longitude and latitude pertain to the leftmost map. The dashed black lines show the locations of major tectonic lines (the Median Tectonic Line and the Itoigawa-Shizuoka Tectonic Line) (Ito et al., 1996). Locations and magnitude of the earthquakes from 1 January 2011 to 26 May 2011 are shown as circles and relative to the rightmost map. The size of each circle indicates the magnitude of each earthquake and the color represents its depth. The yellow star denotes the epicenter of the Tohoku-Oki earthquake. The small Japanese map at the top shows four regions for interpretation in Figure 3.
$A_{C_{after}}$ are the anisotropy coefficients before and after the main shock, respectively. The change in the anisotropy coefficient is shown in the second map from right in Figure 2. In Figure 3, we show a crossplot of the changes in the shear-wave velocity and the anisotropy coefficient in four regions defined by the small map in Figure 2. The changes are reasonably well correlated in region II (which is the closest region to the epicenter) and poorly correlated in region I. Different from other regions, most measurements in region IV are in the lower-right quadrant where the velocity increases and the anisotropy coefficient decreases. Region IV is on the west side of the tectonic lines (the Median Tectonic Line and the Itoigawa-Shizuoka Tectonic Line: the black dashed lines in Figure 2), and the geologic age and geomorphological classification both differ across these lines; the west side is an older mountain area, and the east side consists of younger volcanics and sediments (Wakamatsu et al., 2006).

### 4 COMPARING THE CHANGES IN ANISOTROPY AND STATIC STRESS

The Tohoku-Oki earthquake changed the stress and strain conditions (Hasegawa et al., 2011). Changes in stress and strain induce changes in local permeability and pore pressure (Koizumi et al., 1996), and thereby changes in the anisotropy coefficient (Zatsepin and Crampin, 1997). Fluid-filled microcracks, which cause shear-wave splitting (Zatsepin and Crampin, 1997), usually align with the direction of \textit{in situ} stress (Crampin, 1978). Changes in stress caused by intermediate and large earthquakes have been studied for decades (e.g., Hanks, 1977; King et al., 1994; Baltay et al., 2010). Saiga et al. (2003) compare at two stations the time delays associated with shear-wave splitting with the change in Coulomb stress, which is an indicator of how close a fault is to failure (e.g., King et al., 1994). Toda et al. (2011) and Yoshida et al. (2012) compute the change in stress in northeastern Japan caused by the Tohoku-Oki earthquake.

We compare the change in the anisotropy coefficient with the change in the largest principal stress direction computed by Yoshida et al. (2012) (Figure 4). In Figure 4b, the largest principal change in the stress direction and the change in the anisotropy coefficient indicate a weak positive correlation except for the areas B, J, K, and L. Note that the change in the largest principal stress direction is only one proxy of the changes in stress, and we cannot explain the change in the anisotropy coefficient in area B from the change in the principal stress direction. A large change in the principal stress direction ($\geq 20^\circ$) signifies that the stress condition before and after the main shock is significantly changed; therefore a large change in the principal stress direction might induce the large change in the anisotropy coefficient (\$> \pm 10\%\$) in areas A, C, and G. Likewise, a small change in the principal stress direction ($< 20^\circ$) is coincident with the small change in the anisotropy coefficient ($< \pm 10\%$) in areas D, E, F, H, I, and M.

Areas J and K (the asterisks in Figure 4b) are on the west side of the tectonic lines, and area L is close to these lines; thus the change in stress caused by the main earthquake in the upper few hundred meters (the depth range of the boreholes) might differ on the two sides of the tectonic lines. Kern (1978) found in rock-physics experiments that as the confining pressure increases, velocity increases and anisotropy decreases. We speculate that the increase in the velocity and the decrease in the anisotropy coefficient on the west side of the tectonic lines could be explained by increase in the compressional stress, but since we cannot directly measure the compressional stress in this study, we cannot
validate this hypothesis. Note that the inversion model in Yoshida et al. (2012) does not include possible differences in compaction and rheology across these lines.

5 CONCLUSION

By applying deconvolution-based seismic interferometry to KiK-net data, we measure changes in anisotropy caused by the Tohoku-Oki earthquake. The anisotropy coefficient increases in most parts of northeastern Japan after the Tohoku-Oki earthquake, but the fast polarization direction does not change. The changes in shear-wave velocity and anisotropy both recover with time. Comparison of the changes in the shear-wave velocity and the anisotropy coefficient shows strong correlation in the eastern half of Honshu have a strong correlation. The changes in the anisotropy coefficient and the largest principal stress direction are weakly correlated. On the west side of the tectonic lines, the increase in velocity and the decrease in anisotropy could be explained by a difference of the change in stress across the tectonic lines.

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Figure 4. (a) Anisotropy change in Figure 2 with the largest principal stress directions before (red arrows: the same as the red arrows in Figure 3a in Yoshida et al. (2012)) and after (blue arrows: the same as the red arrows in Figure 3b in Yoshida et al. (2012)) the Tohoku-Oki earthquake. The arrows are estimated in each 0.5°-grid area. Black dashed lines indicate the locations of the major tectonic lines. A–M areas denote the interpreted regions in panel (b) as well as Yoshida et al. (2012). (b) Crossplot of the changes in the largest principal stress direction (Yoshida et al., 2012) and the anisotropy coefficient in each area shown in panel (a). The change in the anisotropy coefficient is the mean value for each 0.5° grid. Asterisks indicate areas on the west side of the tectonic lines.
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