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EXPRESS LETTER

Report on a characteristic oscillation about 38 mHz (26 s) in northeastern Japan following surface wave of the 2011 Tohoku megathrust earthquake

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SUMMARY

We try to detect an unidentified signal from the surface motion at northeastern Japan immediately after the 2011 Tohoku earthquake. A focused frequency range is 10–100 mHz (10–100 s). We find a peaky signal with frequency of about 38 mHz (26 s) based on the horizontal-to-vertical (H/V) spectral ratio using the high-rate GNSS data at 382 GEONET stations. We are not able to identify locality of the signal. The signal appears several minutes after the passing of surface wave fronts. The duration of the signal is about 2 min. Since the origin of the 38 mHz signal is unlikely to be local hydrologic tremors, tectonic tremors, or the tsunami, we speculate that the 38 mHz signal originates from a kind of a characteristic oscillation of Northeastern Japan triggered by the 2011 Tohoku earthquake. A normal-mode simulation implies that high-order radial overtones could create the signal with a spherically-layered velocity structure, however, the detailed mechanism of the signal still remains a mystery.

Key words: Satellite geodesy; Transient deformation; Broad-band seismometers; Surface waves and free oscillations; Asia.

1 INTRODUCTION

The 2011 Tohoku megathrust earthquake at northeastern Japan near the Japan trench ruptured a wide region of the plate interface between the Pacific Plate and the Okhotsk Plate. The moment magnitude of the Tohoku earthquake was over 9 (e.g. Nettles et al. 2011), causing significant seismic wave on the Japan islands (e.g. Furumura et al. 2011). The radiated seismic wave dynamically triggered small earthquakes (Miyazawa 2011; Gonzalez-Huizar et al. 2012) and local landslides (Miyagi et al. 2011). Other triggered phenomena might occur after the 2011 Tohoku earthquake without being discovered.

Global Navigation Satellite System (GNSS) can detect surface deformation in a wide range of frequency from 0 to 50 Hz (e.g. Ge et al. 2000; Larson et al. 2003; Genrich & Bock 2006; Avallone et al. 2011). A GNSS array by Geospatial Information Authority of Japan, GEONET, succeeded in detecting seismic wave precisely (Grapenthin & Freymueller 2011; Psimoulis et al. 2014) and Earth’s free oscillation (Mitsui & Heki 2012). Similar to previous studies, we focus on the 1-Hz data immediately after the 2011 Tohoku earthquake at the GEONET stations in NE Japan (Fig. 1).

2 METHODS

We use GNSS time-series data analysed by the kinematic precise point positioning method (e.g. Kouba & Heroux 2001), provided by the Geospatial Information Authority of Japan (GSI) via Nippon GPS Data Services Company (NGDS) and the VERIPOS/APEX service using the RTnet software (GPS Solutions). The sampling period Δt is 1 s. The absolute coordinates of the GNSS stations were estimated on the basis of the precise ephemeris and clock with the ionospheric-free linear combination of dual-frequency carrier-phase observations.

Errors in the ephemeris or clock, and tropospheric path delay of radio waves might contaminate the analysed data for the land displacement, but there were no organized reports about such an error in the period immediately after the 2011 Tohoku earthquake, to our knowledge. Besides, we can neglect characteristic oscillations of the GNSS antennas since its natural frequency is certainly higher than the sampling frequency of 1 Hz (note that 1 Hz is roughly as large as natural frequencies of buildings).

We focus on the 1-Hz data immediately after the 2011 Tohoku earthquake at the GEONET stations in NE Japan (Fig. 1).
We use GNSS time-series data at 382 GEONET stations (solid triangles) and seismometer data at 13 F-net stations (red stars) in NE Japan, near the source region of the 2011 Tohoku earthquake. The GCMT solution (Nettles et al. 2011) is represented by the beach-ball symbol. The ruptured area estimated by Geospatial Information Authority of Japan (2011) is also represented by the dashed squares. The right-hand figures show the observed three-component displacement data at selected GEONET stations. The 0144 station is at \(142.935^\circ\,E, 42.131^\circ\,N\), the 0920 station is at \(140.728^\circ\,E, 37.825^\circ\,N\), and the 3086 station is at \(138.838^\circ\,E, 34.610^\circ\,N\).

Figure 2. Power spectral densities in three components (north–south, east–west and up–down) over the 382 GNSS stations after the occurrence of the 2011 Tohoku earthquake (14:46). The length of the time window for the running spectrum is 6 min and the increment of the window shift is 2 min.

The 2011 Tohoku earthquake occurred around 14:46 (JST) on March 11. Since the large wave groups of the surface waves had almost passed through the GEONET stations around 14:50 (Grapenthin & Freymueller 2011), we mainly consider the data after 14:51 (JST).

In order to obtain anonymous signals, we first remove linear trends from all the time-series data and then compute their power spectral densities. The power spectral density \(P\) for each data is given by \(\frac{|D|^2}{T}\), where \(D\) is the data in the frequency domain and \(T\) is the measurement period. We obtain \(D\) by the discrete Fourier transform using an FFT method (Singleton 1979) with the Hann window (Blackman & Tukey 1959):

\[
D(k) = \Delta t \sum_{m=0}^{N-1} d(m)(0.5 - 0.5 \cos[2\pi m/(N - 1)]) \times \exp(-2\pi i km/N),
\]

where \(\Delta t\) is the sampling period (1 s), \(N\) is number of the time-series with zero padding, and \(d\) is the original (time-series) data.

3 RESULTS

We first checked the power spectral density at each station but the spectra were rather complex and oscillating. Following our previous study (Mitsui & Heki 2012), we stack the data over many stations to extract some signals. Hence we obtain median values of the power spectral densities over all the stations (382 stations) for three components (north–south, east–west and vertical). Fig. 2 shows the power spectral densities after the occurrence of the main shock (14:46). In all the components, the spectral densities have maximal values immediately after the main shock and decay until about 15:10. The increases in the densities around 15:10 are caused by an \(M7\)-class aftershock, which were the first large aftershock of the Tohoku earthquake (e.g. Asano et al. 2011). There does not seem other notable signal in Fig. 2.
38 mHz oscillation at NE Japan

Next, we take the horizontal-to-vertical (H/V) spectral ratio (Nakamura 1989; Bonnefoy-Claudet et al. 2006), which has been generally used to evaluate local site effects relating to the fundamental mode of Rayleigh waves in a frequency range higher than 1 Hz.

Fig. 3 presents the results. We find an isolated peak around 38 mHz (26 s) in a period about 14:55–14:57. Such a sharp spectral peak following the 2011 Tohoku earthquake has not been reported before. Apparently, the larger spectral density in E-W component and the smaller density in U-D component constructed the larger H/V signal. The extent of the peak concentrates in a range of 36–40 mHz. This range differs from the second Airy phase (about 20 mHz) of the fundamental mode of Rayleigh waves for isotropic PREM (Widmer-Schnidrig & Laske 2007).

In order to check locality of the signal, we divide the observation area into Eastern–Western zones (the boundary is 140.0°E) or northern–southern zones (the boundary is 37.0°N). Fig. 4 shows the results and reveals that the typical signal around 38 mHz appears over the whole observation area.

**Figure 3.** The H/V spectral ratio from the median value over the 382 GNSS stations. The left-hand figure is the running spectrum, and the right-hand figure is the simple spectrum for a period of 14:51–15:00.

**Figure 4.** The H/V spectral ratio from parts of the GNSS data. The left-hand figure shows the median H/V ratios for the 174 eastern stations (>140.0°E) and the 208 western stations (<140.0°E). The right-hand figure presents the median H/V ratios for the 193 northern stations (>37.0°N) and the 189 southern stations (<37.0°N).

4 DISCUSSION

We only used the GNSS displacement data for analysing the post-seismic deformation of the 2011 Tohoku earthquake. Since there are many seismometer stations on Japan, we also use continuous time-series data by the VSE-355G2 and VSE-355G3 strong-motion seismometers to compare with the GNSS results. The distribution of the seismometer stations of the F-net (by National Research Institute for Earth Science and Disaster Prevention) is illustrated in Fig. 1. We analyse the seismometer data by the same method as the GNSS data excluding the following points. First, we must correct the instrument responses of the seismometers in a low frequency range using complex transfer function in the Fourier domain. Next, the sampling rate $\Delta t$ for the seismometer is 0.01 s. Fig. 5 shows the comparative results of the GNSS and the strong-motion seismometer. From the left-hand figure, we find that the 38 mHz peaky signal also exists during 14:51–15:00 in the seismometer data, but several other peaky signals appear especially in the lower frequency range possibly caused by the less accuracy of the seismometer data for such a frequency range (Clinton 2004) or the smaller numbers of the seismometer stations. To check this point, in the right-hand figure, we compare between the GNSS data at one station and the seismometer data at a close station. We focus on a period of 15:09–15:11, which includes a seismic wave by the $M_7$-class aftershock occurred about 15:08 (e.g. Asano et al. 2011). We found that the GNSS and seismometer data are better consistent. It implies that the several peaky signals only appeared in the seismometer data during 14:51–15:00 are owing to the smaller numbers of the seismometer stations.

The origin of the 38 mHz signal is now under discussion. We were not able to identify the locality of the signal. Thus the signal does not originate from local hydrologic tremors (Nishida & Shiomi 2012). Extensively triggered tectonic tremors are also unlikely to operate as the source of the signal, since the occurrence of the tectonic tremors has been rarely reported in Northeastern Japan (e.g. Beroza & Ide 2011). One may consider that the huge tsunami
approaching the coast of Japan caused the 38 mHz signal. However, this mechanism seems also improbable because the tsunami load mainly induces vertical land deformation (e.g., Mitsui & Heki 2013) leading to small H/V ratios. Another possibility is the effect of a thick column of water near the trench (Nakanishi 1992; Yomogida et al. 2002; Noguchi et al. 2013). Noguchi et al. (2013) demonstrated the importance of the water layer on the propagation of the Rayleigh wave around a period of 10–15 s near the Japan trench. This mechanism may affect the land deformation after the 2011 Tohoku earthquake, but the period range seems slightly different from that of the observed peaky signal in this study.

By contrast, a more general mechanism independent of locality may explain the 38 mHz signal. For this purpose, we compute synthetic seismograms of the 2011 Tohoku earthquake by summation of normal modes. We input the GCMT solution (Nettles et al. 2011) and the Preliminary Reference Earth Model (Dziewonski & Anderson 1981) into the MINEOS package (Masters et al. 2011). The shallow ocean part in the preliminary reference earth model is filled with the solid crust part. Fig. 6(a) shows the spherically layered structures of the P-wave velocity, the S-wave velocity, the density, and the quality factor Q, above 250 km depth.

We calculate the seismograms at a representative point (140° E, 39° N) caused by the GCMT solution for the 2011 Tohoku earthquake. For the normal-mode summation, we consider both spheroidal modes, \( nS_l \), and toroidal modes, \( nT_l \), where \( n \) means the radial order number and \( l \) represents the angular order number. Fig. 6(b) exhibits the calculated seismograms when the mode summation is done for a period range longer than 4 s with several
overtones ($n \leq 8$), and only the fundamental mode without overtones ($n = 0$). We found that one spectral peak appears around 35 mHz in the overtone case. Additional experiment implied that the high overtone modes ($n = 6–8$) especially have potentials to make a peak around 35 mHz. It is unclear whether this experimental result well explains our observational result or not, but the experimental result implies that high-order radial overtones of strong seismic wave could make a peaky spectral signal in an ordinary (not local) layered structure.

Besides, we additionally state that the characteristic frequency of 38 mHz (26 s) in this study is the same frequency as an enigmatic microseism observed in the Gulf of Guinea (Oliver 1962; Shapiro et al. 2006). This peculiar consistency might be a key to understand the source mechanisms of the enigmatic signals.

5 CONCLUSION

We found the characteristic oscillation about 38 mHz (26 s) in Northeastern Japan that transiently appeared immediately after the 2011 Tohoku earthquake, using the H/V spectral ratio from the GNSS and the strong-motion seismometer data. We were not able to identify the locality of the characteristic peaky signal. The signal might be caused by the high-order radial overtones, which do not originate from a complex and local velocity structure, but further examination is needed.

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