Millimeter-scale tsunami detected by a wide and dense observation array in the deep ocean:

Fault modeling of an Mw 6.0 interplate earthquake off Sanriku, NE Japan

T. Kubota¹, T. Saito¹, and W. Suzuki¹

¹National Research Institute for Earth Science and Disaster Resilience, Japan.

Corresponding author: Tatsuya Kubota (kubotatsu@bosai.go.jp)

Key Points:

- Millimeter-scale tsunamis from an Mw 6.0 earthquake were captured by the S-net, a new nation-wide pressure gauge array off Sanriku, Japan
- Tsunami signals were identified from the pressure data adjacent to the source, which were contaminated by signals irrelevant to tsunamis
- We inferred the stress drop of the earthquake from the tsunami data more reliably than could be done from seismogram analysis
Abstract
A new dense and widely distributed tsunami observation network installed off northeast Japan detected millimeter-scale tsunamis from an Mw 6.0 shallow interplate earthquake on 20 August 2016. Based on the fault model deduced from this dataset, we obtained a stress drop of 1.5 MPa for this event, similar to those associated with typical interplate earthquakes. The rupture area was unlikely to overlap with regions where slow earthquakes occur, such as low-frequency-tremors and very-low-frequency-earthquakes. The results demonstrated that this new network has dramatically increased the detectability of millimeter-scale tsunamis. Some near-source stations were contaminated by large pressure offset signals irrelevant to tsunami, and we must therefore be careful when analyzing this data. Nonetheless, the new array enables estimations of the stress drops of moderate offshore earthquakes and can be used to elucidate the spatial variation of mechanical properties along the plate interface with much higher resolution than previously possible.

Plain Language Summary
Tsunamis are generated when an earthquake occurs beneath the seafloor. Far fewer tsunami observations have been recorded from moderate earthquakes than large to giant earthquakes because tsunamis created by moderate earthquakes have been too small to be observed. On 20 August 2016, a moderate earthquake occurred off Sanriku, in northeastern Japan, and a tsunami with a height of less than one centimeter was recorded by a new seafloor tsunami observation network. This network has many tsunami sensors distributed much closer to each other and over a much wider area than any other previous network in the world. Using this data, this study
estimated the source location and size, and the slip amount of the 2016 earthquake with higher accuracy, which was impossible to achieve from past observations because they were too far away from the earthquake and the signals were too small. Using this source information, we could estimate the stress drop associated with the earthquake, which is important because the stress drop information deepens our understanding of how and why earthquakes happen.

Keywords

millimeter-scale tsunami
fault modeling
interplate earthquake
Sanriku
S-net
1. Introduction

To understand the mechanics of earthquake faulting, the earthquake source process and mechanical properties in and around the rupture area must be identified and understood. An important parameter in these investigations is the stress drop, the average shear stress reduction on the fault due to the earthquake. It has been commonly thought the stress drops show no systematic change from smaller to larger earthquakes (e.g., Kanamori & Anderson, 1975), suggesting a universal mechanism of earthquake faulting irrespective of size. However, recent studies have reported that the estimated stress drops change depending on the frictional properties along the fault and/or the background stress level (e.g., Lay et al., 2012; Yoshida et al. 2017).

Finite fault modeling using teleseismic data is one of the most common approaches to estimating the stress drops for major (Mw >~7) earthquakes (e.g., Ye et al. 2016). Regional onshore seismograms have also been widely used to measure the corner frequencies and estimate stress drops for minor (Mw <~5) earthquakes (e.g., Uchide et al., 2014; Yamada et al. 2017; Yamashita et al., 2004). However, accurately constraining stress drops for moderate (Mw ~6) offshore earthquakes remains difficult because of the uncertainties attributed to the low signal-to-noise-ratio in these datasets. In fault modeling with seismic waves, a tradeoff between rupture area and rupture velocity may cause considerable uncertainties in the stress drop estimation.

For major offshore earthquakes, tsunami data recorded by ocean-bottom pressure gauges can be used to construct the fault models (e.g., Gusman et al. 2015; Heidarzadeh et al. 2016; Newman et al., 2011; Satake et al. 2013; Williamson et al., 2017; Saito & Kubota, 2020). Given the much slower propagation velocity of tsunamis, tsunamis are much less affected by the
rupture propagation than seismic waves, which contributes to more robust constraints on the earthquake’s horizontal location and fault dimensions, and thus on the earthquake stress drop (e.g., Kubota, Saito et al., 2018; Saito & Kubota, 2020). However, few observations of tsunamis excited by Mw ~6 earthquakes have been reported (Hino et al., 2001), because the typical offshore stations are too few (only one or two stations) and remote from the earthquake (> ~100 km); challenging the capture of tsunamis of reasonable and sufficient quality. Therefore, to constrain the fault models based on tsunamis requires more stations located closer to the focal area with better station coverage.

A new, wide and densely-distributed seafloor pressure gauge network called the Seafloor Observation Network for Earthquakes and Tsunamis along the Japan Trench (S-net) was recently constructed by the National Research Institute for Earth Science and Disaster Resilience (NIED), (Kanazawa et al., 2016; Mochizuki et al., 2016; Uehira et al., 2016, Figure 1). This network observed tsunamis with amplitudes of less than one cm (hereafter termed millimeter-scale tsunamis) associated with a moderate earthquake off Sanriku, in northeastern Japan, on August 20, 2016 (Figure 1, Mw 6.0, Ekström et al., 2012; http://globalcmt.org). This event was located near the up-dip ends of the rupture areas of the 1968 (Mw 8.0, Satake, 1989) and the 1994 Sanriku earthquakes (Mw 7.7, Tanioka et al., 1996), and at the northern edge of the rupture area of the 1896 Mw 8.1 Sanriku tsunami earthquake (Kanamori, 1972; Satake et al., 2017; Tanioka & Satake, 1996). Low-frequency-tremors with a dominant frequency of a few Hz and very-low-frequency-earthquakes with periods of 10–20 s have also been observed around this region (Asano et al., 2008; Matsuzawa et al., 2015; Nishikawa et al., 2019; Tanaka et al., 2019). Because the rupture areas of these slow earthquake phenomena and those of regular earthquakes
are thought not to overlap each other (e.g., Nishikawa et al., 2019), the differences between the regular and slow earthquake source processes and their relation to mechanical properties along the plate interface are of interest.

This study used the S-net tsunami records to estimate the finite fault model and the stress drop for the 2016 Sanriku earthquake and then examine its relationship with other interplate events, including slow earthquakes and past earthquakes. In Section 2, we describe the characteristics of the S-net pressure data from the 2016 event and Section 3 discriminates the tsunami signals from the data, which were contaminated by noise. Section 4 then infers the finite fault model and the stress drop of the 2016 event and gives the estimation uncertainties. We finally discuss the relationship with the other phenomena such as slow earthquakes and the 1896 tsunami earthquake.
Figure 1. (a) Tectonic setting off northeastern Japan. White stars and red hemispheres indicate the Japan Meteorological Agency (JMA)’s epicenter and the global centroid moment tensor solution of the 2016 Mw 6.0 earthquake, respectively. Fault models of the 1896 (Mw 8.1, Satake et al., 2017), 1968 (Mw 8.0 Satake, 1989), and 1994 earthquakes (Mw 7.7, Tanioka et al., 1996) are shown. Green and light blue circles denote very-low-frequency-earthquakes (Matsuzawa et al., 2015; Nishikawa et al., 2019) and low-frequency-tremors (Tanaka et al., 2019), respectively. Small gray circles indicate regular earthquakes with thrust-faulting mechanisms (Fukuyama et al., 1996).
1998, Mw ≥ 3.5, strike: 170–210°, dip: 0–40°, rake: 45–135°). Black arrows indicate the
direction of the plate motion (DeMets et al., 2010). Pink dashed lines in the inset show iso-depth
contours of the plate interface (Iwasaki et al., 2015) at 20-km intervals. (b) Locations of the
pressure gauges. Circles denote the S-net stations (colors denote groups of observation nodes).
Blue and pink triangles denote the Off-Kushiro (KPG) and Deep-Ocean Assessment and
Reporting of Tsunamis (DART) stations, respectively. (c) Vertical profile along the A-B line.
Red and black lines show the plate boundary and the seafloor.

2. Tsunami Data

To identify tsunami, an acausal band-pass filter with a passband of 100–1000 s was
applied to the S-net pressure records (Figure S1). It seems very difficult to recognize the
tsunamis if only from a few stations, given the background pressure changes are almost ~1 cm at
the stations near the epicenter (Figure S1). However, the westward propagation of tsunamis with
amplitudes of a few cm could be recognized when the traces were aligned according to the
station locations (Figures S1 and S2). Some pressure records showed small subsequent phases
after the main tsunami pulses (Figure S2), due to dispersive tsunami associated with the short-
wavelength components (e.g., Kirby et al. 2013; Saito & Furumura 2009; Saito et al. 2010).

Step-like signals were also observed at several stations near the source. At the station
S4N10, closest to the epicenter (~10 km), a large and abrupt pressure step of ~20 hPa was
observed (Figure S3). Similar steps were observed by the pressure gauges located just above the
focal area during the 2016 Off-Mie earthquake (Mw 6.0), off southwest Japan (Kubota, Suzuki et
al., 2018; Wallace et al., 2016). Pressure offset changes are often recognized as vertical seafloor
deformation (Saito & Tsushima, 2016; Tsushima et al. 2012), although in other cases, such changes have been attributed not to the vertical deformation but to the strong shaking of the seafloor (e.g., Hayashimoto et al. 2015; Nakamura & Hayashimoto, 2019). Takagi et al. (2019) examined records from the co-located accelerometer at S4N10 during this earthquake and found the sensor at S4N10 was rolled with a rotation angle of 5.72° (see Figure 3a in Takagi et al., 2019). It has been reported that outputs of pressure sensors identical to those in the S-net (Paroscientific Digiquartz sensors) strongly depend on the sensor rotation (Chadwick et al., 2006; the rotation angles of 5° corresponds to pressure changes of ~5 hPa). Therefore, careful analysis is required to determine whether or not the step-like signals are caused by tsunami.

We also used four other pressure gauges, more than 150 km away from the source (Figure 1b). Two of them are located off Kushiro, northern Japan (Hirata et al., 2002; hereafter KPG) and the other two, from the Deep-Ocean Assessment and Reporting of Tsunamis (DART) system (Bernard et al., 2014), are located outside of the Japan Trench. Although fluctuations with amplitudes of < 1 cm are seen (at ~10–20 min from the focal time), these fluctuations are comparable to the noise (Figure S1). We also observed large pressure changes due to the seismic waves (e.g., An et al., 2017; Kubota et al., 2017) soon after the focal time, which further makes it difficult to identify tsunamis.

3. Preparatory Analysis: Tsunami Forward Simulation

Before estimating the finite fault model of the 2016 event, we conducted several preparatory analyses. We first simulated tsunamis to identify tsunami signals in the pressure change based on the GCMT solution. Using the GCMT seismic moment $M_0 = 1.31 \times 10^{18}$ Nm
(Mw 6.0) and the scaling law of Blaser et al. (2010), we obtain the fault length \( L = 8.3 \text{ km} \) and width \( W = 8.2 \text{ km} \). We also assumed the uniform slip amount of \( D = M_o/\mu LW = 48.2 \text{ cm} \) (rigidity \( \mu = 40\text{GPa} \) was assumed). Horizontal location of the fault center and the strike, dip, and rake angles were fixed to the GCMT centroid, which were consistent with the subducting plate interface (Iwasaki et al., 2015) and the direction of the plate convergence (DeMets et al., 2010). The spatial distribution of the seafloor vertical displacement was calculated assuming elastic half space (Okada, 1992). Then the seafloor displacement was converted to the sea-surface elevation considering the spatial smoothing effect due to the deep seawater (e.g., Kajiura, 1963; Saito, 2019) (seawater depth of \( H_0 = 4 \text{ km} \) was assumed). The rectangular fault model and the sea-surface displacement are shown in Figure S4a. Using the sea-surface elevation as the initial condition, we simulated tsunami by numerically solving the linear dispersive wave equation in the Cartesian coordinates (e.g., Saito, 2019). We used JTOPO30v2 bathymetry data with 30 arcsec spatial resolution, provided by Japan Hydrographic Association's Marine Information Research Center (http://www.mirc.jha.jp/en/), interpolating the grid interval of 2 km. The time step interval was 1 s. After the calculation, we applied the same band-pass filter to the numerically calculated waveforms as applied to the observed waveforms.

Although the tsunami signals in the S-net pressure records could be distinguished by the simulated waveforms, the step-like signals just after the focal time were not reproduced at all (Figure S4b), indicating the step-like signals were related to neither the tsunami nor the seafloor deformation. The apparent pressure changes due to the seismic motion irrelevant to tsunami can be seen only in a few stations closest to the focal area. On the other hand, the simulated pressure change shows a systematic spatial and temporal evolution across the stations. Therefore, we can
distinguish the tsunami signals in the observed records by using the array observation data of the S-net that captures a systematic spatial and temporal tsunami propagation. However, if there were one or a few stations, it would be difficult to distinguish noises from the pressure records.

We also found that at some stations, the simulated tsunamis were slightly delayed and were elongated in time compared to the observation. This implies that the actual fault location should be horizontally shifted from the GCMT centroid. The S-net pressure records were expected to improve the constraint for the fault location, whereas estimating the epicenter based on seismic waves can produce non-negligible errors, as pointed out by Inazu and Saito (2014), Kubota, Saito et al. (2018) and Saito & Kubota (2020).

To further discuss the importance of the near-field tsunami data to resolve the fault model, we additionally inverted the tsunami waveforms to estimate the initial sea-surface height (i.e., tsunami source) distribution and evaluated the spatial resolution of the analysis (the details are in Text S1, Figures S5 to S7). As a result, the tsunami source had a spatial extent of ~40 km and was located ~10 km to the west of the GCMT solution. In addition, when using only the KPG and DART data, the horizontal location was similar; however, the spatial distribution was broader and the total volume of the displaced seawater was almost twice as large as that estimated using the S-net data. This suggests the horizontal location could be reasonably constrained only from the remote stations but the near-field S-net tsunami data is essential to obtain the higher-resolution fault models.

4 Grid-Search for Optimum Fault Model
We then estimated the fault model that could best explain the observations, using a two-step grid-search approach. First, fixing the fault dimension, the horizontal location of the fault was constrained. Second, we searched for the fault dimensions that best matched the observation data by fixing the fault’s horizontal location.

To constrain the horizontal location of the fault in the first step, we assumed a rectangular fault identical to that used in the forward simulation based on a scaling law proposed by Blaser et al. (2010). The fault center was constrained to coincide with the depth of the plate interface. The suitability of each fault model candidate was assessed based on the similarity of the tsunami waveforms, in terms of the variance reduction (VR):

\[
VR = \left[1 - \frac{\sum_{i=1}^{N}(x_{i}^{\text{obs}} - x_{i}^{\text{cal}})^{2}}{\sum_{i=1}^{N}x_{i}^{\text{obs}}^{2}}\right],
\]

where \(x_{i}^{\text{obs}}\) and \(x_{i}^{\text{cal}}\) are the observed and calculated tsunami waveforms of the \(i\)th data sample, respectively (\(N\) is the total number of data samples). We used the main tsunami phases (thick black lines in Figure S8b) to calculate the VR.

The fault location that produced best the pressure records was located \(~\)10 km west from the GCMT centroid, and was consistent with the tsunami source estimated by the inversion (VR = 55%, Figure S8). The uncertainty of the horizontal location was assessed based on the simulation by shifting the centroid location (Figure S9). When shifting the centroid location by 5 km from the best-matched location in the east-west direction, the peak timings of the simulated waveforms at S4N12 (~100 km west of the centroid) did not explain the observation (temporal
differences of ~ 0.5 min, Figure S9a), although a simulation that shifted the centroid by 1 km explained the observations reasonably well. A similar tendency was seen at the station S4N09 (~50 km northeast of the source), when the centroid location was shifted in the north-south direction (Figure S9b). Based on these experiments, we concluded that the horizontal location of the earthquake was identified with a resolution of ~5 km.

We then estimated the fault dimensions that best explained the observations by changing the fault length and width. The centroid location (longitude, latitude, and depth) was fixed to that obtained by the grid-search for the fault’s horizontal location. The slip amount was also estimated, such that VR was maximized.

As a result, we obtained \( L = 17 \text{ km}, W = 5 \text{ km}, \) and \( D = 40.5 \text{ cm} \) (Figure 2). The seismic moment \( M_o = 1.37 \times 10^{18} \text{ Nm} \) (\( \mu = 40 \text{ GPa}, \text{Mw 6.0} \)) was comparable to the GCMT solution. The best-matched model explained the observations as well as the inversion result (VR = 57%; Figure 3). We calculated the average stress drop on the fault using the following equation (e.g., Kanamori & Anderson, 1975; Madariaga, 1977):

\[
\Delta \sigma_s = c \mu \frac{D}{(LW)^{1/2}} = c \frac{M_o}{(LW)^{3/2}},
\]

(3)

where \( c \) is a constant (\( = 8/3\pi \), assuming Poisson’s ratio of 0.25, Kanamori & Anderson, 1975). We thus obtained the stress drop of \( \Delta \sigma_s = 1.5 \text{ MPa} \).
Figure 2. Result of the grid-search for the fault dimensions that best matched the observation data. Pink circles denote the stations used for the analysis. Small green and light blue circles denote very-low-frequency-earthquakes (Matsuzawa et al., 2015; Nishikawa et al., 2019) and low-frequency-tremors (Tanaka et al., 2019), respectively. Small open circles indicate thrust fault earthquakes (Fukuyama et al., 1998).
**Figure 3.** Comparison of the observed (gray) and synthesized waveforms from the results of the grid-search for the fault dimension (red). The observed waveforms indicated by thick black lines were used for the analysis.

We evaluated the uncertainties of the fault dimensions based on the tsunami simulation. Fixing the fault center location and the seismic moment $M_o$, we changed the fault length and width (Figure 4 for the stations S4N09 and S4N12, and Figures S10 and S11 for the other stations). At the station S4N12, when assuming the larger fault width ($W \geq 10$ km), the first up-motion wave (~7 min) was reasonably captured by the simulation, whereas the first down-motion wave (~8 min) and the secondary up-motion waves (~9.5 min) were delayed and their amplitudes were small compared to the observation (Figure 4a). This tendency was also seen when assuming the larger fault length ($L \geq 25$ km, Figure 4b). One reason why the larger fault dimension could not explain the later phases was the differences between the horizontal locations of the uplift and subsidence peak. Furthermore, when assuming the smaller fault length ($L \leq 10$ km), the amplitudes of the secondary up-motion waves at the station S4N09 were larger than the observation (Figure 4b). This was probably due to the tsunami dispersion, because shorter wavelength tsunami components are strongly affected by dispersion (e.g., Saito et al., 2010). We should note that the simulated waveforms showed no difference in the remote stations (e.g., the DART or KPG records, Figure S9 and S10), indicating that the remote stations cannot resolve the fault dimension and stress drop well, and we need to use the near-field records to precisely constrain the fault dimension and stress drop (similar discussion is also made by Kubota et al., 2019).
Taking this result into account, we concluded that the uncertainties of the fault length and width were within the ranges of $15 \leq L \leq 20$ km and $W \leq 7$ km. Given the largest plausible fault dimensions ($20$ km $\times$ $7$ km) and the seismic moment of the best-matched solution, the stress drop of this event was expected to be larger than $\sim 0.7$ MPa. Even if we suppose the fault depth is shallower or deeper than $\sim 5$ km than the plate boundary (Iwasaki et al., 2015), the plausible maximum fault dimension and minimum stress drop was $L = 20$ km and $W = 10$ km, and $\Delta \sigma_s \sim 0.5$ MPa (see Text S2 and Figures S12 and S13 for the detail). Furthermore, in order to discuss the plausible maximum stress drop, we investigated the fault parameters and the VR values for all model candidates in Figure S14 (detail is described in Text S2). Taking a trade-off between each of the model parameters on the fault model candidates into account, the upper limit of the plausible stress drop range could be estimated at $\sim 5$ MPa (Figure S14).
Figure 4. Comparison of the observed (gray) and simulated (red) waveforms at (a) S4N12 with the width changed and (b) S4N09 with the length changed. The timings of the maximum and minimum peaks are denoted by the red arrows. The fault locations are shown in the insets.

5 Discussion and Conclusions

On 20 August 2016, a Mw 6.0 earthquake occurred off Sanriku and a millimeter-scale tsunami was observed by a dense offshore pressure gauge network called S-net. We first identified tsunami signals from the S-net data based on the forward tsunami simulation. The inversion of the tsunami data for the tsunami source showed that the horizontal location was ~10 km west of the GCMT solution. An additional inversion without the S-net data showed that the S-net records were essential to constrain the high-resolution source model. We then conducted
the grid-search analyses to constrain the centroid location and fault dimensions. This grid-search
showed the uncertainty in the horizontal location was $\pm 5$ km. We obtained the fault
parameters of $L = 17$ km, $W = 5$ km, and $D = 40.5$ cm (Mw 6.0), and $\Delta\sigma_s = 1.5$ MPa. Based on
the uncertainty evaluation, the possible range of the model parameters are $15 \leq L \leq 20$ km and $W \leq 7$ km, and $\Delta\sigma_s \sim 0.7–5$ MPa.

The 2016 event occurred at the northern edge of the rupture area of the 1896 Sanriku
tsunami earthquake (Satake et al. 2017; Tanioka & Satake, 1996). Tsunami earthquakes are
defined as earthquakes that generate much larger tsunami than expected from the surface wave
magnitudes and radiate less energy compared to typical interplate earthquakes (e.g., Kanamori,
1972; Newman & Okal, 1998; Ye et al., 2016). Tsunami earthquakes are commonly thought to
have smaller stress drops of less than 1 MPa (e.g., Kanamori and Anderson, 1975; Ye et al.,
2016). Bilek et al. (2016) reported that the mean stress drops associated with minor earthquakes
(M $< \sim 4–5$) within the 1992 Nicaragua tsunami earthquake rupture area were significantly
smaller than those outside of the rupture area. However, in this study, the stress drop of the 2016
event (1.5 MPa) was not as small as expected in tsunami earthquakes. Our result was rather
consistent with the stress drops of small earthquakes around this area determined from the corner
frequencies of onshore seismometers (Yamashita et al., 2004; Uchide et al., 2014; Yamada et al.,
2017); however they did not show systematic differences of the stress drops between inside and
outside of the 1896 rupture area. Conventional stress drop estimations based on onshore
seismograms can contain significant errors originating from the unreliable estimation of the
corner frequency. Therefore, although one plausible interpretation for typical stress drop in the
2016 earthquake is this earthquake was located outside of the 1896 rupture area, we must
investigate additional examples of stress drops of moderate earthquakes to quantify the spatial relationship between moderate earthquakes and the 1896 earthquake with greater confidence.

Low-frequency-tremors and very-low-frequency-earthquakes were also detected in this region (Matsuzawa et al., 2015; Nishikawa et al., 2019; Tanaka et al., 2019). Based on our fault model, the rupture area of the 2016 event was isolated from the low-frequency-tremor and very-low-frequency-earthquakes region. This spatial isolation may reflect a spatial difference in the frictional properties along the plate boundary (e.g., Nishikawa et al. 2019). Therefore, investigating further examples of stress drops of moderate earthquakes in this region would be useful to construct a model of the heterogeneous frictional property along the plate interface.

This study demonstrated that the detectability of a millimeter-scale tsunami was dramatically increased by S-net’s dense and wide pressure gauge network. The ability to resolve the slip distribution of offshore earthquakes was limited by sparse and remote tsunami networks (e.g., Williamson et al., 2018). Although the horizontal location could be reasonably constrained using only the remote tsunami data, the resolution was insufficient to constrain the detailed fault model (e.g., Inazu & Saito 2014; Kubota et al. 2019). In contrast, using the new dense and widely distributed tsunami network significantly improved the constraints on earthquake source parameters: in particular, the fault dimensions and stress drops, even for the moderate earthquake. This would greatly contribute to the source process estimation using the seismogram analysis, which is usually difficult to precisely resolve the fault dimension, especially for the moderate earthquakes, because of the trade-off due to the assumption of the rupture propagation velocity and of the signal-to-noise ratio. Although in the onshore region the InSAR data has played this important role to extract the spatial extent of the coseismic displacement and to
precisely constrain the fault dimension of moderate earthquakes (e.g., Massonnet et al., 1993; Feigl et al., 1995; Kobayashi, 2017), such observations have not been available in the ocean. The near-field dense and wide tsunami network has the great advantage to estimate the initial sea-surface displacement distribution and thus the fault dimension for offshore moderate earthquakes, as we demonstrated. Our results promise that the tsunami records from the nearest-field pressure gauge array data will enable us to obtain the relation between the stress drop and magnitude with the wider magnitude range and spatial extent.

Because some stations near the epicenter can contain large pressure offset signals irrelevant to tsunamis, we must be careful to analyze the signals to distinguish whether such signals are the real ones or the artefacts (Kubota, Suzuki et al., 2018; Tanioka, 2018). Nonetheless, the S-net data show significant promise for constraining earthquake source processes and stress drops, to elucidate mechanical properties along the fault in the Tohoku subduction zone.

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DART data was downloaded from the website of the National Data Buoy Center (NDBC) of the National Oceanic and Atmospheric Administration (NOAA), USA (https://www.ndbc.noaa.gov/dart.shtml). We thank Takanori Matsuzawa for the very-low-frequency-earthquake catalog (also available in Nishikawa et al., 2019) and Sachiko Tanaka for the low-frequency-tremor catalog (Tanaka et al., 2019). The regular earthquake catalog was acquired from the F-net NIED earthquake mechanism catalog (https://doi.org/10.17598/nied.0005; http://www.fnet.bosai.go.jp/event/search.php?LANG=en).

We used the JTOPO30 bathymetry model by the Marine Information Research Center of the Japan Hydrographic Association (http://www.mirc.jha.jp/en/index.html) and the plate boundary model by Iwasaki et al. (2015) (http://evrrss.eri.u-tokyo.ac.jp/database/PLATEmodel/PLMDL_2016/README_en.pdf). The discussion with Ryota Hino, Ryota Takagi, and Naoki Uchida and the members of the Research Center for Prediction of Earthquakes and Volcanic Eruptions at Tohoku University was fruitful.
References


Hino, R., Tanioka, Y., Kanazawa, T., Sakai, S., Nishino, M., & Suyehiro, K. (2001). Micro-tsunami from a local interplate earthquake detected by cabled offshore tsunami observation...

https://doi.org/10.1029/2001GL013297


https://doi.org/10.1109/JOE.2002.1002471


https://doi.org/10.1002/2013JB010892


Nakamura, T., & Hayashimoto, N. (2019). Rotation motions of cabled ocean-bottom seismic
stations during the 2011 Tohoku earthquake and their effects on magnitude estimation for
https://doi.org/10.1093/gji/ggy502

energetic 2010 MW 7.1 Solomon Islands tsunami earthquake. *Geophysical Journal

E/M0 discriminant for tsunami earthquakes. *Journal of Geophysical Research, 103*,
26885–26898. https://doi.org/10.1029/98JB02236

Nishikawa, T., Matsuzawa, T., Ohta, K., Uchida, N., Nishimura, T., & Ide, S. (2019). The slow
earthquake spectrum in the Japan Trench illuminated by the S-net seafloor observatories.

Tokachi-oki tsunami source: in-situ measurements and 3-D numerical modelling. *Natural

the Seismological Society of America, 82*(2), 1018–1040.

https://doi.org/10.1007/978-4-431-56850-6


Introduction

The procedure and result for the inversion for the initial sea-surface height is described in Text S1. Text S2 discusses the uncertainty of the stress drop estimation attributed to the depth uncertainty of the plate boundary model and the trade-off between each of the fault model parameters.

Figure S1 is the filtered pressure gauge waveforms. Figure S2 shows the tsunami dispersion captured by the S-net pressure gauges. The time series at the station S4N10, nearest to the focal area, is shown in Figure S3. Result of the forward tsunami simulation is shown in Figure S4. Figures S5 is the tsunami source distribution estimated from all pressure data. Figure S6 is the result of the recovery test for the inversion for the tsunami source. Figure S7 is the tsunami source distribution from only the KPG and DART data. Figure S8 shows the result of the grid-search for the fault horizontal location. Figure S9 is the simulation result for horizontally shifting the fault location. Figures S10 and S11 are the comparison of the simulated waveforms with the fault width and length changed, respectively. Figures S12 and S13 are the comparison of the waveforms when the fault is deepened and shallowed, respectively. Figure S14 shows the VR and fault parameters searched in the grid-search for the fault dimension.
Text S1

This text describes the detailed procedure and result of the inversion for the spatial distribution of the initial sea-surface height (tsunami source). The procedure is similar to that described in Kubota, Suzuki et al. (2018). Unit source elements of seafloor vertical displacements with a dimension of $8 \times 8$ km were distributed with the horizontal spacing of 4 km. The sea-surface elevation due to the unit source element was calculated considering the spatial smoothing effect due to the deep seawater (e.g., Saito, 2019) (seawater depth of $H_0 = 4$ km was assumed). The calculation scheme of the tsunami Green’s function is identical to that in the forward simulation. In the inversion, the smoothing constraint for the initial sea-surface height distribution was imposed, and its weighting was determined by trial and error. The time-window used for the inversion was manually determined based on the result of forward simulation as described above (thick black lines in Figure S5). After we obtained the solution, we evaluated the solution based on the similarity of the tsunami waveforms between the observation and the simulation from the model, in terms of the variance reduction (VR):

$$VR = 1 - \frac{\sum_{i=1}^{N} (x_{i}^{obs} - x_{i}^{cal})^2}{\sum_{i=1}^{N} x_{i}^{obs^2}},$$

where $x_{i}^{obs}$ and $x_{i}^{cal}$ are the observed and calculated tsunami waveforms of the $i$th data sample, respectively ($N$ is the total number of the data samples). We did not use the pressure gauges near the epicenter which seem to be affected by the large apparent pressure changes due to the seismic strong motion (e.g., S4N10 and S4N22).

As a result, the inversion analysis provided a distribution that paired uplift and subsidence (Figure S5). The tsunami source had a spatial extent of ~40 km and was located ~10 km to the west of the GCMT solution. The main features of the tsunami waveforms were reasonably reproduced (variance reduction of 66%, Figure S5b).

In order to assess the spatial resolution of the inversion analysis, we conducted the recovery test of the sea-surface height inversion (Figure S6). We assume the input sea-surface displacement distribution as:

$$\eta_0(x, y) = a_0 \exp \left[ -\left( \frac{x - x_0}{L_x} \right)^2 - \left( \frac{y - y_0}{L_y} \right)^2 \right]$$

where $a_0$ is the maximum displacement ($a_0 = 5$ cm was assumed), $(x_0, y_0)$ is the center of the displacement (we assumed to coincide with the center of the optimum fault model), and $L_x$ and $L_y$ are the horizontal dimension of the source (we assumed $L_x = L_y$). We inverted the simulated tsunami waveforms under the same condition as the original source inversion. As a result, the displacement was not reconstructed when $L_x = L_y = 4$ km was assumed, whereas that was reconstructed well when we assume $L_x = L_y \geq 8$ km. This indicating our inversion has a spatial resolution of 8 km.

We also evaluated how better the near-field S-net tsunami data increases the resolution of the tsunami source estimation than the far-field data, based on the
inversion using only the KPG and DART data, more than 150 km away (the other inversion settings are identical to the original). We obtained a centroid location similar to the original result (Figure S7). However, the distribution was spatially broader and the total volume of the displaced seawater \((4.0 \times 10^7 \text{ m}^3)\) was almost twice as large as that shown when the S-net data were added \((2.2 \times 10^7 \text{ m}^3)\). Therefore, we concluded that the horizontal location could be reasonably constrained from the remote pressure gauges, although the spatial resolution of the tsunami source was inferior to that obtained with the S-net data. The S-net data was essential to obtaining the higher-resolution tsunami source distribution and fault dimensions.

**Text S2.**

In order to evaluate the uncertainty of the stress drop estimation attributed to the fault depth uncertainty attributed to the uncertainty of the plate boundary model of Iwasaki et al. (2015), we calculated tsunami by changing the depth of the fault of ± 5 km from the optimum model (Figures S12 and S13). When supposing the fault deeper by 5 km, the later dispersive phases could not be reproduced at all even if we suppose the smallest fault dimension \((W = 1 \text{ km})\). If we suppose the shallow fault by 5 km, the fault should have a dimension of \(L \leq 20 \text{ km}\) and \(W \leq 10 \text{ km}\) to reproduce the observation, which corresponds to the stress drop of \(\Delta \sigma_s \geq \sim 0.5 \text{ MPa}\). This possible fault dimension and stress drop was almost identical to the original modeling \((15 \leq L \leq 20 \text{ km}, W \leq 7 \text{ km}, \text{ and } \Delta \sigma_s \geq \sim 0.7 \text{ MPa})\). This is probably because this event occurred at very shallow part of the plate boundary \(\sim 12 \text{ km}\) and the tradeoff between the fault dimension and depth was small.

In order to discuss the upper limit of the plausible stress drop, we further investigated the fault parameters and the VR values for all model candidates (Figure S14, arranged in descending order in terms of VR). We found the variability of the fault dimension parameters for the top 100 model candidates corresponded to the uncertainty of the fault dimension from the additional tsunami simulation (Figures S14d and S14e). We also found the top 100 candidates had almost identical and relatively high VR values (Figure S14a), which indicates that the top 100 candidates reasonably reproduce the observed waveforms. Based on the variability of the stress drop values within the top 100 fault model candidates, we concluded the plausible maximum stress drop was \(\sim 5 \text{ MPa}\) (Figure S14c). It seems that the uncertainties attributed to uncertainty of the fault length or width are not so large \((\text{e.g., uncertainty of fault length of } 10–15 \text{ km corresponds to the stress drop uncertainty of } 1.0–2.0 \text{ MPa})\). However, because of the trade-off between each of the model parameters, the expected uncertainty range of the stress drop became larger.
Figure S1. Pressure waveforms after the data processing. Light gray and red traces denote the de-tided and the band-pass filtered waveforms.
Figure S1. (Continued)
Figure S1. (Continued)
Figure S1. (Continued)
Figure S1. (Continued)
**Figure S2.** The tsunami waveforms at the stations S4N11 to S4N15. The main tsunami pulses and small subsequent dispersive waves are marked by green and gray arrows, respectively.
Figure S3. Time series at the station S4N10. Light gray, dark gray, and red waveforms are the de-tided, the low-pass filtered (100 s), and the band-pass filtered (100 – 1000 s) records, respectively.
Figure S4. Result of the forward tsunami simulation. (a) Spatial distribution of the initial sea-surface height. (b) Comparison of the observed (gray) and synthesized (red) waveforms.
Figure S4. (Continued)
Figure S5. Result of the inversion for the tsunami source. (a) Spatial distribution of the initial sea-surface height. (b) Comparison of the observed (gray) and synthesized (red) waveforms. The rectangle with dashed lines denotes the analytical area.
Figure S5. (Continued)
Figure S6. Result of the recovery test for the tsunami source inversion. Left panels are the input displacement distributions, assuming spatial extent $L_x$ of (top) 4 km, (center) 8 km and (bottom) 12 km (contour interval of 1 cm). Middle panels are the inverted distributions. Small dots are the locations of center of the unit source elements and the dashed lines denote the analytical area (also shown in Figure S5). Right panels show the comparisons of the displacement profile along $y = 26$ km (gray: input, blue: inverted).
Figure S7. Result of the inversion for the tsunami source without the S-net data. (a) Spatial distribution of the initial sea-surface height. (b) Comparison of the observed (gray) and synthesized (red) waveforms.
Figure S7. (Continued)
Figure S8. Result of the grid-search for the horizontal location. (a) Spatial distribution of the initial sea-surface height. Blue contours are the initial sea-surface height distribution estimated by the inversion analysis. (b) Comparison of the observed and simulated waveforms.
Figure S8. (Continued)
Figure S9. Comparison of the observed (gray) and simulated (red) waveforms at (a) S4N12 with the centroid location shifted along the east-west direction and (b) S4N09 with the centroid location shifted along the north-south direction. The timings of the maximum and minimum peaks were denoted by red arrows. The centroid locations are shown in the insets.
Figure S10. Comparison of the observed (gray) and simulated (red) waveforms with the fault width changed. The direction and azimuth are shown in each panel. The timings of the maximum and minimum peaks were denoted by red arrows.
Figure S10. (Continued)
Figure S10. (Continued)
Figure S10. (Continued)
Figure S10. (Continued)
Figure S10. (Continued)
Figure S10. (Continued)
Figure S10. (Continued)
Figure S11 Comparison of the observed (gray) and simulated (red) waveforms with the fault length changed. The direction and azimuth are shown in each panel. The timings of the maximum and minimum peaks were denoted by red arrows.
Figure S11. (Continued)
Figure S11. (Continued)
Figure S11. (Continued)
Figure S11. (Continued)
Figure S11. (Continued)
Figure S11. (Continued)
Figure S11. (Continued)
Figure S12. Comparison of the observed (gray) and simulated (blue) waveforms assuming the fault 5 km deeper than the optimum model, at (a) S4N12 with the width changed and (b) S4N09 with the length changed. The simulated waveforms from the optimum fault without changing depth is also shown (red).
Figure S13 Comparison of the observed (gray) and simulated (blue) waveforms assuming the fault 5 km shallower than the optimum model, at (a) S4N12 with the width changed and (b) S4N09 with the length changed. The simulated waveforms from the optimum fault without changing depth is also shown (red).
**Figure S14.** VR and fault parameters for all fault models in the grid-search for the fault dimension, arranged in descending order in terms of VR. The light green lines denote the maximum and minimum limit of the possible range of the fault dimension.