Comparison of the seafloor displacement from uniform and non-uniform slip models on tsunami simulation of the 2011 Tohoku–Oki earthquake

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ABSTRACT

The numerical simulations of recent tsunami caused by 11 March 2011 off-shore Pacific coast of Tohoku–Oki earthquake ($M_w$ 9.0) using diverse co-seismic source models have been performed. Co-seismic source models proposed by various observational agencies and scholars are further used to elucidate the effects of uniform and non-uniform slip models on tsunami generation and propagation stages. Non-linear shallow water equations are solved with a finite difference scheme, using a computational grid with different cell sizes over GEBCO30 bathymetry data. Overall results obtained and reported by various tsunami simulation models are compared together with the available real-time kinematic global positioning system (RTK-GPS) buoys, cabled deep ocean-bottom pressure gauges (OBPG), and Deep-ocean Assessment and Reporting of Tsunami (DART) buoys. The purpose of this study is to provide a brief overview of major differences between point-source and finite-fault methodologies on generation and simulation of tsunamis. Tests of the assumptions of uniform and non-uniform slip models designate that the average uniform slip models may be used for the tsunami simulations off-shore, and far from the source region. Nevertheless, the heterogeneities of the slip distribution within the fault plane are substantial for the wave amplitude in the near field which should be investigated further.

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1. Introduction

On March 11, 2011 (05:46:24 UTC, 14:46:24 JST), a massive earthquake occurred off the Honshu coast (approximately 130 km east of Sendai and 370 km northeast of Tokyo) on a shallow portion of the subduction zone where the oceanic Pacific Plate subducts beneath the North American plate along the margin of the Japan Trench (Fig. 1). The earthquake generated a tsunami that was observed all over the Pacific region and caused enormous demolition in coastal Japan. This is the fourth largest earthquake in the world and the largest in Japan since instrumental recordings began in 1900. The tsunami is also most deadly and destructive event caused by an earthquake after the 2004 Sumatra earthquake. This devastating tsunami mostly affected coastal districts of Iwate, Miyagi, Fukushima, Ibaraki and Chiba Prefectures, which border the Pacific Ocean. Maximum run-up heights greater than 10 m are distributed along 425 km of coast and maximum run-up heights greater than 20 m are distributed along 290 km of coast, in direct distance (Mori et al., 2011). The disaster resulted in 16.131 fatalities, 3240 missing, 5994 injured and 128.497 house collapses as of 13 January 2012 including the toll from the aftershocks and triggered events (FDMA, 2012). The epicenter was located at Lat. 38.297°N, and Lon. 142.372°E (USGS, 2011a). The moment magnitude determined by JMA (Japan Meteorological Agency) and USGS (United States Geological Survey) was 9.0 with a depth of 23.7 and 30 km respectively. There had been hundreds of aftershocks, many greater than magnitude 5 (Fig. 1). The largest of the aftershocks was $M_w$ 7.7. The main shock is much larger than the expected earthquakes on the asperities of the Off-Miyagi Prefecture (Yamana and Kikuchi, 2004). Although few events greater than magnitude 8.0 occurred in the last several hundred years in the region, none of them triggered such enormous wave heights as this earthquake. In addition to the effects of the huge waves, the other serious effect of this earthquake arose from the Nuclear reactors at the Fukushima which was located close to Pacific Ocean to get the required cooling capacity for the steam condensate. The tsunami swept over the sea wall designed to protect the Fukushima Nuclear Plant and knocked out cooling system causing the reactors to melt down and numerous explosions. The sea walls proved to be no match for the massive tsunami generated by the March 11, 2011 Tohoku–Oki earthquake. However, they might be a time for people to evacuate and also stopped the full force of energy of the tsunami waves. This situation is an indication of the re-evaluation of tsunami simulations to understand the rupture processing and mechanism of tsunamis considering the future earthquakes in the same...
area. Furthermore, it is much more important than past, as a result of increasing population due to the economic development of the coastal provinces of Japan.

Tsunamis are very large ocean or sea wave triggered by various large scale disturbances of the ocean floor such as submarine earthquakes, volcanic activities or landslides (Yolsal et al., 2007). The lengths of tsunami waves may be up to a two hundred kilometers from one wave crest to another. Although they have relatively small heights off-shore, they may become enormous as they approach to shallow coastal areas. Most tsunamis are triggered along trenches and are typically caused by thrust-type subduction zone earthquakes. Subduction zones are convergent plate boundaries characterized geomorphologically by deep ocean trenches and island arcs or continental margins, seismically by landward dipping Wadati–Benioff zones of deep earthquakes, tectonically by regional-scale crustal faulting and terrane movements, and magmatically by arcuate and linear belts of eruptive centers, the so-called volcanic front (Zhao, 2001).

Many subduction zones are characterized by the occurrence of great underthrusting earthquakes (Ruff, 1992). These areas have regularly been the source for about magnitude 8 or more earthquakes in the history. Most large earthquakes that spawn tsunamis were known to occur in these areas, including such as Japan, Chile, Mexico, Alaska, Aleutians, Kamchatka, Cascadia, Sumatra, Philippines, Tonga and Hellenic Arc. However, it should be noted that large earthquakes do not occur uniformly throughout the all worldwide convergent margin subduction zones.

The regional stress distributions in the subduction zones effect the deformation area of ocean floor to generate a significant series of tsunami waves affected coastlines and islands. Hence, regional variations in the rupture characteristics of large thrust-type subduction zone earthquakes in the different subduction zones should be examined for estimating the threat of tsunamiogenic earthquakes. Lay et al. (1982) indicated that there are four fundamental categories of behavior of systematic variations in maximum rupture extents in different subduction zones. These are (1) the Chile-type regular occurrence of great ruptures spanning more than 500 km; (2) the Aleutian-type variation in rupture extent with occasional rupture reaching 500 km in length, and temporal clustering of large events; (3) the Kurile-type repeated failure over a limited zone of 100–300 km in length in isolated events; and (4) the Marianas-type absence of large events. Chile-type convergent margins have many large (M > 7) earthquakes, whereas Marianas-type ones do not (Cloos and Shreve, 1996). The rupture processes associated with each category have distinctive features and it is assumed that the degree of the seismic coupling between the downgoing and overriding plates control the maximum earthquake rupture dimensions in each subduction zone (Lay et al., 1982). The criteria for these classifications were also related to the nature of the back arc regions for both continental and island arcs (Uyeda and Kanamori, 1979). The island arcs were grouped into those having inactive back arc basins and active back arc basins. The Chilean type have no back arc basins. One of the features to be noted is that the Japan arc shows the features of Chilean-type, because the Sea of Japan is not actively spreading, but was probably a Mariana-type when the Sea of Japan had been formed (Uyeda and Kanamori, 1979; Lay et al., 1982). Here, the most well-known worldwide tsunamis caused by thrust-type subduction zone earthquakes were depicted in Table 1. Several authors have provided the source characteristics of these tsunamiogenic earthquakes, rupture descriptions, fault plane solutions, tsunami simulations and the links between earthquake rupture of these special events and subduction zone structures (Table 1). Although most large earthquakes occur along subduction zones, massive earthquakes of magnitude greater than Mw 9.0 have been known to occur in Chile, Alaska, Kamchatka and Sumatra (Ozawa et al., 2011) until 2011 Tohoku–Oki earthquake except for the Jogan earthquake of 13 July 869. The Jogan earthquake of 13 July 869 may be the only documented event to have occurred with a possible magnitude and location similar to that of the 2011 earthquake (Simons et al., 2011). Given the lack of historical earthquakes of the Mw 9.0 size, the Tohoku–Oki earthquake caught most seismologists by surprise (Ritsema et al., 2012). However, the Pacific coast of Japan has experienced tsunamis from the earthquakes around the Pacific Ocean several times in this century (Watanabe, 1985; Satake et al., 1996; Tanioka et al., 1997). The subduction of the Pacific Plate along the Japanese Trench is responsible for many large underthrusting earthquakes and giant tsunamis in that region. The Tohoku–Oki earthquake occurred off the Pacific coast of Honshu in the NE Japan forearc region of the subduction zone.
which is the most tectonically active setting in the region. Many great (M 8.0–8.5) interpolate earthquakes and microearthquakes show a wide distribution in the forearc region from the Japan Trench to the Pacific coast suggesting strong interpolate seismic coupling in that region (Zhao et al., 1997). Understanding the processes that govern the seismic cycle of subduction margins in the region has been a major research by many researchers and it was well studied with various seismological approaches (Utsu, 1971; Kanamori, 1971; Matsuzawa et al., 1990; Zhao et al., 1992, 1997, 2011; Nakajima et al., 2001; Hasegawa et al., 2005, 2009; Kita Kanamori, 1971; Matsuzawa et al., 1990; Zhao et al., 1992, 1997, 1999; Sugawara et al., 2010, 2012), Minoura et al. (2001), Satake et al. (2008), and Watanabe (2000)

Table 1

<table>
<thead>
<tr>
<th>Year</th>
<th>Earthquake location</th>
<th>Estimated magnitude</th>
<th>Tectonic plate interactions</th>
<th>References</th>
</tr>
</thead>
<tbody>
<tr>
<td>1988</td>
<td>1868 Arica</td>
<td>Mw 8.0</td>
<td>Pacific–North American</td>
<td>Okal et al. (2006), Dorbath et al. (1990), and Lomnitz (1970)</td>
</tr>
<tr>
<td>1989</td>
<td>1877 Iquique</td>
<td>Mw 8.0</td>
<td>Nazca–South American</td>
<td>Kulikov et al. (2005), Dorbath et al. (1990), and Comte and Pardo (1991)</td>
</tr>
<tr>
<td>1996</td>
<td>1960 Chile</td>
<td>Mw 8.5</td>
<td>Nazca–South American</td>
<td>Nelson et al. (2009), Moreno et al. (2009), Satake and Atwater (2007), Lange et al. (2007), Cisternas et al. (2005), Okal (2005), Barrientos and Ward (1990), and Kanamori and Cipar (1974)</td>
</tr>
<tr>
<td>2003</td>
<td>2004 Sumatra</td>
<td>Mw 9.2</td>
<td>India</td>
<td>McCaffrey (2009) and Konca et al. (2008)</td>
</tr>
<tr>
<td>2010</td>
<td>2010 Chile</td>
<td>Mw 8.8</td>
<td>Nazca–South American</td>
<td>Moreno et al. (2012), Lay (2011c), Moscoso et al. (2011), Lorito et al. (2011), Lay et al. (2010), and Moreno et al. (2010)</td>
</tr>
</tbody>
</table>

which is the most tectonically active setting in the region. Many great (M 8.0–8.5) interpolate earthquakes and microearthquakes show a wide distribution in the forearc region from the Japan Trench to the Pacific coast suggesting strong interpolate seismic coupling in that region (Zhao et al., 1997). Understanding the processes that govern the seismic cycle of subduction margins in the region has been a major research by many researchers and it was well studied with various seismological approaches (Utsu, 1971; Kanamori, 1971; Matsuzawa et al., 1990; Zhao et al., 1992, 1997, 2011; Nakajima et al., 2001; Hasegawa et al., 2005, 2009; Kita et al., 2006; Zhao, 2009; Wang and Zhao, 2010; Tong et al., 2012).

2. Tsunami source models

Tsunami source is seafloor deformation caused by slip on a fault during an earthquake (Fujii and Satake, 2007). In the case of a seafloor deformation by an earthquake, the water column is disturbed...
by the uplift or subsidence of the seafloor causing generation of a tsunami. The intensity of the generated tsunami wave depends on the size and impact of the source mechanism on the displaced water (Yalciner et al., 2002). Thus, the accurate prediction of maximum heights and tsunami arrival times relies mainly on the estimation of earthquake source parameters and fault plane models. The impulsive fault plane models assume that the seafloor deforms instantaneously and the entire fault line ruptures simultaneously (Wang and Liu, 2006). Assumptions in calculating the seafloor deformation define the initial size and height for tsunami propagation. The main factor which determines the initial size and height of a tsunami is the amount of vertical sea floor deformation. In this common approach, it is expected that the initial sea surface displacement mimics the same form as the seafloor displacement (Piatanesi et al., 2001; Yalciner et al., 2004; Gica et al., 2007; Titov et al., 2005a,b; Okal et al., 2009; Yolsal and Taymaz, 2010; Yolsal-Cevikbilen and Taymaz, 2012). This approach may be controlled by the magnitude and depth of earthquake, fault characteristics, rupture characteristics, abrupt increasing due to splay faulting and coincident sliding of sediments. Then, the approach is based on solving the hydrodynamics equations with boundary conditions at the ocean floor corresponding to a static displacement caused by the earthquake source (Yanovskaya et al., 2003). Thus, it is necessary to identify the magnitude, depth, location of epicentre, size of the fault and source parameters of tsunamiogenic earthquakes for the propagation time and the final amplitude of tsunami at a site due to submarine earthquakes (Fig. 2).

In reality, the epicenter and the magnitude of the earthquake parameters can be difficult to determine and may remain unknown (Synolakis, 2003). In these cases, the size of the source area can be estimated by using empirical formulas among moment magnitude, rupture length and rupture width which are compiled from the fault parameters of historical earthquakes. However, the larger magnitude of the earthquakes (M > 8.0) might be out of the magnitude range used in fitting procedures of the empirical relationships (e.g., Wells and Coppersmith, 1994; Utsu et al., 2001; Papazachos et al., 2004). For this purpose, the aftershock distributions which are expected to cluster within the slip zone are used to delineate earthquake rupture area. The more reliable method to delineate earthquake rupture area is the finite fault model that could be constructed through inversion of teleseismic, strong motion, tsunami and geodetic datasets (Kikuchi and Kanomori, 1982, 1996; Hartzell and Heaton, 1986; Satake, 1987; McCaffrey and Abers, 1988; Taymaz et al., 1990, 1991; Johnson et al., 1996; Ji et al., 2002; Benetatos et al., 2004; Tan and Taymaz, 2006; Weinstein and Lundgren, 2008; Yolsal and Taymaz, 2010; Lay et al., 2011b; Koketsu et al., 2011; Yokota et al., 2011; Romano et al., 2012; Crowell et al., 2012; Tsushima et al., 2012). For this purpose, the earthquake parameters and source characterizations of the Tohoku–Oki earthquake issued by different researchers and organizations that would be an input for the tsunami simulations of the earthquake were gathered (Tables 2 and 3). Here, static vertical deformation of the seafloor resulting from the motion of the fault of the 11 March 2011 Tohoku–Oki earthquake was calculated for the fault models by using an algorithm developed by Okada (1985).

In Tohoku–Oki earthquake, the width and length of the fault plane (LFP) was suggested from the 150 km to 240 width km and from 450 to 640 km length due to the several inversion techniques (Yagi and Nishimura, 2011; Ammon et al., 2011; Koketsu et al., 2011; Hayes, 2011; Fujii et al., 2011; Imamura, 2011; Yoshida et al., 2011; Lay et al., 2011a; Honda et al., 2011; Ritsema et al., 2012). By defining the aftershock area with a rectangular region enclosing the main shock and aftershocks, the size of the surface projected rupture area for this event was assumed as 200 km wide, 500 km long. The aftershock sequence included three aftershocks with magnitudes of 7.0 or larger, nearly 40 with magnitudes between 6.0 and 7.0, and over 350 events in the magnitude range from 5.0 to 6.0 (Ammon et al., 2011). The aftershocks occurred at an edge of the relatively large slip area of the mainshock (Yoshida et al., 2011). It should be note that the earthquake rupture area are delineated from aftershock distribution is the surface projected earthquake rupture area. Thus, the width of the fault plane (WFP) for this event might be estimated as in the following equations due to the measured width yielded by aftershock distribution (e.g. Liu et al., 2009):

$$WFP = \frac{MS}{\cos(\delta)}$$

where WFP is width of fault plane, MS is measured width, and $\delta$ is dip angle. The top of the fault (TOF) was calculated as follows:
wave data and tsunami data (Imamura et al., 1994; Ohmachi et al., slip models and also between the fault rupture models from seismic there might be differences between the uniform and non-uniform on maximum tsunami heights. Moreover, it is often the case that fort to examine and compare of different source and rupture models in this study, it was used a range of uniform and non-uniform slip models for tsunami simulations. This process might be used which are well suited for trans-0001). Adopted the hypocentral depth estimates for the initiation of the rupture process that may render tsunami source estimate for the possible large earthquakes. Clearly, assuming uniformly dis-

Source parameters and focal mechanism solutions of Candidate Sources (CS) used for uniform slip models for tsunami simulations.

<table>
<thead>
<tr>
<th>Source parameters</th>
<th>CS-1 (USGS, 2011a,b)</th>
<th>CS-2 (Shao et al., 2011)</th>
</tr>
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<tbody>
<tr>
<td>$M_w$</td>
<td>9.1</td>
<td>9.0</td>
</tr>
<tr>
<td>Seismic moment</td>
<td>5.31</td>
<td>5.06</td>
</tr>
<tr>
<td>Depth (km)</td>
<td>30</td>
<td>24</td>
</tr>
<tr>
<td>LFP (km)</td>
<td>500</td>
<td>500</td>
</tr>
<tr>
<td>WFP (km)</td>
<td>203</td>
<td>203</td>
</tr>
<tr>
<td>Strike, dip, rake</td>
<td>203/10/88</td>
<td>199/10/92</td>
</tr>
<tr>
<td>TOF (km)</td>
<td>12.36</td>
<td>6.37</td>
</tr>
<tr>
<td>Average slip (m)</td>
<td>13.08</td>
<td>12.46</td>
</tr>
<tr>
<td>Max-vertical dislocation (m)</td>
<td>5.35</td>
<td>5.82</td>
</tr>
<tr>
<td>Min-vertical dislocation (m)</td>
<td>–2.38</td>
<td>–2.37</td>
</tr>
</tbody>
</table>

WFP, TOF and average slip was calculated from Eqs. (1)–(3) respectively Seismic moment scale: $10^{22}$ Nm.

TOF $= h - (\sin \delta) d$  

where TOF is top of the fault in km, $h$ is the hypocentral depth of earthquake, $\delta$ is dip angle and $d$ is the half of fault WFP. It was adopted the hypocentral depth estimates for the initiation of the earthquake is 24 km (JMA, 2011) and 30 km (Ammon et al., 2011; USGS, 2011a,b) respectively. The hypocenter was assumed at the center of the fault. It should also be note that a centroid location, which is a release point of seismic energy mainly, is not always the same as the hypocenter. Therefore, the hypocentral depths were used instead of centroid depths to define the size of rupture area. Then, the remained parameters as it was pointed out before were examined and compiled to provide inputs assuming that the slip is distributed uniformly. This is the case that many tsunami models assume that the slip is distributed uniformly over the entire rupture. This process might be used which are well suited for trans-oceanic tsunamis for early warning strategies. In contrast, however, tsunami models assuming uniformly distributed slip are far not enough to estimate the near field coastal tsunami heights for especially in the case of 26 December 2004 Indonesia, 28 February 2010 Chile and 11 March 2011 Japan earthquakes. In these regions, the subduction takes place only 100–200 km off the coast which might be a smaller distance than that of the potential rupture size for the possible large earthquakes. Clearly, assuming uniformly distributed slip is not the real case, and there are often complications with the rupture process that may render tsunami source estimate procedure inadequate (Weinstein and Lundgren, 2008). Therefore, in this study, it was used a range of uniform and non-uniform slip models to perform tsunami simulations of the earthquake in an effort to examine and compare of different source and rupture models on maximum tsunami heights. Moreover, it is often the case that there might be differences between the uniform and non-uniform slip models and also between the fault rupture models from seismic wave data and tsunami data (Imamura et al., 1994; Ohmachi et al., 2001)

Numerical tsunami simulations by using the static vertical deformation algorithm were performed in five stages by considering the different source and rupture models proposed by different researchers. The different sources used for the tsunami simulation of the earthquake will be mentioned as CS (Candidate Source) herein after in this paper (Table 2): (1) the Global Centroid Moment Tensors (GCMT) using long period surface waves issued by National Earthquake Information Center (NEIC) of the United States Geological Survey (USGS, 2011b), (2) Multiple Double Couples (MDC) analysis conducted using 1-h long period seismic waves (Shao et al., 2011), (3) USGS finite fault model (Hayes, 2011), (4) Tsunami source area constructed from tsunami wave-form inversion (Fujii et al., 2011), (5) Tsunami source area constructed from tsunami waveform inversion (Imamura, 2011).

The first CS was adapted to the tsunami simulations by using the source parameters of Global Centroid Moment Tensor Project Moment Tensor Solution (GCMT) results. The GCMT results include the quantification of earthquake source characteristics on a global scale which is rapidly determined by moment tensors for earthquakes with $M > 5.5$ globally. Initial manual CMT results for this event were obtained within 23 min of origin time and fully automatic results were distributed by e-mail within 33 min (Polet and Thio, 2011). The strike, dip and rake of the fault derived from the GCMTs for large worldwide earthquakes using long period surface waves were given as 203°, 10°, 88° for the nodal plane1 (NP1) and as 25°, 80°, 90° for the nodal plane2 (NP2) respectively (USGS, 2011b). The NP1 was choosen as the main nodal plane by taking into account the known tectonic features of the subduction in the region. The second CS was carried out by using Multiple Double Couples (MDC) analysis derived rapidly from teleseismic body and surface waves of the earthquake yielding a single double couple, whose low angle nodal plane orients $199^\circ$, dips 10°, and rakes $92^\circ$ (Shao et al., 2011). The moment tensors indicate the orientation of the fault and the strength and direction of shearing motion on the fault (Ritsema et al., 2012). The third CS was adapted by using the finite fault model of the earthquake issued by the USGS (Hayes, 2011), based on static and seismic data inversion algorithm (Ji et al., 2002). This solution provides multiple fault segments divided into 325 subfaults (25 km × 25 km). The fourth CS was adapted tsunami simulations by using tsunami source area divided into 40 subfaults (50 km × 50 km) constructed from tsunami waveform inversion (Fujii et al., 2011). The fault plane solution determined from USGS W-phase moment tensor inversion (Hayes, 2011) (strike 193°, dip 14° and rake 81°) was considered for every subfaults in this tsunami source area. The fifth CS was adapted to tsunami simulations by using tsunami source area divided into 10 subfaults (100 km × 100 km) (Imamura, 2011). Strike 193°, dip 14° and rake 81° were also considered from the USGS’s W-phase moment tensor solution to construct the tsunami source area as in the former model. Slip distributions estimated by teleseismic and tsunami waveform inversions are displayed in Fig. 3.

The first two CS models specify the focal geometry (strike, dip, rake) and the scalar moment, but not the slip distribution along the fault; consequently an average uniform slip over the entire fault area was calculated for these models. The average uniform slip was calculated using the following equations for these models:

### Table 2

Source parameters and focal mechanism solutions of Candidate Sources (CS) used for uniform slip models for tsunami simulations.

<table>
<thead>
<tr>
<th>Source parameters</th>
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<tr>
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<td>30</td>
<td>24</td>
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<tr>
<td>LFP (km)</td>
<td>500</td>
<td>500</td>
</tr>
<tr>
<td>WFP (km)</td>
<td>203</td>
<td>203</td>
</tr>
<tr>
<td>Strike, dip, rake</td>
<td>203/10/88</td>
<td>199/10/92</td>
</tr>
<tr>
<td>TOF (km)</td>
<td>12.36</td>
<td>6.37</td>
</tr>
<tr>
<td>Average slip (m)</td>
<td>13.08</td>
<td>12.46</td>
</tr>
<tr>
<td>Max-vertical dislocation (m)</td>
<td>5.35</td>
<td>5.82</td>
</tr>
<tr>
<td>Min-vertical dislocation (m)</td>
<td>–2.38</td>
<td>–2.37</td>
</tr>
</tbody>
</table>

WFP, TOF and average slip was calculated from Eqs. (1)–(3) respectively Seismic moment scale: $10^{22}$ Nm.

TOF $= h - (\sin \delta) d$  

### Table 3

Fault model parameters estimated from seismic-wave and tsunami inversion of Candidate Sources (CS) used for non-uniform slip models for tsunami simulations.

<table>
<thead>
<tr>
<th>Source parameters</th>
<th>CS-3 (Hayes, 2011)</th>
<th>CS-4 (Fujii et al., 2011)</th>
<th>CS-5 (Imamura, 2011)</th>
</tr>
</thead>
<tbody>
<tr>
<td>$M_w$</td>
<td>9.0</td>
<td>9.0</td>
<td>9.0</td>
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<tr>
<td>Seismic moment</td>
<td>4.9</td>
<td>3.8</td>
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<td>Depth (km)</td>
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<td>LFP (km)</td>
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<td>500</td>
</tr>
<tr>
<td>WFP (km)</td>
<td>260</td>
<td>200</td>
<td>200</td>
</tr>
<tr>
<td>Strike, dip, rake</td>
<td>195/10/V</td>
<td>193/14/81</td>
<td>193/14/81</td>
</tr>
<tr>
<td>TOF (km)</td>
<td>7.54</td>
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<td>1</td>
</tr>
<tr>
<td>Average slip (m)</td>
<td></td>
<td></td>
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</tr>
<tr>
<td>Maximum slip (m)</td>
<td>33.47</td>
<td>47.93</td>
<td>20</td>
</tr>
<tr>
<td>Minimum slip (m)</td>
<td>0.05</td>
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<td>1</td>
</tr>
<tr>
<td>Number of subfault</td>
<td>325</td>
<td>40</td>
<td>10</td>
</tr>
<tr>
<td>Subfault area (km)</td>
<td>$25 \times 25$</td>
<td>$50 \times 50$</td>
<td>$100 \times 100$</td>
</tr>
<tr>
<td>Max-vertical dislocation (m)</td>
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<td></td>
</tr>
<tr>
<td>Min-vertical dislocation (m)</td>
<td>–1.95</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

V: Variable rake values for every subfaults. Seismic moment scale: $10^{22}$ Nm.
where \( \mu \) is the rigidity of earth crust, \( S \) is the amount of average slip motion (slip) and \( L \) is the length of the fault plane and \( W \) is the width of the fault plane, \( M_0 \) is the scalar moment of an earthquake and \( M_w \) is the moment magnitude of an earthquake (Aki, 1966; Hanks and Kanamori, 1979). The value of the rigidity varies with location and ranges usually from 1.0 to \( 6.0 \times 10^{11} \) dyn cm \(^{-2} \) due to the geological properties of a location (Gica et al., 2007). In this study, the crustal rigidity was used a typical value of \( 4.0 \times 10^{11} \) - dyn cm \(^{-2} \) for the Pacific Rim regions (Johnson, 1999). For the rest of the CS models (third, fourth, fifth), the prescribed non-uniform slip distributions along the divided subfaults carried out for tsunami simulations. It should be noted that the calculation of average slip via Eq. (3) has limitations due to the correct estimation of fault width and fault length for up-dip fault motions. However, source inversion results of the earthquake show that rupture behavior of the earthquake developed with a maximum slip over 40 m (Ammon et al., 2011; Shao et al., 2011; Fujii et al., 2011; Lay et al., 2011; Lee et al., 2011) and with the average slip over the entire fault model about 15–20 m (Lay et al., 2011b; Lee et al., 2011). Most of the inversions have in common that the earthquake was caused by a fault rupture extending to a shallow part of the subduction zone at the Japan trench and maximum slip is located in the area east off-shore from Minamisanriku in the South to Miyako in the North (Fujiiwara et al., 2011; Løvholt et al., 2012). This is also indicated by various geodetic and joint inversion procedures and bathymetric surveys across the trench in the rupture zone before and after the earthquake (Ide et al., 2011; Fujiwara et al., 2011; Ozawa et al., 2011; Romano et al., 2012). For this reason, non-uniform rupture models were examined by considering the cause of higher tsunami waves from the large slips occurred in the shallow portion of the plate interface near the trench axis (Maeda et al., 2011; Saito et al., 2011).

As mentioned earlier, here, the static vertical deformation of the seafloor for the different CS models by using an algorithm developed by Okada (1985) taking the vertical component of the displacement field. For the finite fault rupture models, the resulting co-seismic vertical bottom displacement was calculated by assuming superposition of subfaults (e.g., Duttykh et al., 2012; Grilli et al., 2012). The vertical dislocations with a 2D view were presented in Fig. 4. Whereas Fig. 4a and b represent simplified cases with uniform slips; the other calculations are all non-uniform slips. The maximum vertical dislocations of sea floor, in other words, the maximum sea surface responses are 5.35, 5.16, 8.76, 12.55, 15.84 m for the CS-1, CS-2,

\[
M_0 = \mu SLW
\]

\[
M_w = \frac{2}{3} \log_{10} M_0 - 10.7
\]
CS-3, CS-4 and CS-5 source models respectively. The wider deformation area can be seen in uniform models. The wider deformation area can also be seen in finite fault model (USGS, 2011a,b) compared to the models estimated from inversions of tsunami datasets (Fujii et al., 2011; Imamura, 2011). However, the wider negative displacement area is present along the coast in the rupture models estimated from tsunami inversions.

3. Tsunami propagation model

Since tsunami wavelengths are much larger than the ocean depth, tsunamis are considered long or shallow water gravity waves (Hebert et al., 2001). A tsunami wave becomes a shallow water wave when the ratio between the water depth and its wave length gets very small. Thus, tsunami numerical models might be based on non-linear shallow-water equations. In this study, the propagation of the tsunami was numerically modeled by using the SWAN (Simulating WAVes Near-Shore) code which solves the non-linear long wave equations of the fluid flow (Mader, 1988, 2001). The model uses finite difference scheme in time which includes Coriolis and frictional effects. The set of equations to be solved are:

\[ u_t + \frac{1}{\cos(\varphi)} u_u + u u_x + \frac{g}{\cos(\varphi)} \eta_x = f v - \frac{g U |u|}{C^2 (D + \eta)} \]  

(5)

Fig. 4. Vertical dislocations of the sea floor and cross section of AB due to different source models.
\[ v_t + \frac{1}{\cos(\phi)} u v_x + u v_y + g \eta_y = -fu - \frac{g |U| v}{C(D + \eta)} \] (6)

\[ u_t + \frac{1}{\cos(\phi)} \left( \left( \eta + D \right) u_x + \left( \left( \eta + D \right) v \cos(\phi) \right) \right)_y = 0 \] (7)

where \( \phi \) is the latitude, \( u \) and \( v \) are the \( x \) and \( y \) components of the velocity \( U \), \( g \) is the gravitational acceleration, \( t \) is the time, \( \eta \) is the wave height above the mean water level, \( f \) is the Coriolis parameter, \( C \) is the coefficient for bottom stress, \( D \) is the depth, and indexes refer to partial derivatives. The model has also been implemented widely by different researchers to simulate tsunami propagations and wave heights (Baptista et al., 2003; Xie et al., 2012; Kaabouben et al., 2008; Franchello and Annunziato, 2012; Gonzalez et al., 1991; Ulutas, 2011). Here, calculations were performed in geographical coordinates and GEBCO_08 Grid (GEBCO-BODC, 2012) bathymetric data set was used. The data set values within the GEBCO_08 Grid represent elevation in meters, with negative values for bathymetric depths and positive values for topographic height and the values cover a 30 arcsec grid of global elevations.

**Fig. 5.** Location of the buoys and the cross section of the bathymetry in covered area A and B.

The sea depth is about 7000 km at the Japan Trench while that in the point Q is about 30 m around Sendai Coastal Area (Fig. 5). The depth difference between shallow and deep areas is significant. It means that non-linearity and bottom friction will govern the transformation of tsunami waves from open Pacific Ocean to coastal areas of Japan contrary to the non-linear and dispersive effects are extremely small and can be neglected in open ocean (Gica et al., 2007). In this case, shallow water equation model can be nested with different grid size from coarser grid to finer grid. Thus, above mentioned SWAN code was chosen for computational analysis of the tsunami simulation and different grid calculations were used for simulating the tsunami.

### 4. Computational analysis, comparisons and results

In this study, two levels of computation grids for proper amplification were used. The grids consist of a 2° grid describing the open ocean and 0.25° grid describing the coasts of Honshu. The computation area was set in two parts as A, B (Fig. 5). The first area covers 28.0–50.0°S and 135.0–160.0°E, on a re-sampled 2 arcmin
grid, the second area covers 36.5°–40.5°S and 140.0°–144.0°E, on a re-sampled 0.25 arcmin grid. The duration of tsunami simulation was 3 h for the computational grid of 2° and 2 h for the computational grid of 0.25°. Numerical tsunami simulations were computed from the proposed fault models and assumptions as summarized in Tables 2 and 3 of the fault parameters of March 11, 2011 Mw 9.0 Tohoku–Oki earthquake. The maps of maximum tsunami heights were shown in Figs. 6 and 7 due to the different parameters and rupture models. The highest estimated wave heights are 9.49 m, 8.50 m, 17.13 m, 14.97 m and 18.81 m for computational grid of 2° in the CS-1, CS-2, CS-3, CS-4 and CS5 respectively. For the computational grid of 0.25°, the highest estimated wave heights are 12.31 m, 10.00 m, 29.83 m, 23.84 m and 26.84 m in the CS-1, CS-2, CS-3, CS-4 and CS5 respectively. The highest wave heights, when comparing to National Geophysical Data Center (NGDC) maximum heights (NGDC, 2011) presented in Fig. 8 are reasonable due to the use of inversion source models. The inversion source models include tsunamigenic slips larger than 30 m in the source fault to a shallow part between the first rupture area and the Japan Trench (Koketsu et al., 2011; Fujii et al., 2011; Pollitz et al., 2011; Romano et al., 2012). However, it should be noted that there are considerable discrepancies in rupture models constructed through tsunami inversion techniques, seismic wave inversion techniques and elastic deformation studies embedded in homogeneous elastic half-spaces. The elastic deformation theory is used where the earth is assumed to follow the laws of the classical linear elastic theory which treats it as a homogeneous, isotropic, and elastic material. On the other hand, in reality, finite fault models based on inversion algorithms cannot be determined immediately after the earthquake. Hence, finite fault models cannot be utilized for warning of tsunamis in real time. For this purpose, the study concentrates on availability of the size of fault area and the average slip of Tohoku–Oki earthquake while the epicenter location and the magnitude can be measured relatively accurately and quickly after the earthquake occurs.

The simulations of the tsunami along the coastlines of Japan and offshore were compared with observation data. The observation data of the 2011 tsunami was recorded instrumentally at various gauges (Hayashi et al., 2011; Fujii et al., 2011) including coastal wave gauges (Nagai, 1998; Nagai et al., 2005), real-time kinematic...
global positioning system (RTK-GPS) buoys (Kato et al., 2005), cabled deep ocean-bottom pressure gauges (OBPG) (e.g. Fujisawa et al., 1986; Hirata et al., 2002), and Deep-ocean Assessment and Reporting of Tsunami (DART) buoys (González et al., 2005). Here, four deep ocean pressure sensors named DART were adopted for comparing the model results to DART buoy measurements named 21401, 21413, 21418, 21419 located throught the Pacific. Then, four GPS buoys named GPS-Fukushima, GPS-Miyagi (Center), GPS-Miyagi (North), GPS-Iwate (South), GPS-Iwate (Center), GPS-Iwate (North), and two OBPGs named TM1 and TM2 were adopted. RTK-GPS buoys are the nearest gauges used in this study for comparisons between observed and predicted waveforms in finer grid.

Fig. 5 shows the distribution of DART, RTK-GPS buoys and OBPGs in the computational area A and B. Comparisons between the computed tsunami waveforms and observed waveforms were evaluated for computation area A and computation area B for the different CS models given in Figs. 9–14. Tsunami Analysis Tool (TAT) developed by Annunziato (2007) were used to visualize and compare tsunami propagation, tsunami travel time and maximum heights with the records of wave gauges. TAT allows a comparison of the calculated value with the available sea level measurements downloaded from Intergovernmental Oceanographic Commission (IOC) and NOAA (National Oceanic and Atmospheric Administration) web sources (e.g. Annunziato et al., 2009; Ulutas et al., 2012).

By comparing the waveforms for the different uniform CS models in RTK-GPS buoys and OBPGs (Figs. 9 and 11), CS-1 and CS-2 provides largest amplitudes and earlier arrival times compared to the records of Fukushima. Although the time series of the waves are earlier for the rest of the comparison results, the amplitudes of the leading waves are in good agreement with the records of Miyagi (North). However, the leading waves were underestimated in Iwate (South), Iwate (Center) and Iwate (North). The earlier time arrivals and underestimations of the leading waves were simulated in TM-1 and TM-2 (Fig. 11). It is deduced that the characteristics of the simulated tsunami waveforms from the CS uniform models do

![Fig. 7. Tsunami maximum heights with grid size of 0.25 min in computation area B.](image-url)
not match those of the observed waveforms calculated in near field. The uniform slips of 13.08 m and 12.46 m for CS-1 and CS-2 respectively over the entire fault area indicate uplift above the locations of RTK-GPS buoys and OBPGs. The simulated values from CS-1 and CS-2 models also demonstrate sudden increases in sea height from 0 to about 2 m immediately after the earthquake. However, these sudden increases were not measured in the earlier part of the records. The lots of GPS buoys and OBPGs tsunami records show a drop in sea heights in the earlier part of the records immediately after the earthquake. Then, the waveform traces show increasing due to the tsunami. This could only be clarified with a subsidence along the coast line with a small slip. Furthermore, the tsunami waveform traces at TM-1 and TM-2 indicate a gradual uplift and then a sudden uplift which are a unique type of waveforms that has not been observed in the OBPG records for the other earthquakes (e.g., Maeda et al., 2011). It could be realized that it is impossible to explain the shape of these types of waveforms by using uniform slip models along the entire faults. Thus, the tsunami amplitudes was simulated by using above mentioned non-uniform models and compared with the observation data. As shown in Figs. 10 and 12, the CS-3 provides small amplitudes compared to the Fukushima, Iwate (South), Iwate (Center), Iwate (North), TM-1 and TM-2. It provides relatively good agreement with Miyagi (North) and Miyagi (Center) where the direction of the largest slips cumulated. However, for the rest of the gauges, CS-3 scenario provides underestimated waveforms. The underestimated waveforms might be resulted from much larger and much wider fault plane (550 km by 260 km) assumption with a lower slip distributions in lots of subfaults compared to the CS-4 and CS-5 models. Examining the waveforms shown in Figs. 10 and 12,
the CS-4 seems to agree better with the observed waveforms than that of CS-3 and CS-5. The calculated vertical dislocation for the CS-4 is 12.55 m which is higher than that of CS-3. Additionally, the CS-4 differs from the CS-1, CS-2 and CS-3 with a larger assumption of subsidence area near to coastal areas of Japan. The measured waveforms show decreasing amplitudes immediately after the earthquake. From these waveforms, it might be concluded that very little sea bottom uplift and subsidence occurred near to coast along the Honshu as proposed in Maeda et al. (2011). The simulated waveforms from CS-4 model match with measurements very well.
compared to the other source models used in this study. The simulated waveforms calculated from CS-5 were compared to RTK-GPS buoys and OBPGs. The waveform estimations from CS-5 match well with the measurements expect for the earlier part of the measurements. This difference might be from the assumption large subsidence in CS-5 model. The maximum subsidence and maximum uplift was estimated as 5.04 m and 15.84 m respectively in the CS-5 model. These values are higher than the other values mentioned above. Thus the CS-5 model engenders higher leading wave values than the others.

Although the maximum simulated waves reasonably match with the highest values of the observations especially in Iwate (North), the arrival times do not coincide with the observations. Furthermore the waveform and maximum height from CS-5 model of the first trailing wave do not reasonably coincide with the observations. This might be from estimated second large cumulated slip distributions with maximum slip exceeding 6 m expanding to the northeastern part of fault plane near trench off Iwate Prefecture (See Fig. 4). The second cumulated slip apart from the slip cumulated around hypocentral area was estimated only in the CS-5 model.

Finally, the simulated waveforms were also compared with the DART gauges (Figs. 13 and 14). By performing the uniform models, the arrival times of the estimated waves match well with the DART 21413. The location of DART 21413 is perpendicular to the strike of the fault. The arrival times of the estimated waves also show a good agreement with the DART 21418. However, the arrival times and the heights of leading crests do not seem to agree with the 21401 and 21419. The differences are about 5 min for the arrival times and about 0.30 cm for the heights of leading crests. These discrepancies were thought to be within reasonable limits due to the assumption of simple uniform fault and the distances of buoys from the source area. By performing the non-uniform models, the amplitudes of the leading crests are rather similar to the CS-3 and CS-5 models on DART 21413. However, the amplitudes were
overestimated for CS-4 model. All the arrival times of the models in DART 21413 are 1 min earlier than the measured arrival times. In DART 21418 which is the nearest DART to the source area, the CS-4 and CS-5 models provide reasonable match in terms of wave height. The CS-3 model tends to underestimate the wave height. The wave heights from CS-3 and CS-5 are 1 and 3 min earlier than the observed one respectively. In DART 21401 and DART 21419, the CS-3 tends to underestimate the wave height in addition to the earlier arrival of wave times. The CS-4 seems to agree better than that of other models in terms of wave heights and arrival times. The CS-5 model provides too early increase and surprisingly second trailing waves coincide with the measured waves. This might be clarified from second large cumulated slip distributions estimated in the CS-5 model near trench.

5. Summary, discussion and concluding remarks

In this study, a range of uniform and non-uniform slip models was used to perform tsunami simulations of the earthquake in an effort to examine and compare of different source and rupture models on maximum tsunami heights. The study was made to judge the set of source models obtained by different researchers and organizations. The simulations were performed with the approximation of shallow water equations adapted to an initial displacement of the ocean bottom deformation due to faulting. By employing the different uniform and non-uniform slip models, the maximum initial vertical displacements of the ocean bottom are 5.35, 5.16, 8.76, 12.55, 15.84 m for the CS-1, CS-2, CS-3, CS-4 and CS-5 source models respectively. The CS-1 and CS-2 models are uniform and CS-3, CS-4 and CS-5 are non-uniform models proposed by different researchers. The characteristics of the simulated waveforms from the uniform CS models do not match those of observed waveforms as previously mentioned. The reason is that the 500 km length and 200 km fault area anticipated from the after-shock distribution is not the effective fault area where the non-uniform largest slips cumulate. Consequently, determination of fault width and length relying on aftershocks only lead to over estimate the true rupture area. Furthermore, the assigned largest amount of slip to the shallower subfaults at the trench site and much lower amount of slip for the rest of the fault area yielded from non-uniform models are responsible for the propagation and orientation of the waves along the Honshu Coasts.

The simulations of the tsunami along the near shore and offshore were compared with observation data recorded at various gauges. Generally, the heights of the leading waves and arrival times were underestimated in uniform source models of CS-1 and CS-2. It is concluded that average slip values due to the predicted length and width of the fault plane by using only the after-shock area do not qualify the correct tsunami predictions in comparison to non-uniform slip distributions. Overall results obtained from the non-uniform models are in reasonably good agreement with the observations compared to those of uniform models. The results obtained for the uniform models showed that the CS-4 model is in better agreement with the observations. It should be noted that CS-4 model is a tsunami inversion source model. Thus, it is expected to be in better agreement with the observations compared to the uniform models and teleseismic inversion models. In this sense, the results of CS-3 model are much more important to show the effects of sea floor displacements caused by the seismic faulting on tsunami simulations. The CS-3 model provides relatively good estimation of tsunami heights especially in locations facing the direction of the largest slips cumulated. For the rest of the simulation area, the underestimated heights might be resulted from much larger and much wider fault plane assumption with lower slip distributions in lots of subfaults compared to the CS-4 and CS-5 models. Overall, the comparisons of the simulated waves with the observed waves performed from the different CS models show much better agreement in far field except for the CS-5 model. The early increase and second higher simulated trailing waves from the CS-5 model show that it could not be a second cumulated slip distributions proposed at the northeastern part of the fault area. The time shifts in the simulated waves from all models might be related to wave dispersion effects in the far field.

The results of the simulated comparison analysis which is seen to agree better with tsunami observations in especially CS-3 and CS-4 models prove the cumulated largest slips to the east of the epicenter by rising and subsiding the ocean floor near the coasts of Honshu. This agreement also indicates the requirements of non-uniform rupture models to accurately estimate the near field tsunamis. Unfortunately, it is still a challenge to determine the finite fault models immediately after occurrence of the earthquake providing multiple fault segments of variable depth, slip, rake angle, rupture time and seismic moment. In this study, through carrying out source models with assuming the uniform slip, it was confirmed that the simulated tsunami heights were underestimated. However, it might be also concluded from the uniform source model results that the advantage of the uniform model approaches is the simplicity and speed of use due to the limited knowledge of the earthquake parameters.

Finally, it was believed that the results of the study may contribute to remark the differences between the uniform and non-uniform source models used in tsunami simulations for the risk mitigation of future earthquakes.

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