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Are inundation limit and maximum extent of sand useful for differentiating tsunamis and storms? An example from sediment transport simulations on the Sendai Plain, Japan

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Abstract

We examined the quantitative difference in the distribution of tsunami and storm deposits based on numerical simulations of inundation and sediment transport due to tsunami and storm events on the Sendai Plain, Japan. The calculated distance from the shoreline inundated by the 2011 Tohoku-oki tsunami was smaller than that inundated by storm surges from hypothetical typhoon events. Previous studies have assumed that deposits observed farther inland than the possible inundation limit of storm waves and storm surge were tsunami deposits. However, confirming only the extent of inundation is insufficient to distinguish tsunami and storm deposits, because the inundation limit of storm surges may be farther inland than that of tsunamis in the case of gently sloping coastal topography such as on the Sendai Plain. In other locations, where coastal topography is steep, the maximum inland inundation extent of storm surges may be only several hundred meters, so marine-sourced deposits that are distributed several km inland can be identified as tsunami deposits by default. Over both gentle and steep slopes, another difference between tsunami and storm deposits is the total volume deposited, as flow speed over land during a tsunami is faster than during a storm surge. Therefore, the total deposit volume could also be a useful proxy to differentiate tsunami and storm deposits.

Keywords: tsunami deposit, storm deposit, numerical simulation, Delft-3D, SWAN

1. Introduction

Many studies have tried to identify the origin of deposits formed by tsunamis and storms by investigating their sedimentological characteristics (e.g., Nanayama et al., 2000; Tuttle et al., 2004; Kortekaas and Dawson, 2007; Morton et al., 2007; Komatsubara, 2012; Phantuwongraj and Choowong, 2012). However, the sedimentological characteristics of both types of deposits can be very similar (e.g., Goff et al., 2004; Phantuwongraj and Choowong, 2012; Kain et al., 2014; Goto et al., 2015), so that differentiating them is complex (e.g., Goff et al., 2012). Morton et al. (2007) proposed identification criteria for tsunami vs. storm deposits by reviewing the difference in their sedimentological characteristics and the hydraulic difference between tsunami and storm waves/surge. However, Watanabe et al. (2017) found that quantitative differentiation between both types of deposits is sometimes not possible based on field surveys only. Inclusion of other tools such as geochemical analysis and/or numerical simulations is needed for a better quantitative differentiation.

Recently, some studies have tried to investigate the distribution of marine-sourced deposits based on geochemical analyses. Chagué-Goff et al. (2012) identified mud deposits formed by the 2011 Tohoku-oki tsunami that were distributed inland up to 95% of the inundation limit, based on their geochemical characteristics, and this two months after the tsunami. Shinozaki et al. (2015a) collected and analyzed samples from the 2011 Tohoku-oki tsunami's deposits in Odaka, in northeast Japan. They suggested that biomarkers can be used as proxies to identify marine-sourced deposits on coastal land. In this way, geochemical analyses can be used to reveal the distribution of marine-sourced deposits from past inundation events such as tsunamis and storms. However, geo and biochemical analyses alone may not be used to distinguish between tsunami and storm deposits.

Numerical simulation is also an important method for the quantitative examination of the distribution of both types of deposits. Apotsos et al. (2011a) revealed that the initial distribution of sand is critical for determining the distribution of sandy deposits formed by tsunamis based on sediment transport calculations of the 2009 Samoa tsunami. Cheng and Weiss (2013) conducted numerical experiments in order to reveal a relationship between the inundation limit and the distribution limit of sandy deposits formed by tsunamis. They suggested that the deposition ratio (ratio of the distribution limit of sandy deposits to the tsunami inundation limit) was determined by the amplitude of the tsunami and topography, while grain size was not important. Watanabe et al. (2017) examined parameters which determine the distribution limit of storm deposits based on inundation and sediment transport simulations of the 2013 Typhoon Haiyan. Their simulated maximum inundation distance was 3.1 km inland from the coastline, while sediments were distributed only 0.2 km inland. They also revealed that roughness on land (primarily due to vegetation) and typhoon size are important for determining the distribution limit of storm deposits.

Other studies that also quantitatively examined the distribution of deposits based on numerical simulations are those by Apotsos et al. (2011a) and Sugawara et al. (2014). However, no previous numerical modeling studies have directly compared distributions of tsunami and storm deposits in the same area. Moreover, to better understand the processes governing sedimentation of sandy deposits and to differentiate between the two types of deposits, bedload and suspended load transport during each type of event should be investigated in detail (e.g., Watanabe et al., 2017).

We conducted numerical simulations of inundation and sediment transport due to tsunamis and storms, washing over identical topography, in order to reveal the quantitative difference in sediment transport processes and distribution of deposits during these events. We also determined which physical factors are important for differentiation between the two types of deposits.

2. Study area

We selected the Sendai Plain, Japan as our study area (Fig. 1, 2), because of the presence of sandy deposits formed by the 2011 Tohoku-oki tsunami (e.g., Goto et al., 2011; Abe et al., 2012) and detailed measurements of tsunami height (e.g., Mori et al., 2011), which can be used for validation of our numerical model. The Sendai Plain is located along Sendai Bay. The typical elevation of the plain is approximately 0–3 m relative to the vertical datum of Tokyo Peil (TP) and it extends approximately 4–5 km inland (Sugawara et al., 2014). Concrete-armored coastal dikes, with crest height 6.2 m above TP, were constructed well before the 2011 Tohoku-oki tsunami to protect the hinterland from high tides and storm surges. A coastal forest extended 100 to 700 m inland from the shoreline before the tsunami and was bordered inland by the Teizan Canal, which is approximately 20–30 m wide and ~2 m deep (Fig. 2).

On the Sendai Plain, the maximum inland inundation distance of the 2011 Tohoku-oki tsunami was 5.4 km from the coastline (Goto et al., 2012a), and the maximum flow depth was 9.6 m (Mori et al., 2011). There, the inverse model of Jaffe et al. (2012) showed that the current tsunami velocity ranged from 2.2 to 9.0 m s^{-1} based on data collected in trenches located from about 250 to 1350 m inland from the shoreline. Hayashi and Koshimura (2012) measured the flow speed based on the analysis of aerial video ~5 km south of our study area. They estimated that the current velocity of the tsunami front was 7 m s^{-1} at a location 1 km inland from the shoreline. Sugawara et al. (2014) conducted numerical simulations of inundation and sediment transport of the tsunami on the Sendai Plain. The calculated tsunami current velocity was generally less than 10 m s^{-1}, decreasing in an inland direction. Calculated major sources of sand deposited by the tsunami were the sea bed, the beach, and sand dunes which had been covered by coastal forest (Sugawara et al., 2014), consistent with field observations (e.g., Szczuciński et al., 2012). Sugawara et al. (2014) also noted, based on the calculation of sediment transport during the tsunami, that engineered structures such as coastal dikes heavily affected the transport of suspended sediments.

Many researchers conducted tsunami deposit surveys on the Sendai Plain after the 2011 Tohoku-oki tsunami (e.g., Goto et al., 2011, 2012b; Abe et al., 2012; Chagué-Goff et al., 2012; Richmond et al., 2012; Szczuciński et al., 2012; Shinozaki et al., 2015a). Among them, Abe et al. (2012) revealed that sandy deposits of thickness greater than 5 mm were distributed inland up to 57-76% of the inundation limit where it was more than 2.5 km from the shoreline, and all the way up to the inundation limit where it was less than 2.5 km. In our study area (Fig. 1, 2), the distribution limit of sandy deposits was 2.3 km from the shoreline on Transect A (which was adopted from Goto et al., 2012a) and 3.0 km on Transect B (which was adopted from Abe et al., 2012). Therein, transect A was offset near a pond (Fig. 2).

3. Methods

3.1 Numerical model used for this study

We used the Delft-3D and SWAN models (Deltares, 2011) for simulation of tsunami and storm hydrodynamics and sediment transport because such applications of these models have already been extensively validated (e.g., Apotsos et al., 2011a, 2011b; Bricker and Nakayama, 2014; Bricker et al., 2014; Watanabe et al., 2017). For tsunami simulation, we used Delft-3D alone, which implements the shallow water equations when applied with 1 vertical layer. We note that the shallow water equations cannot resolve tsunami soliton fission, which occurs over shallow seas such as the Sendai Bay (Murashima et al., 2012). Soliton fission is a process in which a long wave divides into several short waves due to non-linearity and dispersion effects (Japan Electric Power Civil Engineering Association, 2017). Fukazawa et al. (2002) revealed that the inundation extent of tsunamis can be simulated well without considering soliton fission, as it has little effect on the overall transport of tsunami mass or momentum, and thus it is not an essential aspect to resolve when investigating the inland extent of sediment transport.

For simulation of storm waves and surge, we used both Delft-3D and SWAN. Delft-3D calculates current fields, then passes the numerical result to SWAN (Deltares, 2011). SWAN calculates spectral parameters of the wave field via the conservation of wave action. Then, SWAN passes the numerical result for radiation stresses back to Delft-3D, so that mean current fields induced by wave setup are calculated. In this way, by running Delft-3D and SWAN together, wave fields and current fields induced by waves were calculated. We note that SWAN is a phase-averaged spectral wave model that cannot resolve low-frequency infragravity wave motion such as surf beat. However, this effect should be small on the Sendai Plain because it is located inside the broad and shallow Sendai Bay, the bathymetry of which would dissipate infragravity motions (e.g., Roeber and Bricker, 2015). Sediment transport was calculated with Delft-3D (Deltares, 2011) as in Apotsos et al. (2011a, 2011b), while bedload and suspended load were calculated using the formulation proposed by Van Rijn (1993).

Watanabe et al. (2017) conducted their sediment transport simulation with one vertical layer and did not account for the density stratification adopted by Apotsos et al. (2011a, 2011b). Nonetheless, their simulation could reproduce the measured maximum extent of sand and the distribution of sandy deposits. This result reinforces the assertion that one vertical layer is enough to simulate sediment transport due to extreme waves. In this study, we run the sediment transport model with one vertical layer as in Watanabe et al. (2017) in order to reduce computational load.

Model resolution was 3645 m, 1215 m, 405 m, 135 m, 45 m, and 15 m in domains 1, 2, 3, 4, 5, and 6, respectively (Fig. 1). Topographic data were generated from the pre-2011 earthquake DEM data used in Sugawara et al. (2014). To reduce computational load, sediment transport was calculated only in domain 6.

Roughness coefficients were determined based on the landuse map in Sugawara et al. (2014). This also affected the sediment transport calculation because flow speed was reduced at sites with high roughness coefficients.

Sugawara et al. (2014) revealed that sources of tsunami deposits were the seabed, beach, and sand dunes. The grain size in sand dunes and on the seabed in 2~10 m of water depth was 1.5-2.4 phi and 1.2-2.4 phi, respectively (Matsumoto, 1985). Therefore, we used a grain size of 0.267 mm (=1.9 phi) for the sediment transport simulation. We determined the initial distribution of sand via a landuse map together with data from Sugawara et al. (2014). Following Watanabe et al. (2017), we assumed the initial sediment layer thickness was 5 m because a finite initial sediment layer thickness helps avoid model instability due to excessive erosion and sedimentation. Hereafter, we define "distribution limit" as the distance from the shoreline up to which sand sheets with thickness of more than 5 mm were deposited (Abe et al., 2012, 2015; Sugawara et al., 2014; Watanabe et al., 2017).

3.2 Tsunami and storm model boundary conditions

The tsunami simulation was run over 5 hrs to include the effects of both incidence and backwash.

The composite fault model proposed by Imamura et al. (2012) was used as the wave source, in order to reproduce the general extent of the observed inundation area, so that the calculated flow depth was consistent with measured values as described in Section 4.1. The model is composed of 10 fault segments which are 100 km long and wide, and are arranged along the Japan Trench in two rows (Fig. 3a). The crustal deformation of the seafloor was calculated based on the elastic model proposed by Odaka (1985), with the initial tsunami waveform assumed to be identical to it.

For the simulation of storm waves and surge, 24 hrs of storm were simulated to capture the period during which a modeled typhoon moved from 300 km offshore until after landfall on the Sendai Plain. Characteristics of the typhoon used in this calculation were generated by using the method of Bricker et al. (2014). The path and central pressure of the typhoon were input into the parametric hurricane model of Holland (1980) for estimation of the air-pressure and wind fields. To account for asymmetry of these fields due to forward motion of the typhoon, the method was modified as in Fuji and Mitsuta (1986). The radius to maximum winds was estimated by using the empirical relation of Quiring et al. (2011). The track of the typhoon (Table 1) was also assumed to be that which maximizes the storm surge on the Sendai Plain as shown in Fig. 3b. The propagation speed of the typhoon was set to be the same as the 2013 Typhoon Haiyan (Japan Meteorological Agency, 2017).

To determine the required strength of the modeled typhoon, we review historical typhoons passing nearby Japan. The recorded strongest typhoon in the world (which also affected Japan) is the 1979 Typhoon Tip, with a pressure of 870 hPa and 10 min sustained winds of 140 knots (Kitamoto, 2017). The 1934 Muroto typhoon, the strongest historic typhoon to have made landfall in Japan, had a minimum central pressure of 911.6 hPa (Japan Meteorological Agency, 2017). During the 1961 Typhoon Nancy, pressure was 920-925 hPa near 38 N degrees (Kitamoto, 2017), which is the same latitude as the Sendai Plain.

Based on these past typhoons, we assumed a typhoon with 140 kt maximum wind speed and 870 hPa central pressure in our calculation. As noted above, a typhoon of this strength has never been recorded on the Sendai Plain (or anywhere at 38 N degrees) and thus the intensity of this typhoon is unrealistically strong. However, our main objective is to prove that even an unrealistically strong storm surge is not energetic enough to transport sediment far inland, as discussed below. Thus, the assumed typhoon is suitable for our study.

3.3 Validation of numerical model

To validate our numerical model, we conducted a numerical simulation of inundation and sediment transport during the 2011 Tohoku-oki tsunami (Fig. 3a). We examined the accuracy of our model by comparing our numerical results with measured water levels (Mori et al., 2011) as shown in Fig. 4. We also examined the accuracy of the sediment transport simulation by comparing calculated results with measured sand thickness (Abe et al., 2012; Goto et al., 2012a) on Transect A (Fig. 5) and Transect B (Fig. 6). Validation of the roughness distribution has already been conducted by Sugawara et al. (2014), and we also checked that our simulation accurately reproduced the measured flow depth by using this roughness distribution as described in Section 4.1. We used a topography with pre-2011 coastal dikes included, because we used measured water depth and tsunami deposit data from the 2011 Tohoku-oki tsunami for the validation.

3.4 Inundation and sediment transport during a storm vs. a tsunami

After our numerical model, we conducted inundation and sediment transport calculations during a hypothetical tsunami and storm using the "natural" topography from which Sendai's coastal dike and the Sendai Tobu road (Fig. 2) were removed. This is in order to investigate the processes of inundation and sediment transport (and formation of deposits) during storm vs. tsunami events and distribution of deposits under natural environmental conditions.

In general, both marine-sourced deposits and deposits originating from sand dunes could come from either storms (e.g., Kortekaas and Dawson, 2007) or tsunamis (e.g., Szczuciński et al., 2012). Thus, hereinafter, a movable sediment bed was assumed to exist everywhere from the sea floor to the sand dunes, and the initial sediment layer thickness was 5 m as in Watanabe et al. (2017).

4. Results

4.1 Validity of the tsunami simulation results

Results of the numerical tsunami simulation were verified by comparing measured (Mori et al., 2011) and computed flow depths at 39 points using the parameters K and κ (Aida, 1978). K and κ indicate geometric average value and fluctuation, respectively, in the ratio of observed to computed amplitudes. In our case, K=0.95 and κ=1.21 (Fig. 4). Takeuchi et al. (2005) suggested that $0.8 \leqq K \leqq 1.2$ and $\kappa \leqq 1.6$ are required to accurately reproduce measured values. Our results showed that both κ and K were within these required ranges. Calculated flow depths were less than the measured values at sites where large flow depths (> 8 m) were recorded (Fig. 4). All these sites were located near the shoreline, with measurements in or behind the coastal forest. When a tsunami strikes a coastal forest, trees can fall due to hydraulic force; therefore the effect of the coastal forest on reducing the energy of the tsunami might weaken as inundation continues. However, this effect is not included in our simulation, and this may explain why simulated flow depths at these sites were not consistent with measured values. Nevertheless, flow depths at the other points were consistent with the measured values, and this qualitative agreement is considered a practical indicator of model validity for tsunami inundation modelling.

We then verified the reproducibility of the sediment transport calculation. Numerical results indicate that sandy deposits extended up to 3.3 km inland from the coast along Transects A and B (Fig. 5, 6), while the measured inland extents of sand on Transects A and B were 2.3 km and 3.0 km, respectively. The calculated extent of sand slightly overestimated the measured values on both transects.

The modeled volume of deposits along transect B was 3.02×10^2 m^2, and the volume of deposits up to the measured distribution limit (defined as sand sheets greater than 5 mm thick) was 2.92×10^2 m^2. Thus, 97% of sand on the transect was deposited up to the measured distribution limit. On Transect A, the modeled volume of deposits along the transect was 3.14×10^2 m^2, and 2.98×10^2 m^2 up to the measured distribution limit; thus 95% of sand was deposited up to the measured distribution limit.

We also compared the observed and calculated volumes of sandy deposits 0~1 km, 1~2 km, 2~3 km, and 3~4 km inland from the shoreline, respectively (Table 2). On both transects, modeled sand volumes overestimated the measured volumes at 0~1 km and 2~3 km (Table 2). A reason for this discrepancy may be that density stratification for sediment transport or a suspended load concentration limit was not included in our sediment transport modelling. Thus, the modeled suspended load was high, and the calculated sand volumes overestimated the measured volumes.

The other reason for this discrepancy at 0~1 km is probably because the sediment-trapping effect of the coastal forest is not included in our simulation. For inundation modelling, this effect was included by using a spatially variable roughness coefficient, so that flow speed was reduced in the coastal forest zone due to the high roughness coefficient. This reduced flow speed and reduced sediment transport there as well. However, in reality, sediments within a coastal forest become even more difficult to erode and transport because of trapping and shielding by vegetation. This shielding effect was not included in the sediment transport simulation, so it is possible that too much sediment was eroded within the coastal forest zone, and that the calculated volumes of sandy deposits at 0~1 km might be overestimated. At 2~3 km inland, the calculated current velocity of tsunami is ~4 m s^{-1}, so that eroded sediments started to be deposited. However, because the effects of sediment trapping and density stratification were not explicitly included in the simulation, modeled deposition overestimates measured deposition at this site.

The calculated tsunami deposit sand thickness is generally consistent with measured values (Fig. 5, 6). However, some discrepancies exist. On transect B, the measured thickness 600 m inland from the shoreline was 11 cm, while the calculated thickness at this site was 47 cm. From 2200~3000 m inland, the calculated sand thickness was overestimated compared to measured values by more than 10 cm. On transect A, the measured thickness was several cm from 440~600 m inland, but the calculated value was 24~39 cm. Furthermore, from 2100~2400 m inland, the calculated deposit thickness was overestimated by more than 10 cm. These inconsistencies are partially due to the coarseness of the model grid resolution. The calculated sediment thickness at each grid cell is an averaged thickness of the 15 m × 15 m area. Thus, as reported by previous studies (e.g., Sugawara et al., 2014), the comparisons between simulated and observed thickness on a point by point basis is difficult, even though overall trends agree well.

Total calculated sand volumes did not show quantitative agreement with measured volumes (Table 2), but calculated deposition of sand ceased near the measured distribution limit of deposits, and the trend of measured deposit thickness was reproduced well in our simulation. For tsunami sediment transport modeling, such qualitative agreement is considered a practical indicator of model validity (Sugawara et al., 2014).

4.2 Inundation and sediment transport during a tsunami

In Section 4.1, we used the topography with coastal structures to validate the model. However, as described in section 3.4, in order to compare inundation and sediment transport during a hypothetical storm with tsunami under the natural environment, we use the topography without these artificial structures hereafter.

During the tsunami, the maximum water level near the shoreline in the finest computational domain was 13.1 m (Fig. 7a). The inundation limits along Transects A and B were 4.4 km and 4.0 km from the shoreline, respectively. The maximum flow speed during the incident wave (Fig. 8a) was 12.6 m s^{-1} at a point 0.16 km inland from the shoreline.

The simulated distribution limits of deposits on Transects A and B were 3.3 km and 3.9 km, respectively (Fig. 9a). The calculated maximum thickness of sandy deposits was 31 cm at a point 0.53 km inland from the shoreline on Transect A, and 29 cm at a point 0.60 km inland on Transect B. In the finest computational domain, the total volume of sand deposited over land was 1.1×10^6 m^3, while the total volume eroded from the sea floor to the sand dunes was 1.8×10^6 m^3. The calculated volume deposited is not equal to the volume eroded because the rest of the sediments are transported and deposited offshore.

We also show the distribution of suspended transport (Fig. 10) and bedload transport (Fig. 11) when the tsunami reached the shoreline (Fig. 10a), when it ran overland (Fig. 10b), and when it reached its maximum inland inundation extent (Fig. 10c).

The simulated maximum horizontal suspended transport flux when the tsunami reached the shoreline (Fig. 10a) was 6.5 m^2 s^{-1} at a point 0.82 km inland as this was near the front of the tsunami. The maximum flux of suspended transport when the tsunami ran overland (Fig. 10b) was 0.42 m^2 s^{-1} at a point 2.5 km inland. When the tsunami reached its maximum inland inundation extent (Fig. 10c), the maximum suspended transport flux was 0.24 m^2 s^{-1} at a location 0.015 km inland from the shoreline. Here, suspended transport on land had almost ceased. As the tsunami propagated inland from the coastline, flow speed gradually decreased, and so did the flux of suspended sediment.

On the other hand, the maximum volume flux of bedload sediment was 5.1 m^2 s^{-1} at 0.17 km when the tsunami reached the shoreline (Fig. 11a). When the tsunami ran overland (Fig. 11b), bedload flux was 1.1×10^{-2} m^2 s^{-1} at a point 2.2 km inland from the shoreline. When the tsunami reached its maximum inland inundation extent (Fig. 11c), it was 5.1×10^{-3} m^2 s^{-1} at a point 0.015 km inland from the shoreline.

We also calculated the ratio of total bedload flux to total suspended load flux (bedload flux /

suspended load flux) in the final domain, and the results are shown in Table 3. The ratio of bedload to suspended load flux when the tsunami ran overland (75 min) is small. This is because the velocity on land is high enough to transport sediments in suspension at this time, so that eroded sediments were transported as suspended load over land instead of as bedload.

4.3 Inundation and sediment transport during storm surge

The maximum water level over land in the finest model domain was 6.1 m (Fig. 7b). The maximum extents of inundation on Transects A and B were 5.8 km and 7.2 km inland, respectively. Maximum storm surge flow speed and maximum near-bottom mean orbital velocity induced only by storm waves in this domain were 6.4 m s^{-1} (occurring near the estuary) and 2.1 m s^{-1} (occurring near the coastline), respectively (Fig. 8b, c). We note that the maximum near-bottom mean orbital velocity induced by storm waves is small over land, so bed shear stress due to storm waves was also small. Thus, sediment transport overland was not generated by storm waves.

The calculated thicknesses of sandy deposits on Transects A and B were 64 cm and 105 cm, respectively (Fig. 9b). The inland extents of these deposits along Transects A and B were 0.19 km and 0.27 km from the shoreline, respectively. In the finest model domain, the total volume of sand deposited over land was 1.2×10^5 m^3, while the total volume eroded was 1.4×10^6 m^3. The majority of the sand was deposited offshore because the maximum near-bottom mean orbital velocity of storm waves and the current velocity of storm surge over land were very small, so sediments were not transported over land. Significant sedimentation and erosion occurred near the estuary in the northeast part of the domain (Fig. 9b) because the elevation at this site is low.

We also differentiated the distribution of suspended load (Fig. 12) and bedload transport (Fig. 13) during the storm when surge inundated on the shoreline (Fig. 12a), when the surge propagated inland (Fig. 12b), and when the surge reached its maximum inland extent (Fig. 12c). The maximum volume flux of suspended transport when the surge reached the shoreline (Fig. 12a) was 1.9×10^{-2} m^2 s^{-1} in the river 0.53 km inland from the shoreline. When the surge propagated inland (Fig. 12b), the flux was 1.9×10^{-2} m^2 s^{-1} in the river 0.57 km inland from the shoreline. When it reached its maximum inland extent (Fig. 12c), the flux was 2.1×10^{-1} m^2 s^{-1} at the shoreline. The volume of suspended sediments transported over land during the storm was smaller than that transported by the tsunami. Regarding bedload transport, the maximum volume flux when the surge reached the shoreline (Fig. 13a), when it propagated inland (Fig. 13b), and when it reached its maximum inland extent (Fig. 13c) was 3.0×10^{-4} m^2 s^{-1} at a point 0.10 km inland from the shoreline, 3.8×10^{-4} m^2 s^{-1} at a point 0.21 km inland, and 2.0×10^{-3} m^2 s^{-1} at a point 0.021 km inland, respectively. Bedload sediment transport flux was also smaller than during the tsunami.

We also calculated the ratio of total bedload flux to total suspended load flux, and the results are shown in Table 4. The ratio when the surge reached its maximum inland extent (975 min) is small because a strong return flow occurred near the estuary, generating a large suspended load.

5. Discussion

5.1 Difference between inundation and sediment transport extents during tsunami vs. storm

The overland propagation speed of the tsunami was much greater than that of the storm surge. The tsunami only took about 20 min from the time of incidence on the coast to reach its maximum inland inundation extent, while the storm surge took almost 4 hrs. This is because a storm surge during a passing typhoon can last as long as several hrs (Ministry of Agriculture, Forestry and Fisheries, 2015), while tsunami wave period is several tens of min (e.g., Nagai et al., 2007).

In our simulation, the distance inundation extended from the coastline was larger for storm surge than for tsunami by several km (Table 5). This is because the elevation of the Sendai Plain is low (0~3 m TP). Moreover, the duration of the storm surge was longer than that of the tsunami, and lower storm surge flow speeds (Fig. 8b) corresponded to less dissipation of energy by bottom friction than in a tsunami. Furthermore, the typhoon used in the storm surge simulation was stronger than any event which has historically hit the region.

The ratio of bedload flux to suspended load flux is small in the case of a tsunami compared to a storm surge. This is because tsunami flow speed over land is high compared to storm surge flow speed, and sediments are transported overland as suspended load.

As described in Section 3.2, our assumed worst-case typhoon is unrealistically strong for the latitude of the Sendai Plain. Thus, we additionally conducted a simulation with a typhoon which can realistically make landfall there (see Appendix A) in order to investigate the storm surge inundation and sediment distribution by such a typhoon. The modeled inundation distances of storm surge and waves during this typhoon were 4.6 km and 4.2 km on Transect A and B, respectively, which are slightly longer than those of the modeled tsunami inundation distances on these transects (4.4 km and 4.0 km, respectively) (Table 5). Therefore, it should be emphasized that even a realistic typhoon-generated storm surge can inundate farther inland than a tsunami. This is because of the topographic setting of the Sendai Plain. The height of sand dunes located near the coastline is about 3 m, while the elevation along both transects up to 4~5 km inland from the coastline is <3 m. Thus, once storm surge overtops the sand dune, it can easily inundate 4~5 km inland.

Some studies (e.g., Srinivasalu et al., 2008; Chagué-Goff et al., 2011, 2015) tried to detect evidence for saltwater inundation of paleotsunamis based on chemical analysis. Other studies (e.g., Shinozaki et al., 2015a, 2015b) tried to identify paleotsunami sand deposits based on the analysis of diatoms or other marine-sourced biomarkers. These approaches are indeed important to estimate the tsunami inundation distance correctly. However, in our case, the storm surge inundated farther inland than the tsunami, so the maximum extent of sandy tsunami deposits is smaller than the inundation extent of the storm surge. Thus, identification of a deposit's origin cannot be conducted by using chemical methods alone, as these cannot distinguish whether the geochemical signature was a result of a tsunami or a storm surge.

The volume of tsunami-induced erosion in the smallest model domain was larger than the volume of storm-induced erosion. This is because tsunami-induced flow speeds near the coastline (the region of most erosion) are stronger than storm surge flow speeds (Fig. 8), resulting in larger bed shear stresses. Likewise, the volume of tsunami deposits was larger than that of storm deposits. This is because of the difference in flow speeds farther upland (where most of the deposition occurred). On both transects A and B, the overland flow speed of the tsunami was faster than that of the storm surge (Table 5), so that the former was able to keep sediments in motion until they were transported far inland (Fig. 9). Slow flow speeds under the storm surge, on the other hand, caused sediments to settle out and cease motion much closer to shore. This held true for both suspended and bedload transport (Fig. 12, 13).

5.2 Relationship of tsunami and storm deposits to topography

The Sendai Plain consists of mildly sloping land with elevation ranging from 0-3 m TP within 4-5 km of the coastline. Over such topography, tsunami velocity over land is large, causing the maximum extent of sand to be large compared to the case of storm surge. Inoue et al. (accepted for publication) conducted a field survey of tsunami deposits and a simulation of the maximum likely storm surge in Noda village, Iwate Prefecture, Japan. At that site, the elevation 600 m inland from the coastline is 10 m TP; this topography is much steeper than on the Sendai Plain. In Noda Village, a gravel layer deposited by the possible AD 869 Jogan tsunami was distributed up to 700 m inland from the coastline (elevation: 11 m TP), while the calculated inundation extent of maximum credible storm surge is only 450 m from the shoreline (elevation: 7.33 m TP). In contrast to the Sendai Plain, tsunamis inundate farther inland than the maximum credible extent of storm surge.

Over any topography, the maximum inland extent of storm deposits is small, not greater than several hundred m from the shoreline. However, there are some observations where thin (mm-thick) deposits were locally formed in low-lying depressions far inland (Watanabe et al., 2017). Pilarczyk et al.

(2016) also reported that isolated sandy storm deposits on Leyte Island (Philippines) formed during Typhoon Haiyan were found up to 1.7 km inland from the shoreline, where the inundation limit was 2.0 km. Therefore, the maximum extent of discontinuous sand deposits may also be affected by local topography.

5.3 Identification of tsunami vs. storm deposits

Previous studies identified the origin of deposits based on the assumption that the storm surge inundation limit is smaller than the tsunami inundation limit (e.g., Inoue et al., accepted for publication). They identified the sandy deposits distributed farther inland than the possible storm surge inundation limit (estimated by the simulation of storm waves and surge) as tsunami deposits. However, in the case of a gentle land slope such as on the Sendai Plain, only confirming the existence of inundation is insufficient for differentiation of tsunami vs. storm deposits, because the storm inundation limit may be larger than that of tsunami (as in our study, Fig. 14a, c). If the land slope is steep such as in Noda village, the computed inundation limit of the maximum credible storm is small (Inoue et al., accepted for publication). However, tsunami deposits can be formed up to several km inland from the shoreline over steep topography. Apotsos et al. (2011b) conducted a simulation of tsunami inundation and sediment transport on an ideal topography where the land slope extended 2000 m inland from the shoreline up an elevation of 20 m. On such steep topography, the tsunami inundated more than 1000 m inland from the shoreline and sandy deposits also formed more than 1000 m inland. These studies reinforce the hypothesis that sandy deposits distributed several km inland can be confidently identified as of tsunami origin (Fig. 14b, d).

Morton et al. (2007) also compared the inundation distance and maximum extent of sand due to a tsunami and a storm. They found that the inundation limit of storm surge is 10^2-10^4 m and the maximum extent of its sand deposit is less than several hundred m, while the tsunami inundation limit is 10^2-10^3 m and the maximum extent of its sand deposit is also 10^2-10^3 m. If land topography is gently sloping, the trends of inundation distance and maximum extent of sand due to the tsunami and the storm shown in Fig. 14 are similar to those of Morton et al. (2007). However, if the land slope is steep, the inundation distance of storm surge and waves will be small, in contrast to Morton et al. (2007), while the inundation distance of the tsunami and the maximum extent of sand deposits are 10^2-10^3 m, similar to Morton et al. (2007).

Some researchers reported that the distribution distance of storm deposits is smaller than that of tsunamis (e.g., Goff et al., 2004; Tuttle et al., 2004; Kortekaas and Dawson, 2007; Morton et al., 2007). This is because the storm surge flow speed (and therefore bed shear stress) several hundreds to thousands of m inland is smaller than that of tsunamis over either steeply or gently sloping land.

Thus, sand layers distributed up to several km inland may be identified as of tsunami origin. However, thin (mm-thick) storm deposits may be locally formed in low-lying depressions even several km inland from the shoreline (Fig. 14a) as mentioned in Watanabe et al. (2017). Soria et al. (2017) also suggested that the inland extent and thickness of storm deposits are governed by local variations in topography or vegetation, so that the identification of tsunami or storm deposits may not be conducted by assessment of the inland extent and thickness of storm deposits alone.

We also showed that the total volume of deposits may be much smaller in the case of storms than tsunamis, even when we assumed an unrealistically strong typhoon. Therefore the volume of deposits could be a useful proxy to differentiate tsunami vs. storm deposits for any topographic conditions, although the deposit volume formed by a small tsunami may also be small.

The thickness of sandy storm deposits also tends to be larger than that of tsunami deposits (e.g., Morton et al., 2007; Phantuwongraj and Choowong, 2012). In our simulation, storm deposits were thicker than tsunami deposits near the shoreline (Fig. 5, 6), because storms persist longer than tsunamis (Watanabe et al., 2017). Therefore, as Morton et al. (2007) mentioned, the formation of thick deposits near the shoreline may be characteristic of storms, in agreement with the modelling results of Watanabe et al. (2017). However, these sedimentological characteristics of storm deposits might also be affected by the wave state, local topography, or local vegetation.

For identification of tsunami deposits, confirming the inundation limit is important because the inundation limit of storms is small if the land slope is steep. In that case, sandy deposits distributed several km inland over a broad range can be confidently identified as of tsunami origin. If land topography has a gentle land slope, the storm inundation limit may be larger than that of tsunamis (Fig. 14a, c). Even in this topographic setting, if marine-sourced deposits are distributed inland several km over a broad range, we may be able to identify these deposits as of tsunami origin.

Although the volume of sediments, thickness and inundation limit could be useful indicators to differentiate storm vs. tsunami deposits, many factors such as sediment source, topography and wave state can complicate the identification of deposit source. Therefore in addition to field investigation, numerical modeling is also a useful method for the quantitative identification of tsunami deposits.

5.4 Future research on differentiation of tsunami vs. storm deposits

In areas such as coral reefs with very steep offshore slopes, energetic low-frequency wave motions such as surf beat can be generated by storm waves (Roeber and Bricker, 2015). In this case, suspended and bedload sediment transport due to storm waves and surge overland could be large, and the corresponding sand deposits could extend far inland.

As described in Section 5.4, thicker storm deposits than tsunami deposits were formed near the shoreline in our simulation (Fig. 5, 6). However, the deposit thickness near the shoreline may also be affected by topography or wave force. Moreover, the two processes of suspended and bedload transport may be affected by topography. Furthermore, sandy deposits formed by a small tsunami may have similar characteristics (e.g., limited inland extent of sand, small volume of deposits) to storm deposits. However, the sedimentological characteristics of sandy deposits formed by small tsunamis is not understood well. The above points must be investigated for better identification of tsunami and storm deposits. Moreover, which parameters determine the difference between inundation distance and maximum extent of sand during tsunami or storm also should be revealed in order to precisely identify a deposit’s origin.

6. Conclusions

We quantitatively examined the difference in the distribution of tsunami and storm deposits on the Sendai Plain based on sediment transport simulations during both types of events. Our numerical results indicated that the simulated inland inundation distance of a large hypothetical storm surge was greater than that of the 2011 Tohoku-oki tsunami by several km. Even a realistic typhoon can produce a storm surge inundation extent slightly greater than that of the 2011 event. Based on the assumption that storm surge inundation extent is generally smaller than that of tsunamis, previous studies identified sandy deposits which are distributed farther inland from the coastline as being of tsunami origin. However, over gently sloping topography such as the Sendai Plain, storm surge inundation can be more extensive than tsunami inundation, so that differentiation of their deposits is not possible by only confirming the inundation extent. Over steep topography such as in Noda village, the maximum inundation distance and inland extent of sand deposits due to storm surge become small, so sandy deposits distributed several km inland from the coastline can be identified as of tsunami origin. Under any topographic condition, the maximum inland extent of continuous storm sand deposits is small because the flow speed of storm surge and waves over land is small. Thus, sandy deposits which are continuously distributed several km inland can be identified as of tsunami origin. For the same reason, the total volume of storm deposits may be much smaller than the total volume of tsunami deposits over any topography, so the total deposit volume could also be a useful proxy to differentiate these deposits. However, discontinuous sandy storm deposits might be formed in low-lying depressions several km inland from the coastline, and the total volume of deposits formed by a small tsunami may also be small.

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Appendix A. Simulation of storm surge due to a typhoon which could realistically make landfall on the Sendai Plain

Our assumed worst case typhoon (Typhoon Tip) is unrealistically strong for the Sendai Plain. This choice of typhoon strength was made to put a limit on the inland extent of sand deposits that could occur in general for the case of a gently sloping topography like the Sendai Plain. However, to identify tsunami deposits on the Sendai Plain itself based on the calculated inundation limit of storm surge, this assumed typhoon is too strong. Thus, we also conducted simulation of a storm based on the 1961 Typhoon Nancy which affected Japan at 38 N degrees (the same latitude as the Sendai Plain). The track of the typhoon was assumed to be that which maximizes the storm surge on the Sendai Plain as shown in Fig. 3b. The propagation speed of the typhoon was assumed to be the same as the 2013 Typhoon Haiyan (Japan Meteorological Agency, 2017). During the 1961 Typhoon Nancy, the central low pressure was 920-925 hPa at 38 N degrees (Kitamoto, 2017) and its maximum wind speed at Sakata city (which is at the same latitude as the Sendai Plain) was 38 m s^{-1} (Yamamoto, 1963). Thus, the central pressure of the typhoon and the wind speed used for the simulation were 920 hPa and 38 m s^{-1}, respectively. The calculated distribution of maximum water level was as shown in Fig. A.1. The inundation distances along Transects A and B were 4.6 km and 4.2 km, respectively.

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Figure Captions

Table 1. Time, location, central pressure, and wind speed of the modeled typhoon used for our simulation.

Table 2. Measured and modeled volumes of sandy deposits along transects A and B.

Table 3. Maximum volume flux of suspended load, maximum volume flux of bedload, and ratio of total bedload to total suspended load in the fine domain during tsunami simulation.

Table 4. Maximum volume flux of suspended load, maximum volume flux of bedload, and ratio of total bedload to total suspended load in the fine domain during the storm surge simulation.

Table 5. Measured inundation distance from the coastline, maximum extent of sand, and maximum sand thickness on Transects A and B, and calculated inundation distance from the coastline, maximum velocity, maximum extent of sand, and maximum sand thickness on both transects.

Figure 1. Locations of the study area and (a) domains 1~4 and (b) domains 4~6.

Figure 2. Topography of domain 6 and locations of two transects set by Abe et al. (2012). Yellow points indicate locations of pits in the tsunami deposit surveys by Abe et al. (2012). Other features are labeled as follows: CD: coastal dike, TC: Teizan Canal, TR: Sendai Tobu Road, IL: Inundation limit of tsunami, SL: measured maximum extent of sand.

Figure 3. (a) Initial water level of the 2011 Tohoku-oki tsunami using the tsunami source model proposed by Imamura et al. (2012). (b) Distribution of wind speed of the category 5 typhoon used for simulation of storm waves and surge. The red line is the path of the typhoon.

Figure 4. Comparison of measured (Mori et al., 2011) and calculated flow depths.

Figure 5. (a) Comparison of measured thickness of sandy deposits formed by the 2011 Tohoku-oki tsunami (Abe et al., 2012) and calculated sand layer thickness on Transect A (see text for explanation). (b) Cross-sectional topography of Transect A.

Figure 6. (a) Comparison of measured thickness of sandy deposits formed by the 2011 Tohoku-oki tsunami (Abe et al., 2012) and calculated sand layer thickness on Transect B (see text for explanation). (b) Cross-sectional topography of Transect B.

Figure 7. Comparison of maximum water levels between (a) tsunami and (b) storm surge.

Figure 8. Comparison of (a) maximum current velocity of tsunami, (b) maximum current velocity of storm surge, and (c) maximum near-bottom mean orbital velocity of storm waves.

Figure 9. Comparison of calculated erosion and sedimentation between (a) tsunami and (b) storm.

Figure 10. Comparison of calculated water level at (a) 68 min, (b) 75 min, (c) 88 min after the start of the simulation and suspended transport due to the tsunami at (d) 68 min, (e) 75 min, (f) 88 min.

Figure 11. Comparison of calculated water level at (a) 68 min, (b) 75 min, (c) 88 min after the start of the simulation and bedload transport due to the tsunami at (d) 68 min, (e) 75 min, (f) 88 min.

Figure 12. Comparison of calculated water level at (a) 735 min, (b) 855 min, (c) 975 min after the start of the simulation and suspended transport due to the storm at (d) 735 min, (e) 855 min, (f) 975 min.

Figure 13. Comparison of calculated water level at (a) 735 min, (b) 855 min, (c) 975 min after the start of the simulation and bedload transport due to the storm at (d) 735 min, (e) 855 min, (f) 975 min.

Figure 14. Differences in inundation distances and sediment-transport distances for sand beds deposited by tsunamis and storms over two types of topographies. (a) Storms on flat topography, (b) storms on steep topography, (c) tsunamis on flat topography, and (d) tsunamis on steep topography are shown. This figure was modified after Morton et al. (2007).

Figure A.1. Maximum water level of storm surge induced by the typhoon which could realistically make landfall on the Sendai Plain.