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

6
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19

20

21 **Abstract**

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

37

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

39

40 **1. Introduction**

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

53

54 Recently, some studies have tried to investigate the distribution of marine-sourced deposits based on
55 geochemical analyses. Chagué-Goff et al. (2012) identified mud deposits formed by the 2011
56 Tohoku-oki tsunami that were distributed inland up to 95% of the inundation limit, based on their

57 geochemical characteristics, and this two months after the tsunami. Shinozaki et al. (2015a) collected
58 and analyzed samples from the 2011 Tohoku-oki tsunami's deposits in Odaka, in northeast Japan.
59 They suggested that biomarkers can be used as proxies to identify marine-sourced deposits on
60 coastal land. In this way, geochemical analyses can be used to reveal the distribution of
61 marine-sourced deposits from past inundation events such as tsunamis and storms. However, geo and
62 biochemical analyses alone may not be used to distinguish between tsunami and storm deposits.

63

64 Numerical simulation is also an important method for the quantitative examination of the
65 distribution of both types of deposits. Apotsos et al. (2011a) revealed that the initial distribution of
66 sand is critical for determining the distribution of sandy deposits formed by tsunamis based on
67 sediment transport calculations of the 2009 Samoa tsunami. Cheng and Weiss (2013) conducted
68 numerical experiments in order to reveal a relationship between the inundation limit and the
69 distribution limit of sandy deposits formed by tsunamis. They suggested that the deposition ratio
70 (ratio of the distribution limit of sandy deposits to the tsunami inundation limit) was determined by
71 the amplitude of the tsunami and topography, while grain size was not important. Watanabe et al.
72 (2017) examined parameters which determine the distribution limit of storm deposits based on
73 inundation and sediment transport simulations of the 2013 Typhoon Haiyan. Their simulated
74 maximum inundation distance was 3.1 km inland from the coastline, while sediments were

75 distributed only 0.2 km inland. They also revealed that roughness on land (primarily due to
76 vegetation) and typhoon size are important for determining the distribution limit of storm deposits.

77

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

84

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

89

90 **2. Study area**

91 We selected the Sendai Plain, Japan as our study area (Fig. 1, 2), because of the presence of sandy
92 deposits formed by the 2011 Tohoku-oki tsunami (e.g., Goto et al., 2011; Abe et al., 2012) and

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

101

102 On the Sendai Plain, the maximum inland inundation distance of the 2011 Tohoku-oki tsunami was
103 5.4 km from the coastline (Goto et al., 2012a), and the maximum flow depth was 9.6 m (Mori et al.,
104 2011). There, the inverse model of Jaffe et al. (2012) showed that the current tsunami velocity
105 ranged from 2.2 to 9.0 m s⁻¹ based on data collected in trenches located from about 250 to 1350 m
106 inland from the shoreline. Hayashi and Koshimura (2012) measured the flow speed based on the
107 analysis of aerial video ~5 km south of our study area. They estimated that the current velocity of the
108 tsunami front was 7 m s⁻¹ at a location 1 km inland from the shoreline. Sugawara et al. (2014)
109 conducted numerical simulations of inundation and sediment transport of the tsunami on the Sendai
110 Plain. The calculated tsunami current velocity was generally less than 10 m s⁻¹, decreasing in an

111 inland direction. Calculated major sources of sand deposited by the tsunami were the sea bed, the
112 beach, and sand dunes which had been covered by coastal forest (Sugawara et al., 2014), consistent
113 with field observations (e.g., Szczuciński et al., 2012). Sugawara et al. (2014) also noted, based on
114 the calculation of sediment transport during the tsunami, that engineered structures such as coastal
115 dikes heavily affected the transport of suspended sediments.

116

117 Many researchers conducted tsunami deposit surveys on the Sendai Plain after the 2011 Tohoku-oki
118 tsunami (e.g., Goto et al., 2011, 2012b; Abe et al., 2012; Chagué-Goff et al., 2012; Richmond et al.,
119 2012; Szczuciński et al., 2012; Shinozaki et al., 2015a). Among them, Abe et al. (2012) revealed that
120 sandy deposits of thickness greater than 5 mm were distributed inland up to 57-76% of the
121 inundation limit where it was more than 2.5 km from the shoreline, and all the way up to the
122 inundation limit where it was less than 2.5 km. In our study area (Fig. 1, 2), the distribution limit of
123 sandy deposits was 2.3 km from the shoreline on Transect A (which was adopted from Goto et al.,
124 2012a) and 3.0 km on Transect B (which was adopted from Abe et al., 2012). Therein, transect A
125 was offset near a pond (Fig. 2).

126

127 **3. Methods**

128 **3.1 Numerical model used for this study**

129 We used the Delft-3D and SWAN models (Deltares, 2011) for simulation of tsunami and storm
130 hydrodynamics and sediment transport because such applications of these models have already been
131 extensively validated (e.g., Apotsos et al., 2011a, 2011b; Bricker and Nakayama, 2014; Bricker et al.,
132 2014; Watanabe et al., 2017). For tsunami simulation, we used Delft-3D alone, which implements
133 the shallow water equations when applied with 1 vertical layer. We note that the shallow water
134 equations cannot resolve tsunami soliton fission, which occurs over shallow seas such as the Sendai
135 Bay (Murashima et al., 2012). Soliton fission is a process in which a long wave divides into several
136 short waves due to non-linearity and dispersion effects (Japan Electric Power Civil Engineering
137 Association, 2017). Fukazawa et al. (2002) revealed that the inundation extent of tsunamis can be
138 simulated well without considering soliton fission, as it has little effect on the overall transport of
139 tsunami mass or momentum, and thus it is not an essential aspect to resolve when investigating the
140 inland extent of sediment transport.

141

142 For simulation of storm waves and surge, we used both Delft-3D and SWAN. Delft-3D calculates
143 current fields, then passes the numerical result to SWAN (Deltares, 2011). SWAN calculates spectral
144 parameters of the wave field via the conservation of wave action. Then, SWAN passes the numerical
145 result for radiation stresses back to Delft-3D, so that mean current fields induced by wave setup are
146 calculated. In this way, by running Delft-3D and SWAN together, wave fields and current fields

147 induced by waves were calculated. We note that SWAN is a phase-averaged spectral wave model
148 that cannot resolve low-frequency infragravity wave motion such as surf beat. However, this effect
149 should be small on the Sendai Plain because it is located inside the broad and shallow Sendai Bay,
150 the bathymetry of which would dissipate infragravity motions (e.g., Roeber and Bricker, 2015).
151 Sediment transport was calculated with Delft-3D (Deltares, 2011) as in Apotsos et al. (2011a,
152 2011b), while bedload and suspended load were calculated using the formulation proposed by Van
153 Rijn (1993).

154

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

161

162 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,
163 respectively (Fig. 1). Topographic data were generated from the pre-2011 earthquake DEM data used
164 in Sugawara et al. (2014). To reduce computational load, sediment transport was calculated only in

165 domain 6.

166

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

170

171 Sugawara et al. (2014) revealed that sources of tsunami deposits were the seabed, beach, and sand
172 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
173 1.2-2.4 phi, respectively (Matsumoto, 1985). Therefore, we used a grain size of 0.267 mm (=1.9 phi)
174 for the sediment transport simulation. We determined the initial distribution of sand via a landuse
175 map together with data from Sugawara et al. (2014). Following Watanabe et al. (2017), we assumed
176 the initial sediment layer thickness was 5 m because a finite initial sediment layer thickness helps
177 avoid model instability due to excessive erosion and sedimentation. Hereafter, we define
178 “distribution limit” as the distance from the shoreline up to which sand sheets with thickness of more
179 than 5 mm were deposited (Abe et al., 2012, 2015; Sugawara et al., 2014; Watanabe et al., 2017).

180

181 **3.2 Tsunami and storm model boundary conditions**

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

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

189

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

200

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

208

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

215

216 **3.3 Validation of numerical model**

217 To validate our numerical model, we conducted a numerical simulation of inundation and sediment
218 transport during the 2011 Tohoku-oki tsunami (Fig. 3a). We examined the accuracy of our model by

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

227

228 **3.4 Inundation and sediment transport during a storm vs. a tsunami**

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

234

235 In general, both marine-sourced deposits and deposits originating from sand dunes could come from
236 either storms (e.g., Kortekaas and Dawson, 2007) or tsunamis (e.g., Szczuciński et al., 2012). Thus,

237 hereinafter, a movable sediment bed was assumed to exist everywhere from the sea floor to the sand
238 dunes, and the initial sediment layer thickness was 5 m as in Watanabe et al. (2017).

239

240 **4. Results**

241 **4.1 Validity of the tsunami simulation results**

242 Results of the numerical tsunami simulation were verified by comparing measured (Mori et al.,
243 2011) and computed flow depths at 39 points using the parameters K and κ (Aida, 1978). K and κ
244 indicate geometric average value and fluctuation, respectively, in the ratio of observed to computed
245 amplitudes. In our case, $K=0.95$ and $\kappa=1.21$ (Fig. 4). Takeuchi et al. (2005) suggested that $0.8 \leq K \leq$
246 1.2 and $\kappa \leq 1.6$ are required to accurately reproduce measured values. Our results showed that both κ
247 and K were within these required ranges. Calculated flow depths were less than the measured values
248 at sites where large flow depths (> 8 m) were recorded (Fig. 4). All these sites were located near the
249 shoreline, with measurements in or behind the coastal forest. When a tsunami strikes a coastal forest,
250 trees can fall due to hydraulic force; therefore the effect of the coastal forest on reducing the energy
251 of the tsunami might weaken as inundation continues. However, this effect is not included in our
252 simulation, and this may explain why simulated flow depths at these sites were not consistent with
253 measured values. Nevertheless, flow depths at the other points were consistent with the measured
254 values, and this qualitative agreement is considered a practical indicator of model validity for

255 tsunami inundation modelling.

256

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

262

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

268

269 We also compared the observed and calculated volumes of sandy deposits 0~1 km, 1~2 km, 2~3 km,
270 and 3~4 km inland from the shoreline, respectively (Table 2). On both transects, modeled sand
271 volumes overestimated the measured volumes at 0~1 km and 2~3 km (Table 2). A reason for this
272 discrepancy may be that density stratification for sediment transport or a suspended load

273 concentration limit was not included in our sediment transport modelling. Thus, the modeled
274 suspended load was high, and the calculated sand volumes overestimated the measured volumes.

275

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

288

289 The calculated tsunami deposit sand thickness is generally consistent with measured values (Fig. 5,
290 6). However, some discrepancies exist. On transect B, the measured thickness 600 m inland from the

291 shoreline was 11 cm, while the calculated thickness at this site was 47 cm. From 2200~3000 m
292 inland, the calculated sand thickness was overestimated compared to measured values by more than
293 10 cm. On transect A, the measured thickness was several cm from 440~600 m inland, but the
294 calculated value was 24~39 cm. Furthermore, from 2100~2400 m inland, the calculated deposit
295 thickness was overestimated by more than 10 cm. These inconsistencies are partially due to the
296 coarseness of the model grid resolution. The calculated sediment thickness at each grid cell is an
297 averaged thickness of the 15 m × 15 m area. Thus, as reported by previous studies (e.g., Sugawara et
298 al., 2014), the comparisons between simulated and observed thickness on a point by point basis is
299 difficult, even though overall trends agree well.

300

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

306

307 **4.2 Inundation and sediment transport during a tsunami**

308 In Section 4.1, we used the topography with coastal structures to validate the model. However, as

309 described in section 3.4, in order to compare inundation and sediment transport during a hypothetical
310 storm with tsunami under the natural environment, we use the topography without these artificial
311 structures hereafter.

312

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

317

318 The simulated distribution limits of deposits on Transects A and B were 3.3 km and 3.9 km,
319 respectively (Fig. 9a). The calculated maximum thickness of sandy deposits was 31 cm at a point
320 0.53 km inland from the shoreline on Transect A, and 29 cm at a point 0.60 km inland on Transect B.

321 In the finest computational domain, the total volume of sand deposited over land was $1.1 \times 10^6 m^3$,
322 while the total volume eroded from the sea floor to the sand dunes was $1.8 \times 10^6 m^3$. The calculated
323 volume deposited is not equal to the volume eroded because the rest of the sediments are transported
324 and deposited offshore.

325

326 We also show the distribution of suspended transport (Fig. 10) and bedload transport (Fig. 11) when

327 the tsunami reached the shoreline (Fig. 10a), when it ran overland (Fig. 10b), and when it reached its
328 maximum inland inundation extent (Fig. 10c).

329

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

337

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

343

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

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

349

350 **4.3 Inundation and sediment transport during storm surge**

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

358

359 The calculated thicknesses of sandy deposits on Transects A and B were 64 cm and 105 cm,
360 respectively (Fig. 9b). The inland extents of these deposits along Transects A and B were 0.19 km
361 and 0.27 km from the shoreline, respectively. In the finest model domain, the total volume of sand
362 deposited over land was $1.2 \times 10^5 \text{ m}^3$, while the total volume eroded was $1.4 \times 10^6 \text{ m}^3$. The majority of

363 the sand was deposited offshore because the maximum near-bottom mean orbital velocity of storm
364 waves and the current velocity of storm surge over land were very small, so sediments were not
365 transported over land. Significant sedimentation and erosion occurred near the estuary in the
366 northeast part of the domain (Fig. 9b) because the elevation at this site is low.

367

368 We also differentiated the distribution of suspended load (Fig. 12) and bedload transport (Fig. 13)
369 during the storm when surge inundated on the shoreline (Fig. 12a), when the surge propagated inland
370 (Fig. 12b), and when the surge reached its maximum inland extent (Fig. 12c). The maximum volume
371 flux of suspended transport when the surge reached the shoreline (Fig. 12a) was $1.9 \times 10^{-2} \text{ m}^2 \text{ s}^{-1}$ in
372 the river 0.53 km inland from the shoreline. When the surge propagated inland (Fig. 12b), the flux
373 was $1.9 \times 10^{-2} \text{ m}^2 \text{ s}^{-1}$ in the river 0.57 km inland from the shoreline. When it reached its maximum
374 inland extent (Fig. 12c), the flux was $2.1 \times 10^{-1} \text{ m}^2 \text{ s}^{-1}$ at the shoreline. The volume of suspended
375 sediments transported over land during the storm was smaller than that transported by the tsunami.
376 Regarding bedload transport, the maximum volume flux when the surge reached the shoreline (Fig.
377 13a), when it propagated inland (Fig. 13b), and when it reached its maximum inland extent (Fig.
378 13c) was $3.0 \times 10^{-4} \text{ m}^2 \text{ s}^{-1}$ at a point 0.10 km inland from the shoreline, $3.8 \times 10^{-4} \text{ m}^2 \text{ s}^{-1}$ at a point 0.21
379 km inland, and $2.0 \times 10^{-3} \text{ m}^2 \text{ s}^{-1}$ at a point 0.021 km inland, respectively. Bedload sediment transport
380 flux was also smaller than during the tsunami.

381

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

385

386 **5. Discussion**

387 **5.1 Difference between inundation and sediment transport extents during tsunami vs. storm**

388 The overland propagation speed of the tsunami was much greater than that of the storm surge. The
389 tsunami only took about 20 min from the time of incidence on the coast to reach its maximum inland
390 inundation extent, while the storm surge took almost 4 hrs. This is because a storm surge during a
391 passing typhoon can last as long as several hrs (Ministry of Agriculture, Forestry and Fisheries,
392 2015), while tsunami wave period is several tens of min (e.g., Nagai et al., 2007).

393

394 In our simulation, the distance inundation extended from the coastline was larger for storm surge
395 than for tsunami by several km (Table 5). This is because the elevation of the Sendai Plain is low
396 (0~3 m TP). Moreover, the duration of the storm surge was longer than that of the tsunami, and
397 lower storm surge flow speeds (Fig. 8b) corresponded to less dissipation of energy by bottom
398 friction than in a tsunami. Furthermore, the typhoon used in the storm surge simulation was stronger

399 than any event which has historically hit the region.

400

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

404

405 As described in Section 3.2, our assumed worst-case typhoon is unrealistically strong for the latitude
406 of the Sendai Plain. Thus, we additionally conducted a simulation with a typhoon which can
407 realistically make landfall there (see Appendix A) in order to investigate the storm surge inundation
408 and sediment distribution by such a typhoon. The modeled inundation distances of storm surge and
409 waves during this typhoon were 4.6 km and 4.2 km on Transect A and B, respectively, which are
410 slightly longer than those of the modeled tsunami inundation distances on these transects (4.4 km
411 and 4.0 km, respectively) (Table 5). Therefore, it should be emphasized that even a realistic
412 typhoon-generated storm surge can inundate farther inland than a tsunami. This is because of the
413 topographic setting of the Sendai Plain. The height of sand dunes located near the coastline is about
414 3 m, while the elevation along both transects up to 4~5 km inland from the coastline is <3 m. Thus,
415 once storm surge overtops the sand dune, it can easily inundate 4~5 km inland.

416

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

426

427 The volume of tsunami-induced erosion in the smallest model domain was larger than the volume of
428 storm-induced erosion. This is because tsunami-induced flow speeds near the coastline (the region of
429 most erosion) are stronger than storm surge flow speeds (Fig. 8), resulting in larger bed shear
430 stresses. Likewise, the volume of tsunami deposits was larger than that of storm deposits. This is
431 because of the difference in flow speeds farther upland (where most of the deposition occurred). On
432 both transects A and B, the overland flow speed of the tsunami was faster than that of the storm
433 surge (Table 5), so that the former was able to keep sediments in motion until they were transported
434 far inland (Fig. 9). Slow flow speeds under the storm surge, on the other hand, caused sediments to

435 settle out and cease motion much closer to shore. This held true for both suspended and bedload
436 transport (Fig. 12, 13).

437

438 **5.2 Relationship of tsunami and storm deposits to topography**

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

449

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

453 (2016) also reported that isolated sandy storm deposits on Leyte Island (Philippines) formed during
454 Typhoon Haiyan were found up to 1.7 km inland from the shoreline, where the inundation limit was
455 2.0 km. Therefore, the maximum extent of discontinuous sand deposits may also be affected by local
456 topography.

457

458 **5.3 Identification of tsunami vs. storm deposits**

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

471 the tsunami inundated more than 1000 m inland from the shoreline and sandy deposits also formed
472 more than 1000 m inland. These studies reinforce the hypothesis that sandy deposits distributed
473 several km inland can be confidently identified as of tsunami origin (Fig. 14b, d).

474

475 Morton et al. (2007) also compared the inundation distance and maximum extent of sand due to a
476 tsunami and a storm. They found that the inundation limit of storm surge is 10^2 - 10^4 m and the
477 maximum extent of its sand deposit is less than several hundred m, while the tsunami inundation
478 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
479 is gently sloping, the trends of inundation distance and maximum extent of sand due to the tsunami
480 and the storm shown in Fig. 14 are similar to those of Morton et al. (2007). However, if the land
481 slope is steep, the inundation distance of storm surge and waves will be small, in contrast to Morton
482 et al. (2007), while the inundation distance of the tsunami and the maximum extent of sand deposits
483 are 10^2 - 10^3 m, similar to Morton et al. (2007).

484

485 Some researchers reported that the distribution distance of storm deposits is smaller than that of
486 tsunamis (e.g., Goff et al., 2004; Tuttle et al., 2004; Kortekaas and Dawson, 2007; Morton et al.,
487 2007). This is because the storm surge flow speed (and therefore bed shear stress) several hundreds
488 to thousands of m inland is smaller than that of tsunamis over either steeply or gently sloping land.

489 Thus, sand layers distributed up to several km inland may be identified as of tsunami origin.
490 However, thin (mm-thick) storm deposits may be locally formed in low-lying depressions even
491 several km inland from the shoreline (Fig. 14a) as mentioned in Watanabe et al. (2017). Soria et al.
492 (2017) also suggested that the inland extent and thickness of storm deposits are governed by local
493 variations in topography or vegetation, so that the identification of tsunami or storm deposits may
494 not be conducted by assessment of the inland extent and thickness of storm deposits alone.

495

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

500

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

507 might also be affected by the wave state, local topography, or local vegetation.

508

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

515

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

520

521 **5.4 Future research on differentiation of tsunami vs. storm deposits**

522 In areas such as coral reefs with very steep offshore slopes, energetic low-frequency wave motions
523 such as surf beat can be generated by storm waves (Roerber and Bricker, 2015). In this case,
524 suspended and bedload sediment transport due to storm waves and surge overland could be large,

525 and the corresponding sand deposits could extend far inland.

526

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

537

538 **6. Conclusions**

539 We quantitatively examined the difference in the distribution of tsunami and storm deposits on the
540 Sendai Plain based on sediment transport simulations during both types of events. Our numerical
541 results indicated that the simulated inland inundation distance of a large hypothetical storm surge
542 was greater than that of the 2011 Tohoku-oki tsunami by several km. Even a realistic typhoon can

543 produce a storm surge inundation extent slightly greater than that of the 2011 event. Based on the
544 assumption that storm surge inundation extent is generally smaller than that of tsunamis, previous
545 studies identified sandy deposits which are distributed farther inland from the coastline as being of
546 tsunami origin. However, over gently sloping topography such as the Sendai Plain, storm surge
547 inundation can be more extensive than tsunami inundation, so that differentiation of their deposits is
548 not possible by only confirming the inundation extent. Over steep topography such as in Noda
549 village, the maximum inundation distance and inland extent of sand deposits due to storm surge
550 become small, so sandy deposits distributed several km inland from the coastline can be identified as
551 of tsunami origin. Under any topographic condition, the maximum inland extent of continuous storm
552 sand deposits is small because the flow speed of storm surge and waves over land is small. Thus,
553 sandy deposits which are continuously distributed several km inland can be identified as of tsunami
554 origin. For the same reason, the total volume of storm deposits may be much smaller than the total
555 volume of tsunami deposits over any topography, so the total deposit volume could also be a useful
556 proxy to differentiate these deposits. However, discontinuous sandy storm deposits might be formed
557 in low-lying depressions several km inland from the coastline, and the total volume of deposits
558 formed by a small tsunami may also be small.

559

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565

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

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

579 1963). Thus, the central pressure of the typhoon and the wind speed used for the simulation were 920
580 hPa and 38 m s^{-1} , respectively. The calculated distribution of maximum water level was as shown in
581 Fig. A.1. The inundation distances along Transects A and B were 4.6 km and 4.2 km, respectively.

582

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740 **Figure Captions**

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

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

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

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

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

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

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

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

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

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

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

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

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

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

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

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

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

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

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

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