

Spatial extension of the mid to late Holocene sedimentary record of tsunamis along the Southern Kuril Trench, Hokkaido Japan

Ryo Nakanishi^{1,1}, Juichiro Ashi^{2,2}, Yosuke Miyairi^{2,2}, and Yusuke Yokoyama^{2,2}

¹Atmosphere and Ocean Research Institute, The University of Tokyo

²University of Tokyo

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Abstract

Infrequent megathrust earthquakes, with their complex cycles and rupture modes, require a high-resolution spatiotemporal record of tsunami inundations over thousands of years to provide more accurate long-term forecasts. The geological record suggests that $M_w > 8$ earthquakes in the Kuril Trench occurred at intervals of several hundred years. However, uncertainties remain regarding the rupture zone, owing to the limited survey areas and chronological data. Therefore, we investigated the tsunami deposits in a coastal wetland of southeastern Hokkaido, Japan, to characterize the tsunamis that have originated from the Kuril Trench over the last 4000 years. On the Erimo coast, more than seven sand layers exhibited the common features of tsunami deposits, such as sheet distributions of several hundred meters, normal grading structures, and sharp basal contacts. According to numerical tsunami simulations, the 17th-century sand layer could be reproduced by using a multiple rupture zone model ($M_w \sim 8.8$). We used high-resolution radiocarbon dating and tephras to correlate the tsunami deposits from the last 4000 years with those reported from regions ~ 100 km away. The tsunami history revealed here shows good agreement with histories of adjacent regions. However, the paleotsunamis reported to have occurred in regions > 200 km away include some events that differ from those in this study, which suggests a diversity of $M_w > 8$ earthquakes in the Kuril Trench. We clarified the history and extents of earthquake-generated tsunamis along the southwestern end of the Kuril Trench, which were previously unknown. Our results provide a framework for magnitude estimations and long-term forecast of earthquakes.

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45

46 **Keywords:** Tsunami deposits, Kuril Trench, Radiocarbon dating, Numerical simulation,
47 Hokkaido, Erimo

48

49 Plain Language Summaries

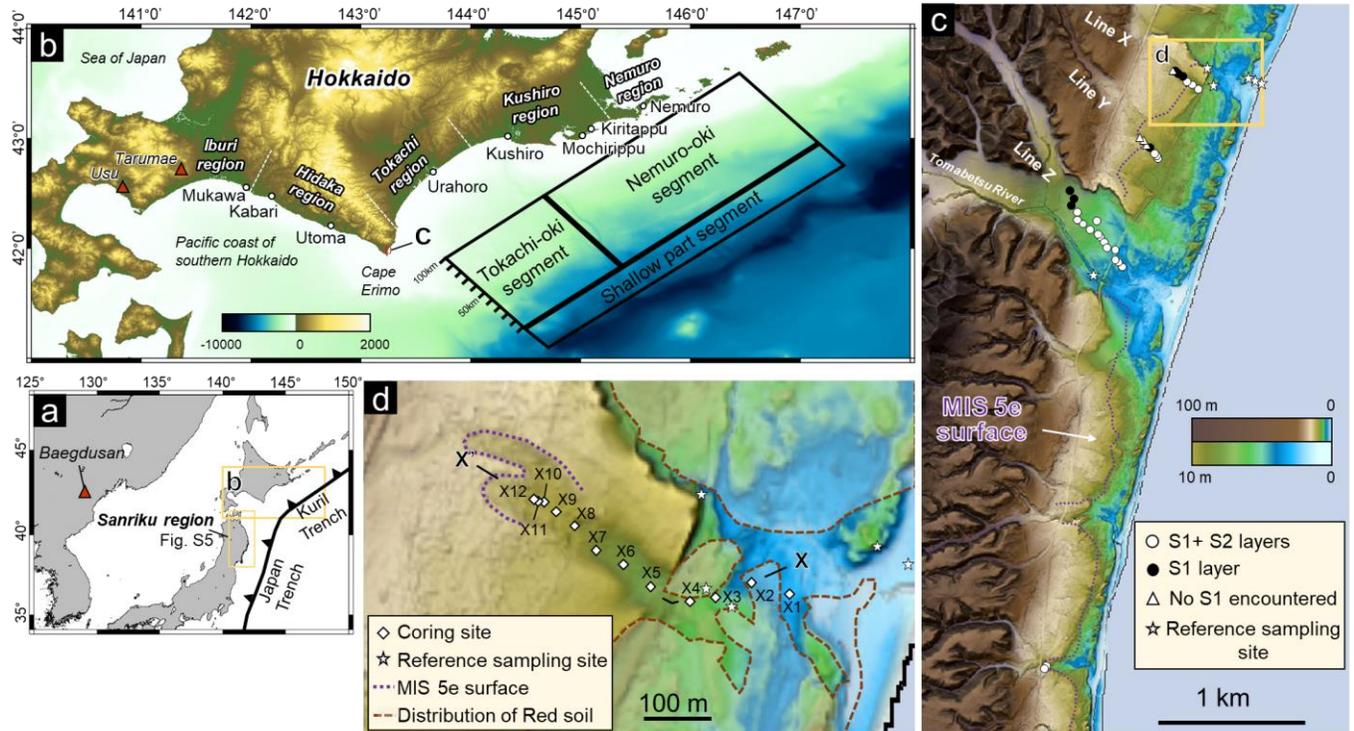
50 Long-term assessments of infrequent large earthquakes and tsunamis are highly uncertain due to
51 the complexity of their cycles and rupture zone. Unveiling the nature of past tsunamis is critical
52 for understanding the complex mechanisms of megathrust earthquakes. However, the period of
53 time covered by observational data is often insufficient to characterize infrequent events. High-
54 resolution spatiotemporal records of coastal inundation caused by tsunamis over thousands of
55 years are therefore required to constrain geophysical models. In this study, we performed
56 geological surveys along the southeastern coast of Hokkaido, Japan. Ten tsunami-derived sand
57 layers were newly observed. According to numerical simulations, the 17th century sand layer
58 could be reproduced by a tsunami from the earthquake model in the Kuril Trench. The tsunami
59 history is well matched with that of the surrounding region facing the Kuril Trench. However,
60 some layers did not correlate with sand layers in in regions > 200 km away, suggesting that
61 another process produced these sand layers. We found that some of the earthquakes categorized

62 as "unusually large earthquakes" included multiple ruptures, and the spatial distribution of sand
63 layers suggests localized or differing rupture modes from that of the 17th century event.
64

65 **1 Introduction**

66 Infrequent megathrust earthquakes (M~9: e.g., the 1700 Cascadia, 1960 Chile, 2004
67 Sumatra, and 2011 Tohoku-oki earthquakes) have been reported to occur over wider rupture
68 zones and with prolonged recurrence intervals of hundreds to thousands of years compared to
69 M7–8 earthquakes (Sieh et al., 2008; Goldfinger et al., 2013; Philibosian & Meltzner, 2020;
70 Salditch et al., 2020). To study these infrequency megathrust earthquakes, prehistoric data with
71 timescales ranging from thousands to tens of thousands of years are required. Prehistoric data
72 can be obtained only from the geological record. For instance, these data can be obtained from
73 uplifted terraces, deep-sea turbidites, and tsunami deposits (Goldfinger et al., 2013; Satake,
74 2015; Hutchinson & Clague, 2017). Every subduction zone has a high danger of M~9 earthquake
75 recurrence because the instrumentally recorded earthquakes do not completely relieve the strain
76 in such zones. Therefore, it is necessary to investigate the case studies of infrequency megathrust
77 earthquakes from various regions as event clusters, and the changes in rupture zone patterns
78 differ for each subduction zone (Nelson et al., 2006; Philibosian & Meltzner, 2020; Salditch et
79 al., 2020).

80 The geological evidence provided by tsunami deposits and seismic crustal deformation
81 along the Pacific coast of eastern Hokkaido indicates that the southern Kuril Trench has
82 repeatedly ruptured at intervals of several hundred years, with the most recent event occurring in
83 the 17th century (Nanayama et al., 2003; Sawai et al., 2004, 2009; Kelsey et al., 2006; Sawai,
84 2020). Based on the evidence provided by the tsunami deposits from the 17th century, which
85 have been found at many sites in eastern Hokkaido, it is suggested that this earthquake was Mw
86 8.8 or greater based on the numerical tsunami simulations of (Nanayama et al., 2003; Ioki &
87 Tanioka, 2016). Therefore, another rupture is expected to occur in the near future, as ~400 years
88 have passed since the 17th-century earthquake (Sawai, 2020). It is necessary to clarify the
89 detailed magnitudes of Mw>8 earthquakes via geological surveys over wide areas since there are
90 limited historical documents regarding the paleotsunamis in Hokkaido. More than 17 sand layers
91 have been reported to have been deposited over several thousand years in east Hokkaido
92 (Nanayama et al., 2003; Sawai et al., 2009: Fig. 1). Some of these sand layers were deposited by
93 “unusually large” earthquake-generated tsunamis, such as the 17th-century tsunamis, which could
94 not be explained by the observed historic tsunamis. For events that took place prior to the 17th
95 century, chronological correlations have rarely been examined, but accurate correlations of these
96 tsunami records are important for understanding the recurrence intervals and rupture modes of
97 megathrust earthquakes.



98

99 Figure 1. Study area. a: Location map of Hokkaido and Baegdusan. The black barbed line indicates the Pacific
 100 Plate subduction zone. b: Bathymetric and topographic map of Hokkaido. The paleotsunami investigation
 101 sites and local sources of tephra described by Nakanishi et al. (2020b) are marked. The black boxes indicate
 102 the area of the 17th-century earthquake fault model (Ioki & Tanioka, 2016). c: Topographic map based on 5
 103 m-grid digital elevation model data from the Erimo area. The circles and triangles indicate the locations of
 104 coring sites. Additionally, the circles show the distributions of layers S1 and S2. The purple dashed lines
 105 indicate the distribution of the MIS 5e surface (Koike & Machida, 2001). The stars show the sand sampling
 106 locations for beaches, rivers, and Red soil. The yellow box indicates the map shown in Fig. 1d. d: Close-up
 107 map of the coring sites that are located along Line X. The distribution of Red soil in 1947 is shown, as
 108 interpreted from aerial photographs. The stars are the same as in Fig. 1c.

109 In this study, we investigated the eastern coast of Cape Erimo to identify tsunami
 110 deposits and to estimate their spatial distributions and to also estimate the wave sources of the
 111 17th-century tsunami deposits that were newly discovered in this area by using numerical
 112 simulations. The tsunami deposits are correlated with the event layers in the surrounding regions
 113 where the event ages have already been reported based on ¹⁴C dating and tephrochronology. This
 114 study discusses the tsunami history and coastal extents over the last 4,000 years along the Kuril
 115 Trench from the Tokachi to Hidaka region based on tsunami age correlations.

116

117 2 Background

118 2.1 Historical tsunami records

119 Instrumentally recorded earthquakes over the past 200 years in the southern part of the
120 Kuril subduction zone are known to occur at magnitudes of Mw 7–8 and fault lengths of ~100
121 km with maximum lengths of 200 km (Satake, 2015). The rupture zone is classified into two
122 segments, Tokachi-oki and Nemuro-oki (Fig. 1). Interseismic GPS data and seismic wave
123 velocity studies suggest a gap in the asperity distributions in these rupture zones (Hashimoto et
124 al., 2009; Liu et al., 2013). Mw 7–8 earthquakes are known to occur at intervals of 50–100 years,
125 with the Tokachi-oki or Nemuro-oki segments acting as rupture faults (Satake, 2015; Sawai,
126 2020). The most recent southern Kuril Trench earthquake to cause significant damage was the
127 2003 Tokachi-oki earthquake (Mw 8.0), which generated a tsunami with a maximum wave
128 heights of 4 m (Tanioka et al., 2004). This tsunami caused traces debris deposition and minimal
129 erosion of up to 3.5 m above present sea level (asl) in the Erimo area of Hokkaido, however, no
130 sandy deposits were formed (Nishimura et al., 2004). Earlier historical tsunamis originating from
131 the Kuril and Japan trenches generated wave heights ranging as 3–4 m along the Erimo coast.
132 However, no significant sandy deposits were created by these tsunamis (Nakanishi et al., 2020a).

133 2.2 Previous work on geological records of tsunami

134 There are no historical records for Hokkaido before the 19th century (Satake, 2015), but
135 geological investigations of the tsunami deposits and seismic crustal deformations have been
136 reported (Nanayama et al., 2003; Sawai et al., 2004, 2009; Kelsey et al., 2006; Szczuciński et al.,
137 2016; Ishizawa et al., 2017; Nakanishi et al., 2020a). In the region from Tokachi to Nemuro,
138 extensive surveys of tsunami deposits have been carried out (Nanayama et al., 2003). Based on
139 tephrostratigraphic correlations, previous studies have identified two sand layers that were
140 deposited between the 17th and 10th centuries, and 2–4 sand layers were identified to have formed
141 between 1000 and 2400 years ago. These tsunami deposits have anomalous distribution heights
142 and extents that cannot be reproduced by tsunamis that were generated by Mw~8 earthquakes,
143 such as the 1952 and 2003 Tokachi-oki earthquakes (Nanayama et al., 2003, 2007; Sawai et al.,
144 2009). These data suggest that these deposits were caused by tsunamis that were generated by
145 Mw~9 earthquakes that occurred in the rupture zones along the Tokachi-oki and Nemuro-oki
146 segments. However, detailed correlations among these sand layers are unclear due to a lack of
147 comprehensive radiocarbon data (Sawai, 2020). The tsunami history of Kushiro to Nemuro
148 regions has been investigated based on widespread tephra and radiocarbon dating of tsunami
149 deposits for the past 7,000 years (Nanayama et al., 2003, 2007; Sawai et al., 2009). However, the
150 age comparisons of these regions are difficult because of the large errors in the obtained ¹⁴C
151 ages. Precise radiocarbon dating has been reported for the tsunami deposits over the past 3000

152 years in the Urahoro, Tokachi region (Ishizawa et al., 2017), which is located ~100 km from
153 Erimo (Fig. 1).

154 2.3 Regional setting

155 The eastern coast of Cape Erimo is located at the southern end of the Hidaka region and
156 faces the Tokachi-oki segment (Fig. 1). This coastal area contains an ~200 m wide sandy beach
157 where the maximum dune height is ~10 m asl. The first recorded settlement with a large number
158 of immigrants was in 1872. Desertification of the area progressed owing to deforestation until a
159 greening project was initiated in 1953 (Sakuraba, 2019). Aerial photographs taken in 1947 show
160 that the coast was originally covered with aeolian redeposited volcanic ash material called "Red
161 soil", which extended over 500 m perpendicular to the coastline (Fig. 1d). The seafloor
162 sediments on the eastern side of Cape Erimo consist of poorly sorted very fine- to fine-grained
163 sand (Noda & Katayama, 2011). Glauconite grains and sessile organisms were found east of
164 Cape Erimo (Noda & TuZino, 2010).

165 Terrace surfaces of marine isotope stage (MIS) 5e are widely distributed in this area (Fig.
166 1c), and relative sea-level change can be inferred from this elevation. Based on glacio-hydro
167 isostatic adjustment (GIA) models, the relative sea-level change due to the GIA over the past
168 6000 years in the Erimo area was estimated to be 0 ± 1 m (Okuno et al., 2014). Marine terraces
169 during MIS 5e extend parallel to the coast at ~15–25 m asl (Koike and Machida, 2001). The
170 uplift rate is 0.02–0.17 mm/yr based on the MIS 5e terraces on the Erimo coast, when GIA
171 estimations of 125 ka are considered (Okuno et al., 2014). The Erimo coast consists of a
172 microtidal area where the maximum tidal range is < 1 m and the difference between mean sea
173 level and mean higher high water is ~0.4 m (J-DOSS:
174 <https://www.jodc.go.jp/jodcweb/JDOSS/index.html>).

175 We surveyed the wetlands along three survey lines (Fig. 1). Lines X and Y traverse the
176 depressions on the marine terraces. A depression oriented perpendicular to the coastline on the
177 marine terrace is considered to be a former riverbed, which is flat near the terrace surface of MIS
178 5e, suggesting that the rivers ceased to flow before at least 125 ka. Line Z is parallel to the
179 Tomabetsu River, and this area is covered by a thick peat layer. The inland area was not
180 surveyed owing to the construction of residential areas.

181 3 Methods

182 3.1 Field surveys

183 Core samples were obtained from three survey lines (40 sites) to investigate the spatial
184 distribution of the tsunami deposits, especially those that were laid down after the 10th century.
185 The samples were extracted using a handy Geoslicer (length of 0.6 m: Takada et al., 2002) and
186 peat sampler (diameter of 7 cm, length of 2.5 m). We sampled the pre-10th-century sediments

187 based on the B-Tm tephra layer to depths of 1–2.5 m to investigate the changes in the sand layers
188 in the inland direction along Line X. The variation of layer thickness was confirmed to be as
189 small as 1 cm from the average value by the repeat hand-boring check at several sites (Table S1).
190 The core samples were described by their sedimentary facies (e.g., colors, grain sizes,
191 thicknesses, bedforms) and were photographed. Thereafter, the whole cores were placed in
192 plastic cases and were subsampled in the laboratory for each laboratory analysis. Beach, river,
193 and aeolian sand (Red soil) samples around the peatland were collected to examine the sources
194 of the sand layers (Fig. 1). The elevation profile along survey Line X was measured via real-time
195 kinematic positioning using multiband global navigation satellite system receivers (GNSS: ZED-
196 F9P U-blox). The data pertaining to forests, where the signal reception was not consistent, were
197 supplemented by correcting the Digital Elevation Model (DEM:
198 <https://fgd.gsi.go.jp/download/menu.php>) data of the Geographical Survey Institute (GSI) to
199 ensure data matching with respect to the GNSS data.

200 3.2 X-ray diffraction analysis

201 Since the mineral combinations of the deposits may reflect the sediment sources as well
202 the sedimentation conditions (Jagodziński et al., 2012; Nakamura et al., 2012), X-ray
203 diffractometry (XRD) analyses were conducted to study the mineralogy of the sand layers.
204 Twenty-three sand and mud samples were collected from the core samples, and the candidate
205 source sand samples. The sand samples were sieved to separate the 0.18–0.25 mm fraction. We
206 also analyzed the 0.090–0.063 mm fraction of the sand samples to determine the grain size
207 effects. The mineralogical compositions were determined by XRD (Bruker D2 PHASER) and
208 were determined by the same method described in Nakanishi et al. (2020a). The analysis
209 conditions were set at 30 kV, 10 mA, 0.02 steps for 1 s, and a 2θ registration range from 5° to
210 65° .

211 3.3 Grain size analysis

212 The grain size distributions of sand layers are widely used to determine their origins and
213 transport modes (e.g., Folk & Ward, 1957; Morton et al., 2007). We performed grain size
214 analyses by using the sieve method on the sand layers that were obtained from the cores and the
215 surrounding sands (e.g., beach, river, and aeolian sands). We sampled 47 samples in total, which
216 were obtained every 1 cm in the vertical direction at site X5 and from the bulk layers at other
217 sites. Samples of 10–20 g were pretreated with H_2O_2 to remove the organic matter and disperse
218 the particles. The dried samples were sieved by using a set of 14 sieves with mesh sizes ranging

219 from 4.5 phi to -2.0 phi. The basic statistics such as the mean (Mz), standard deviation (σ_I), and
220 skewness (Sk_i) were calculated by following the method outlined in Folk & Ward (1957).

221 3.4 Computed tomography (CT) and magnetic susceptibility analysis

222 X-ray CT image scanning was performed to confirm invisible sedimentary structures of
223 the core samples due to density differences. Analysis was performed with a slice width of 0.5
224 mm using an Aquilion PRIME Focus Edition (Canon Medical Systems Corporation) at the Kochi
225 Core Center.

226 Magnetic susceptibility measurements were performed to characterize the sediments and
227 to identify magnetite. Cores sampled in plastic cases were kept at room temperature and then
228 measured every 1 cm using an MS3 susceptometer (Bartington Instruments) and a susceptibility
229 accuracy of $\pm 2 \times 10^{-6}$ SI. We corrected for atmospheric conditions per 10 measurements.

230 3.5 Diatom analysis

231 Diatom fossils transported with sand layers have information about the source, as some
232 diatoms have specific habitats. Samples were taken from pure sand layers that were not muddy
233 and peat layers below sand layers. The sand layers were sampled from the middle of the sand
234 layer to avoid the inclusion of peat materials such as rip-up clasts. Optical microscopic
235 observations were performed in following Nakanishi et al. (2022a). In total, 200 diatom valves
236 were counted and, expressed as percentages; species or genera present by $> 3\%$ were used for
237 assemblages analysis.

238 3.6 Radiocarbon dating

239 Radiocarbon dating was used to determine the depositional ages of the tsunami deposits
240 (Table 1). Analytical samples (units of 1 cm) were collected above and below the sand layers and
241 consisted of plant fragments in peat while avoiding roots. When plant fragments were not
242 available, bulk samples were used for the measurements. Graphitization of the samples was
243 achieved via the same method employed by Nakanishi et al. (2020a), and ^{14}C dating was
244 prepared at the Atmosphere and Ocean Research Institute using single-stage accelerator mass
245 spectrometry (Yokoyama et al., 2019). We calibrated the ^{14}C ages to the calendar ages using
246 OxCal 4.4 with the IntCal20 dataset (Reimer et al., 2020). The P_Sequence and Sequence model
247 and general outlier model in OxCal were used to constrain the calibration ages by using the
248 stratigraphic order (Bronk Ramsey, 2008, 2009a, 2009b). Each model was constructed separately
249 for individual peat layers that were located among the tsunami deposits (Ishizawa et al., 2017,
250 2020). The tsunami recurrence intervals were estimated from the age differences between each
251 sand layer (Lienkaemper & Bronk Ramsey, 2009: Table S2). We recalibrated the ^{14}C ages of the

252 reported tsunami deposits (Ishizawa et al., 2017; Nakanishi et al., 2020a) using updated IntCal20
 253 (Reimer et al., 2020).

254 Table 1. ^{14}C dating results.

Sample name	Site	Depth (cm)	Material	^{14}C age	Error	Modelled age (1σ)		Modelled age (2σ)		Mean	Lab number
						from	to	from	to		
S1_2U	X5	-30	Bulk peat	392	± 25	500	461	511	335	469	YAUT-053023
S1_2L		-35	Bulk peat	814	± 24	714	683	733	677	702	YAUT-053019
S2L		-41	Bulk peat	862	± 22	776	732	897	725	766	YAUT-053018
S5U		-54	Bulk peat	1475	± 24	1379	1315	1393	1307	1352	YAUT-053017
S5_6U		-60	Bulk peat	1801	± 26	1728	1630	1780	1608	1684	YAUT-053016
S5_6L		-66	Bulk peat	1971	± 24	1925	1838	1977	1827	1888	YAUT-053015
S6_7U		-70	Bulk peat	2053	± 27	2051	1950	2098	1940	2012	YAUT-053013
S6_7L		-71	Bulk peat	2119	± 23	2120	2048	2145	2004	2082	YAUT-053012
S7L		-78	Plant fragment	2304	± 25	2350	2324	2358	2182	2322	YAUT-053011
S8U		-83	Plant fragment	2524	± 23	2719	2515	2732	2492	2587	YAUT-053009
S8_9U		-100	Plant fragment	2679	± 24	2841	2755	2850	2750	2792	YAUT-053006
S8_9L		-106	Plant fragment	3010	± 26	3240	3156	3331	3075	3198	YAUT-053005
S9_10U		-112	Plant fragment	3216	± 25	3460	3401	3480	3379	3428	YAUT-053004
S9_10L		-115	Plant fragment	3411	± 24	3672	3578	3703	3568	3631	YAUT-053003
S10L		-121	Plant fragment	3563	± 25	3897	3833	3969	3728	3863	YAUT-053002

Sample name	Site	Depth (cm)	Material	^{14}C age	Error	Calibrated age (1σ)		Calibrated age (2σ)		Mean	Lab number
						from	to	from	to		
X4_195cm	X4	-195	Plant fragment	3628	25	3976	3900	4071	3849	3942	YAUT-053024

The calibrated ages are reported in solar years before 1950 CE.

Analytical Laboratory: Atmosphere and Ocean Research Institute, the University of Tokyo.

255 Half-life of ^{14}C is 5730 year.

256 3.6 Numerical simulation

257 We performed numerical simulation to reproduce the sand layers in this area by assuming
 258 tsunami inundation that is consistent with the observed distribution of tsunami deposits in eastern
 259 Hokkaido using the source faults of the 17th-century earthquake (Satake et al., 2008; Ioki &
 260 Tanioka, 2016). If the model was inconsistent, the fault parameters were modified for
 261 optimization. The numerical simulations were performed using the tsunami calculation code
 262 “JAGRUS” (Baba et al., 2015, 2017). We used bathymetric data obtained from the Japan
 263 Oceanographic Data Center and M7000 series data obtained from the Japan Hydrographic
 264 Association. The ASTER GDEM Version 3 from the Shuttle Radar Topography Mission
 265 (<https://doi.org/10.5067/ASTER/ASTGTM.003>) and 5-m mesh DEM data from the GSI were
 266 used as topographical data. To avoid underestimating the inundation area, modern roads that are
 267 assumed to have not existed in the 17th century were manually removed and flattened using the
 268 surface command of the generic mapping tools (Wessel et al., 2013). We nested the geographical
 269 data in grid size in the following order: 450, 150, 50, 16, and 5 m. The calculations were
 270 performed using linear long-wave equations for the 450 m grid size system and non-linear long-
 271 wave equations for the finer grid size systems in a staggered-grid, leap-frog finite differential

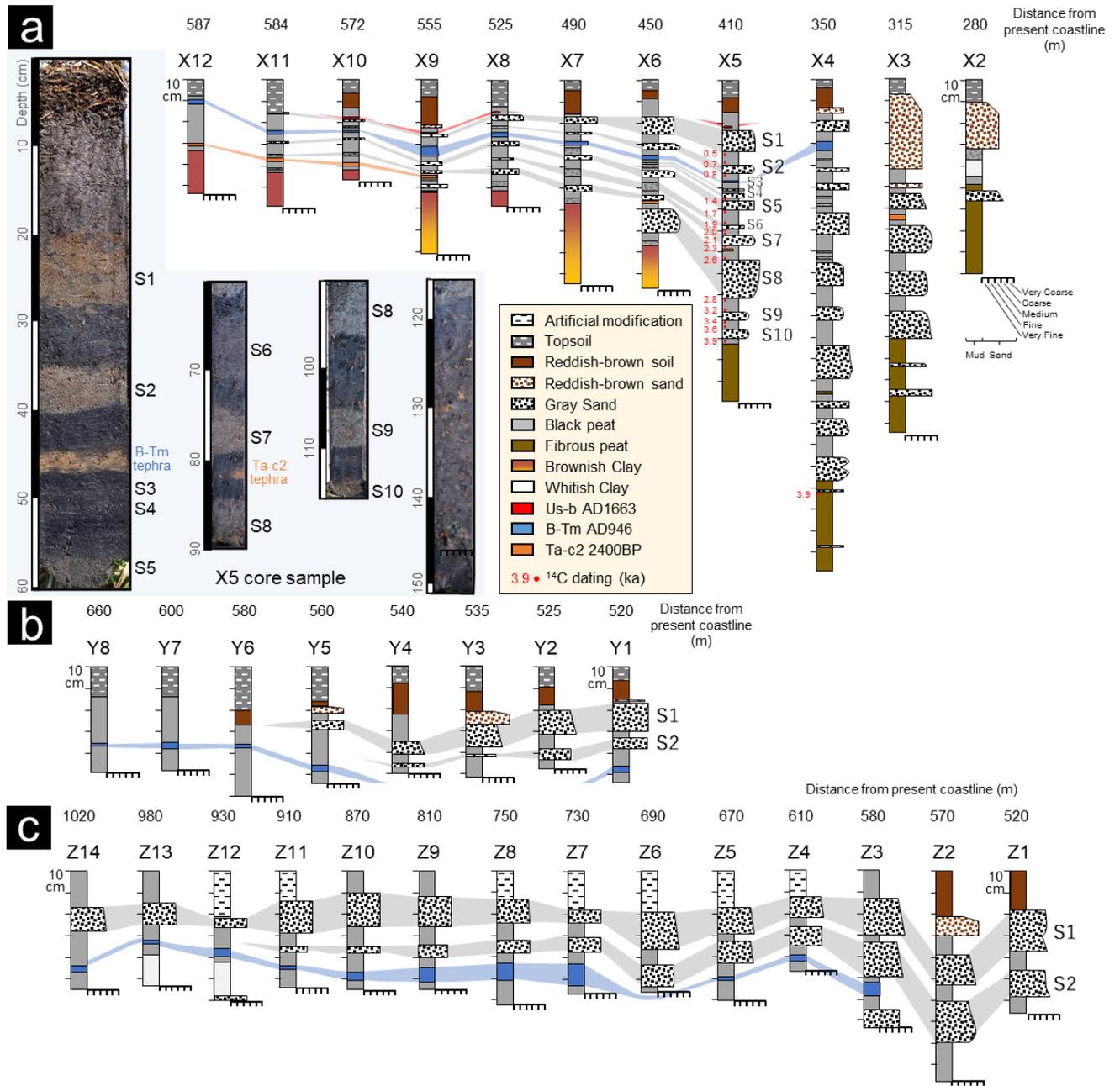
272 scheme. The absorbing boundary condition was applied at the edge of the computational domain.
273 The time step was 0.1 s to safely reach a stable condition. The values of Manning's roughness
274 coefficient was $0.03 \text{ m}^{-1/3}\text{s}$, based on Ioki & Tanioka (2016). The tide levels were assumed to be
275 constant at the present mean sea level. Seismic deformations were computed based on the
276 formula of Okada (1985), and the rise times were assumed to be 60 s (deformation was included
277 in the geographical data). The tsunami propagation calculations were conducted for as long as
278 7200 s.

279 **4 Description of tsunami deposits**

280 4.1 Stratigraphy and depositional ages

281 An outline of the stratigraphy on the survey lines observed by hand borings is described
282 in Fig. 2. From bottom to top: fibrous dark brown peat or clay layer, black peat intercalated by
283 gray sand layers, reddish-brown soil or sand as "Red soil", and topsoil. On the landward side of
284 the X6 site, the clay layer consists of pinkish-brown to yellowish-brown sticky clay at depths of
285 ~0.5 m or more from the ground surface (Figs. 2 and Fig. S1). On the seaward side of the X5

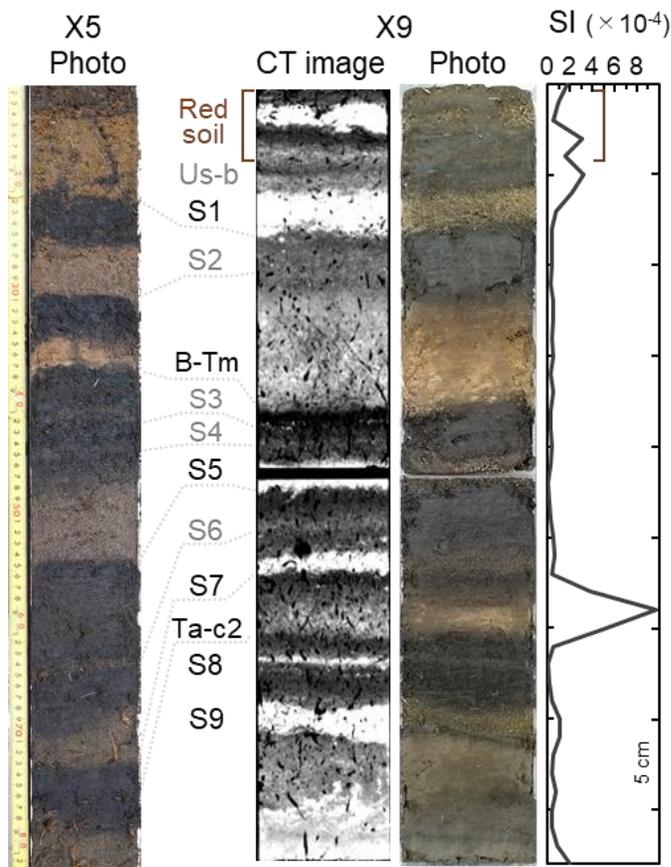
286 site, a fibrous peat layer was observed below ~1.5 m from the ground surface. Here, the sand
 287 layers were either thinner than 5 cm thick, or no sand layers were observed.



288
 289 Figure 2. Correlated stratigraphic columns along survey Line X, Y, and Z and photographs of the core samples
 290 from site X5.

291 Three tephra layers and ten sand layers were intercalated within the black peat (Fig. 2).
 292 The tephra layers were identified in this area by Nakanishi et al. (2020b), and from top to bottom
 293 consist of: a patchy white volcanic ash layer (1663 AD, Usu Volcano-b tephra: Us-b), a

294 yellowish-white volcanic ash layer with a thickness of ~2 cm (946 AD, Baegdusan Volcano-
 295 Tomakomai tephra: B-Tm), and an orange volcanic ash layer with patchy 2-cm thickness within
 296 the peat layers (~2400 BP, Tarumae Volcano-c2 tephra: Ta-c2). The chemical compositions of
 297 their volcanic glasses differ, reflecting the different source volcanoes (Fig. S2). Us-b is
 298 characterized by the presence of hornblende, mainly volcanic glasses. The distribution is
 299 discontinuous owing to the thin thickness and its disturbance by the Red soil layer. B-Tm is
 300 particularly easy to identify because it is composed of > 90% volcanic glass. The Ta-c2 layer is
 301 characterized by high magnetic susceptibility due to a high magnetite content (Fig. 3).

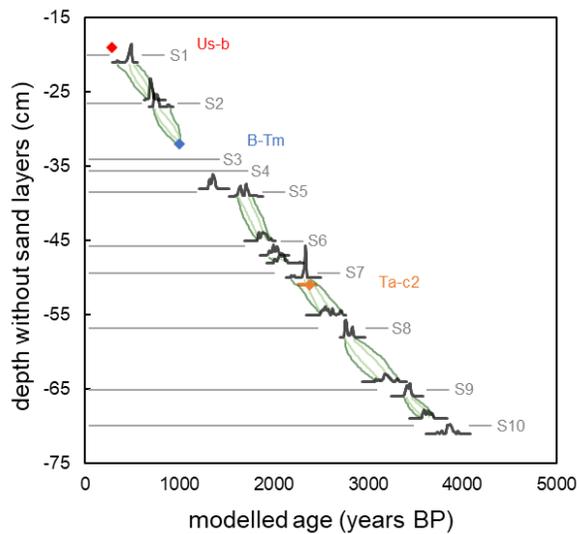


302

303 Figure 3. Computed tomography (CT) image and magnetic susceptibility of the X9 core. Gray letters show
 304 invisible sand layers identified in the CT image. A photograph of the X5 core is shown for comparison.

305 We observed two sand layers between the Us-b and B-Tm tephras, five sand layers
 306 between the B-Tm and Ta-c2 tephras, and three to five sand layers below the Ta-c2 tephra
 307 (named layers S1–10 from top to bottom). In the age–depth model, the depositional ages were
 308 obtained from the peat layers by ¹⁴C dating and the tephras were linearly correlated (Fig. 4). This
 309 model is also consistent with the known ages of the volcanic ash layers. The accumulation rate
 310 was constant (0.14 mm/yr), which indicated continuous deposition of the peat layers. The most

311 basal sand layer (i.e., layer S10) was deposited after 3900 BP. The recurrence intervals for layers
 312 S1–10 ranged from 110 to 620 years (Table S2).

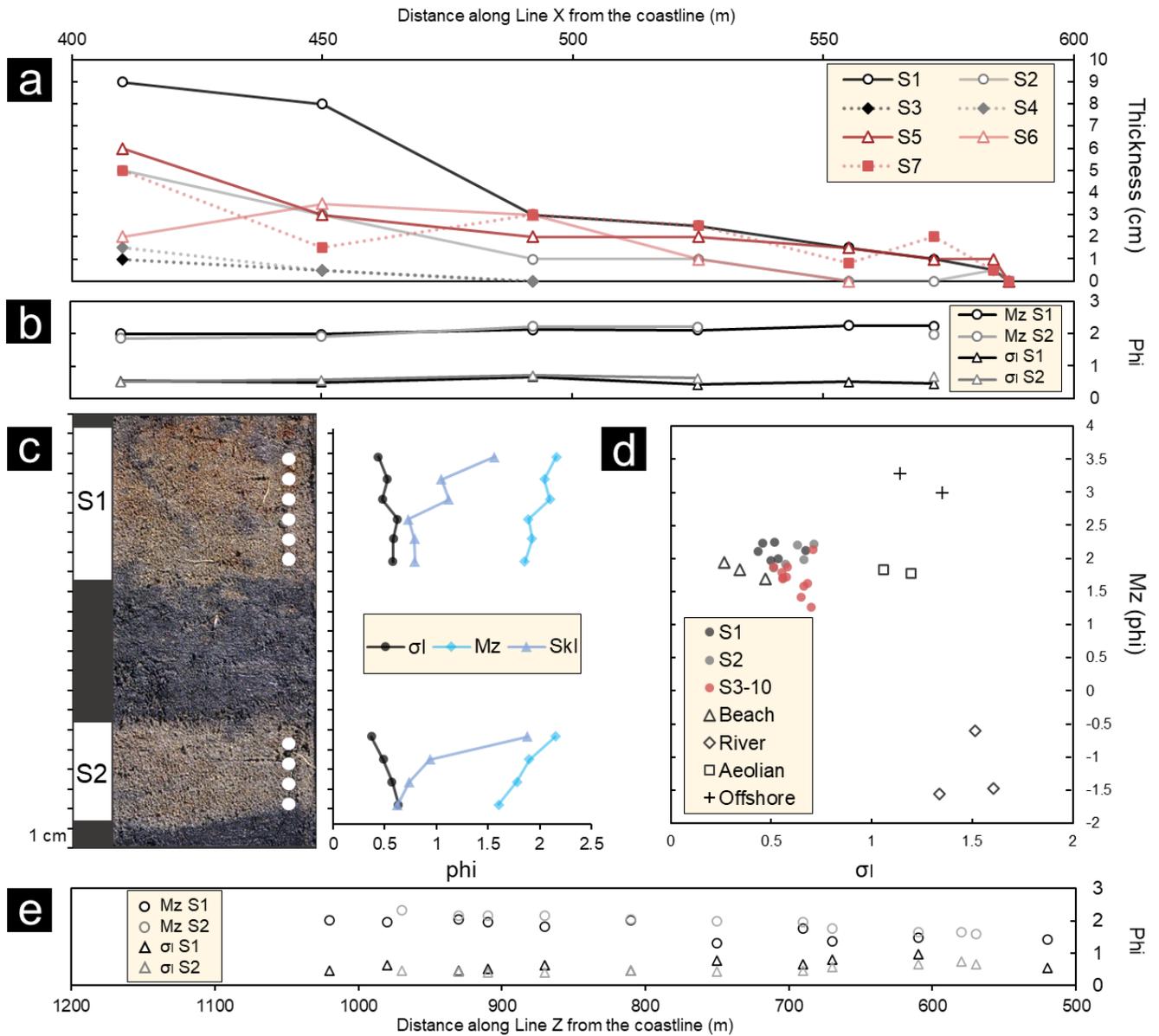


313

314 Figure 4. Age-depth model without the sand layers at site X5. Black histograms show the modeled ^{14}C ages
 315 (Table 1). The dark and light green lines indicate the ranges of the 2σ and 1σ modeled ages, respectively. Red
 316 and blue diamonds indicate the widespread tephras (e.g., Us-b: 1663 AD, B-Tm: 946 AD) that were
 317 incorporated into the age-depth model. Yellow diamonds and bars indicate the Ta-c2 median ages and 2σ
 318 ranges that were calculated by the same methods as used for the event layers, respectively. The calibrated
 319 ages are reported in solar years before 1950 CE.

320 A reddish-brown sand layer and soil were found as “Red soil” near the ground surface.
 321 The reddish-brown sand layer has a mode of 1.5 phi with poor sorting and a high background
 322 and low mineral peak XRD intensity (Figs. 5 and 6, Figs, S3 and S4), which indicates high levels
 323 of amorphous components such as a pumice. The magnetic susceptibility is higher than that of
 324 peats and gray sand layers because of its volcanic ash origin (Fig. 3). Therefore, the reddish-
 325 brown sand layers can be distinguished from the gray sand layers by their distributions near the

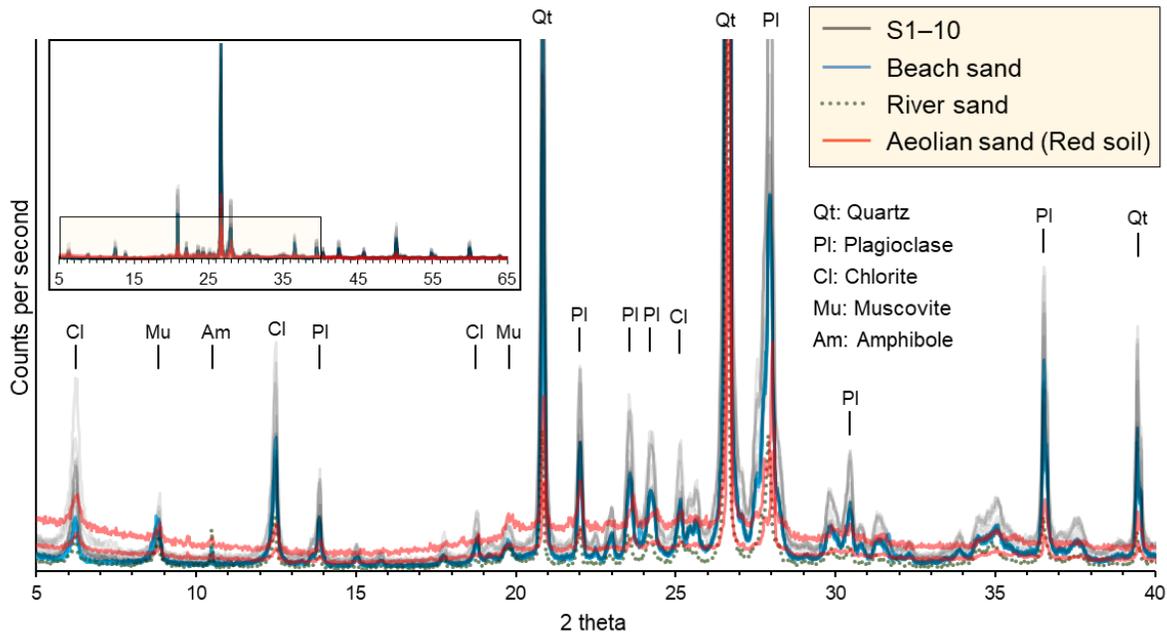
326 ground surface and poor sorting. The topsoil is black to dark brown, but the boundary with the
 327 Red soil is unclear.



328

329 Figure 5. Layer thickness and grain size analysis results for the sand layers and candidate source sands. a:
 330 Layer thickness changes of layers S1–7 on Line X. b: Horizontal changes in the mean diameters (Mz) and
 331 standard deviations (σ) of layers S1 and S2 on Line X. c: Vertical changes in the mean diameters, standard
 332 deviations, and skewness values (Sk_i) for layers S1 and S2. d: Scatter plot of the mean diameters and standard
 333 deviations in layers S1–10 on Line X, beach, river, aeolian (red soil), offshore sand. The offshore grain size

334 composition data are from Noda & Katayama (2011). e: Horizontal changes in the mean diameters standard
 335 deviations of layers S1 and S2 on Line Z.



336

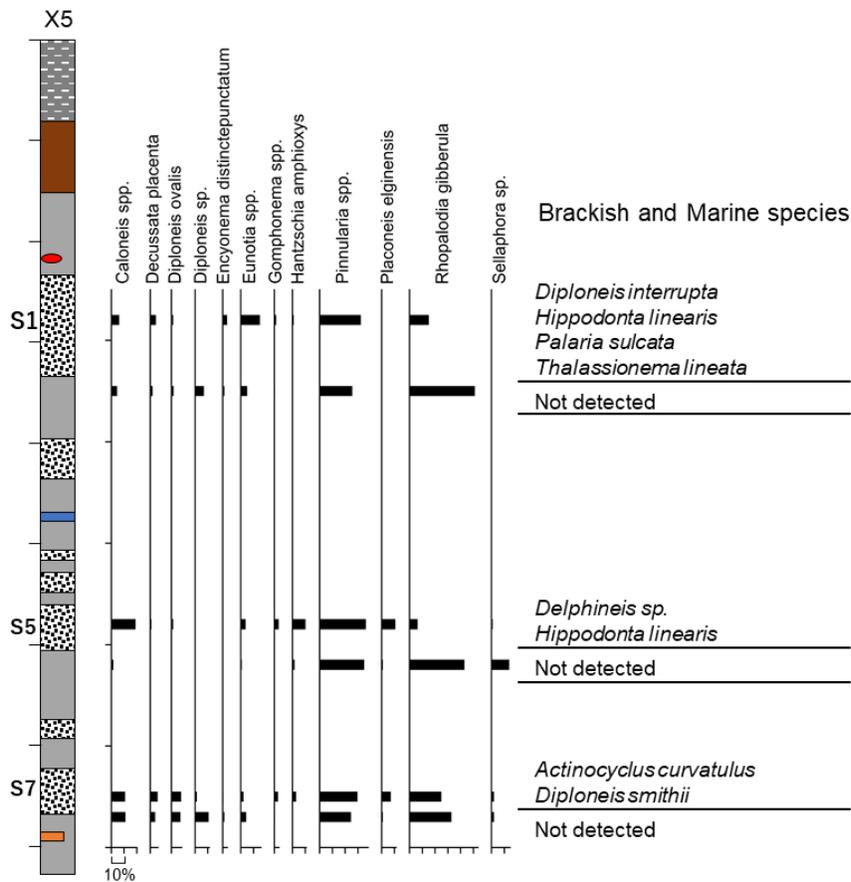
337 Figure 6. Results of mineralogy analysis using XRD for layers S1–10 at site X5 (10 samples), beach sand (3
 338 samples), river sand (1 sample) and Red soil (2 samples). The figures show the full view and the zoomed-out
 339 part. Each abbreviation indicates Am: Amphibole, Cl: Chlorite, Mu: Muscovite, Pl: Plagioclase, Qt: Quartz.

340 4.2 Sand layer features

341 The S1–10 layers consist of well- to moderately well-sorted fine to medium sands (Fig.
 342 5). These sand layers commonly exhibit single normal grading and sharp basal contacts. Some of
 343 the thick sand layers contained rip-up clasts of black peat (Fig. S1). The S1–8 layers exhibited
 344 modes of 2.5 phi and demonstrated very positive skewness. The grain size compositions of layers
 345 S9 and S10 consisted of 0.5–2.5 phi and moderately well sorted sand along X5 (Fig. S3). The
 346 mineral assemblages of layers S1–10 mainly consisted of quartz, feldspar, and chlorite, with low
 347 background values, which indicated few amorphous components (Fig. 6). The mineral
 348 assemblages do not vary significantly with the sample grain sizes or distances from the coastline.
 349 The mineral assemblages of the beach and river sands are similar to those of layers S1–10, but
 350 the river sands differ from layers S1–10 in that they have higher proportions of amphibole and
 351 lower proportions of other minerals.

352 Diatom assemblages in the peat and overlying sand layers are very similar, but the sand
 353 layers contained brackish and marine species (Fig. 7). The diatom assemblages in the peat layers
 354 are dominated by freshwater benthic or aerophilic species such as *Pinnularia*, *Caloneis*,
 355 *Rhopalodia*, and *Diploneis* genera. In addition to these species, the sand layer contains different

356 habitat species such as *Placoneis elginensis* (Chiba & Sawai, 2014), an indicator species for
 357 swampy wetlands, and epontic *Gomphonema* genera. In addition, the site is characterized by the
 358 presence of a few species that are known to inhabit tidal flats and the open sea.



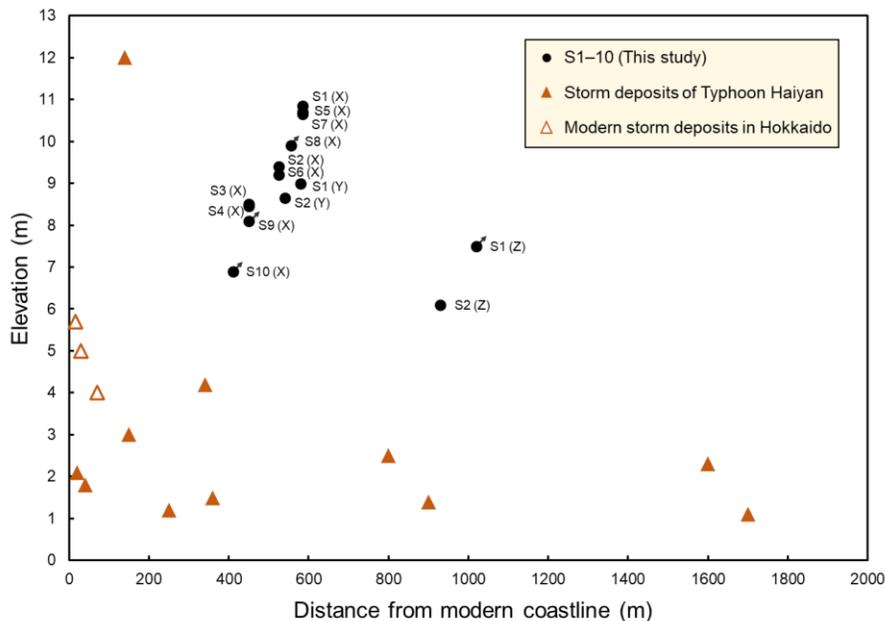
359

360 Figure 7. Results of diatom analysis on the sand layers and the below peat layers at X5. A list of brackish and
 361 marine species found in the analyzed samples is shown.

362 Correlations among the cores were estimated based on the tephtras and the thicknesses of
 363 the peat layers. The peat layer up to the Red soil is homogeneous and stable with very little
 364 clastic detritus; moreover, it has very low density, as indicated by the CT image (Fig. 3). In
 365 addition, there is a gradation of density in the peat above the fine volcanic ash layers, while the
 366 boundary is clear above the sand layers. Diatom assemblages supported a calm environment
 367 where no running water was involved, as benthic species were dominant, and diversity was low
 368 (Fig. 7). The sand layers inland from X9 are thin and shows slight normal grading of a few
 369 centimeters, which suggests that these sediments were transported by suspension and deposited
 370 without significant erosion. Moreover, the black peat layer exhibited a constant accumulation
 371 rate, as shown by the age–depth model (Fig. 4). Therefore, peat thickness was assumed to be
 372 proportional to the duration of the accumulation period. The S1–S7 layers are visible between
 373 Us-b and Ta-c2 until X6, but at X9, only three sand layers were visible (Fig. 2). However, the

374 CT image of the X9 core shows sand layers invisible by naked eye, corresponding to layers S2,
 375 S3, S4, and S6 at X5(Fig. 3), which were observed in X6 and X5. Therefore, the three sand
 376 layers identified up to X11 are S1, S5, and S7 from top to bottom. The depositional ages
 377 estimated from the relationship between the thickness of the peat layer and the stratigraphic
 378 position also supports that the three layers found inland are S1, S5, and S7 (Table S3). The sand
 379 layers below Ta-c2 were difficult to compare because of limited dating, but layers S8–10 were
 380 likely comparable because of the presence of three or more sand layers on the seaward and
 381 landward sides (Fig. 2).

382 Sand layers S1–7 exhibited sheet distributions as far as 450–600 m inland (up to 7–11 m
 383 asl) on Line X (Figs. 5 and 8). The thicknesses of layers S1–7 exhibited trends of gradual
 384 thinning inland; eventually, the sand layers became sandy peat or unrecognizable. We focused on
 385 layers S1 and S2 to clarify their distributions over a wide area of the Erimo coast (Figs. 1 and 2).
 386 For Line Y and Z, two gray sand layers (layers S1 and S2) were identified in the peat layers
 387 above B-Tm (Fig. 2). These sand layers show a thinning trend toward the inland. S1 of Line Z
 388 has a varied thickness trend due to artificial disturbance near the surface. Their distributions
 389 could be observed along the coast for several kilometers and also inland for over 1 km along
 390 Line Z, where the topography was relatively gentle. Layer S1 was distributed further inland than
 391 S2 along all of the survey lines (Fig. 1). Layers S1 and S2 indicated that the grain sizes become
 392 gradually fined inland from medium to fine sand on Line X (Fig. 5). The grain size composition
 393 of Line Z tend to be finer inland, as in Line X, but slightly coarser, reflecting the coarser-grained
 394 river sand (Fig. 5).



395
 396 Figure 8. Distribution heights and distances from the present coastline for layers S1–10 on each survey line.
 397 The black plots with arrows indicate the sand layers that are potentially distributed farther inland due to

398 unconfirmed distribution limits. The letters in parentheses indicate the names of the survey lines. The filled
399 and white triangles indicate the maximum distributions of the storm deposits of Typhoon Haiyan on the
400 central Philippine coast and the modern storm deposits reported in Hokkaido, respectively (Brill et al., 2016;
401 Soria et al., 2017, 2018; Nishimura & Miyaji, 1996; Ganzei et al., 2010; Shigeno & Nanayama, 2016; Chiba &
402 Nishimura, 2018; Switzer et al., 2020). The data from Typhoon Haiyan are used for comparison as the largest
403 storm, not as the one that actually affected Hokkaido.

404 4.3 Formation of anomalous sand layers

405 Layers S1–10 were observed in the peat with sharp boundaries with rip-up clasts (Fig.
406 S1). The sand layers show single normal grading structures and landward grain size fining (Fig.
407 5). The grain size compositions of layers S1–10 were similar to those of the beach and dune
408 sands (i.g., fine-grained offshore sands and coarse-grained river sands). The narrow grain size
409 range of the fining trend along the survey lines can be interpreted as being due to the fact that the
410 major sand source was well-sorted beach sand. However, because the sand layers tend to have
411 poorer sorting than the beach sands, it is assumed that the sand layers were transported by using
412 sand from near the estuary and offshore. The mineral assemblages also support the fact that
413 beach sand was the major source of layers S1–10 (Fig. 6). Glauconite grains and sessile
414 organisms, which are characteristic of offshore sediments (Noda & TuZino, 2010), were not
415 detected in layers S1–10, which suggest that the contributions from offshore sediments are small.
416 The sand layers were supplied from the beach sand and were transported from the sea toward the
417 inland area, as demonstrated by the thinning and fining trends. This observation is supported by
418 the inclusion of brackish and marine diatom valves in the sand layers, which are not present in
419 the peat layer (Fig. 7). In addition, it is unlikely that layers S1–10 were transported by river
420 floods because the sediments of the nearby rivers consist of gravel (Fig. 5). The Red soil is
421 distinguished from layers S1–10 by poorly sorted sands owing to the presence of coarse-grained,
422 low specific density amorphous components such as pumice and fine-grained detritus (Fig. 6).
423 The repetition of sharp contrasts between peat and sand layers with rip-up clasts could have been
424 formed only by abrupt events such as extreme waves.

425 Tsunamis and storm surges are known to form sand layers beyond the back barrier (e.g.,
426 Morton et al., 2007; Takashimizu et al., 2012; Brill et al., 2016). In general, although extreme
427 waves from storms show high inundation heights, the distributions of sand layers are often
428 limited around beach areas (Morton et al., 2007; Watanabe et al., 2018). Storm surges also
429 consist of gradual rises in the water level, which inundate beaches from the lowest points. Thus,
430 storm deposit distributions are often scattered and lobe-shaped (Nishimura & Miyaji, 1996; Brill
431 et al., 2016; Chiba & Nishimura, 2018). Storm deposits have also been reported in Hokkaido,
432 and their distributions were reported up to 4–6 m asl and a few tens of meters from the shoreline
433 (Nishimura & Miyaji, 1996; Ganzei et al., 2010; Shigeno & Nanayama, 2016; Chiba &
434 Nishimura, 2018; Switzer et al., 2020). Additionally, tsunamis result in erosion and inland
435 transportation of large sediment volumes in short time intervals (Morton et al., 2007; Szczuciński
436 et al., 2012; Goto et al., 2014). As a result of erosion, the contacts between the transported sand

437 layers and the background sediments are very sharp and often include rip-up clasts (Szczuciński
438 et al., 2012; Takashimizu et al., 2012). Inundation flows are slowed by flooding on land; grains
439 with higher specific gravity settle and are eventually recorded in the sediment as grading
440 structures (Jaffe et al., 2012; Yoshii et al., 2017). The amounts of transported sediments are
441 limited owing to decreasing tsunami flow velocities and inundation depths; thus, sheet-like
442 distributions and gradual thinning are observed (Szczuciński et al., 2012; Takashimizu et al.,
443 2012; Goto et al., 2014).

444 By comparing the features of the modern tsunami and storm deposits described above, it
445 was shown that the sand layers in this area are tsunami deposits. Layers S1–10 exhibited normal
446 grading and sharp basal contacts with rip-up clasts; these sedimentary structures indicate
447 erosional flow and decreased velocities of single flows. Their distributions extended to several
448 hundred meters inland from the beach and they are spread over wide areas along the shoreline. In
449 addition, the elevations of the sand layer distributions are at least 7–11 m asl (Fig. 8), which is
450 significantly higher than those of modern storm deposits (Nishimura & Miyaji, 1996; Ganzei et
451 al., 2010; Brill et al., 2016; Shigeno & Nanayama, 2016; Soria et al., 2017, 2018; Chiba &
452 Nishimura, 2018; Switzer et al., 2020). The elevations of the deposits are > 5 m asl even when
453 considering the relative sea levels of the past 4000 years based on hydro-isostasy and crustal
454 uplift (up to + 2.1 m at a mean higher high water level: Koike & Machida, 2001; Okuno et al.,
455 2014). For the past 4000 years, the Erimo coast has been limited to the area located ~100 m from
456 the present coastline with elevations under 2 m asl. Therefore, the distributions of layers S1–10
457 along Line X are estimated to be present for more than 300–500 m from the past coastline.
458 Considering that the distributions of layers S1–10 cannot be explained by anomalous storms and
459 that some sedimentary structures are common in tsunami deposits, these are interpreted as
460 anomalous sand layers that were deposited by high tsunami waves.

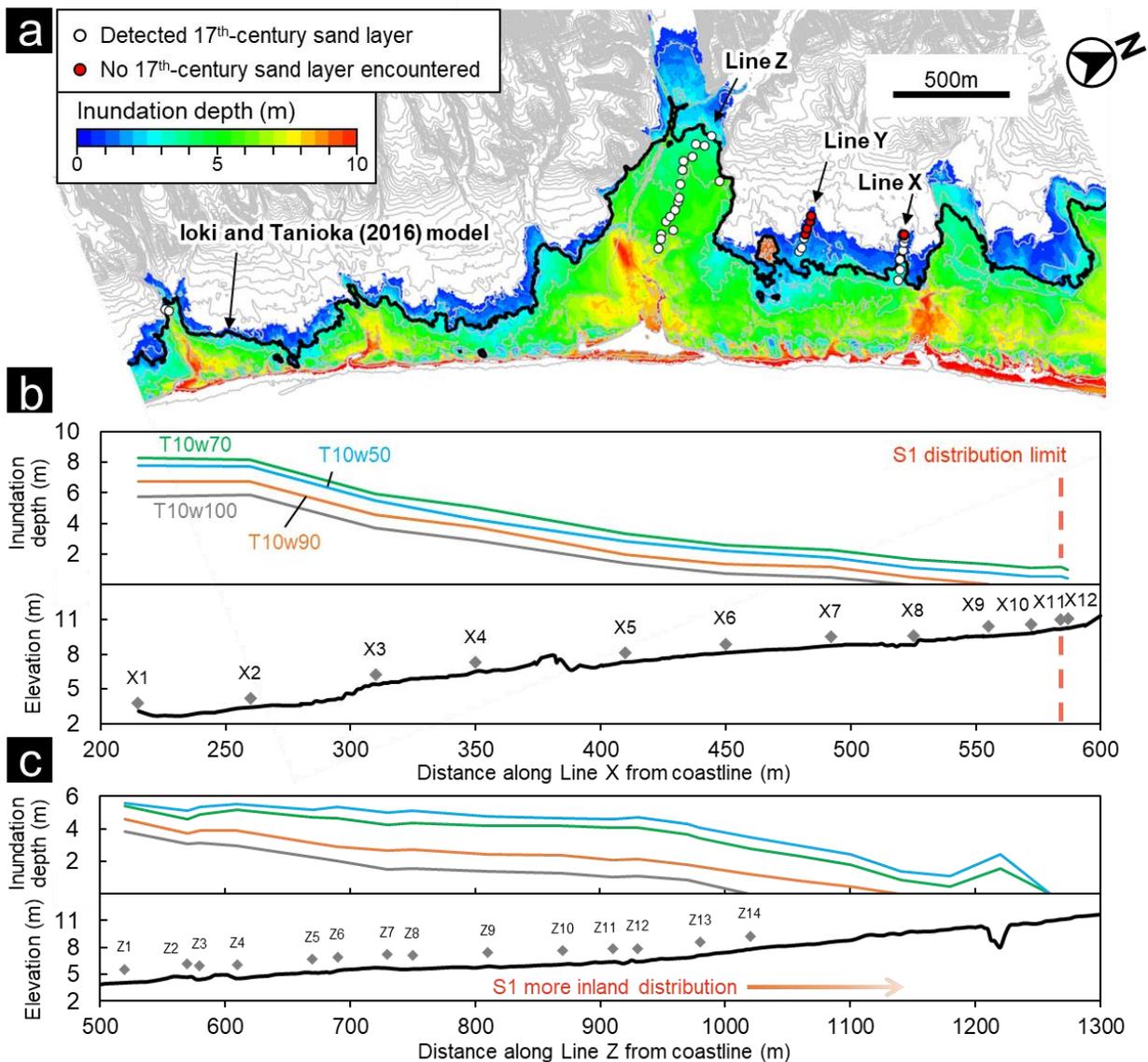
461 **5 Numerical simulations**

462 We used numerical simulations to reconstruct tsunami inundation consistent with
463 observed distributions of tsunami deposits along the eastern Hokkaido coast. Tsunamis generated
464 by Mw~8 earthquakes in the Kuril Trench are known to have recurrence periods of several
465 decades (e.g., the 1952 and 2003 Tokachi-oki earthquakes). Given that no sandy deposits were
466 transported into the back barrier by these tsunamis (Nishimura et al., 2004), it is unlikely that the
467 tsunamis generated by Mw~8 earthquakes transported the S1–10 tsunami deposits. A source
468 rupture model based on geological data has been proposed for 17th-century earthquakes
469 (Nanayama et al., 2003; Satake et al., 2008; Ioki & Tanioka, 2016). The T10N5S25 model (Mw
470 8.8, where T10 denotes 10 m slip on the Tokachi-oki segment, N5 denotes 5 m slip on the
471 Nemuro segment, and S25 denotes 25 m slip on the shallow part segment; Fig. 1) produces an
472 inundation area that covers the distribution of the tsunami deposits extending from the Tokachi
473 to Nemuro regions (Ioki & Tanioka, 2016). In addition to the distribution of tsunami deposits,
474 these source rupture models are required to meet the constraint that tsunami wave heights should

475 not exceed 3 m in the Sanriku region because there is no description of an earthquake like this in
476 the historical documents (Satake et al., 2008). However, this requirement has not yet been
477 appropriately satisfied.

478 We tested whether the T10N5S25 model can be applied to 17th-century tsunami deposits
479 in the Erimo area, which are further to the west. The resulting inundation area did not cover the
480 entire distribution of layer S1 (Fig. 9). To avoid affecting the tsunami simulation results in the
481 Tokachi to Nemuro regions, we attempted to adjust the fault model to expand the tsunami
482 inundation area in the Erimo area. If, in the Tokachi-oki segment, the rupture model parameter is
483 wider or the slip amount is increased, the maximum wave height along the Sanriku coast will be
484 higher. Therefore, we attempted to reproduce the inundation area in the Erimo area by narrowing
485 the fault width, as it is known that the wavelengths affect inundation extents depending on the
486 topography (Satake et al., 2013). The fault widths were set at 10 km intervals that ranged from

487 40 to 90 km (T10w40–T10w90) along the Tokachi-oki segment; N5 and S25 are constant for all
 488 conditions. The magnitudes of these rupture models were Mw 8.7–8.8.



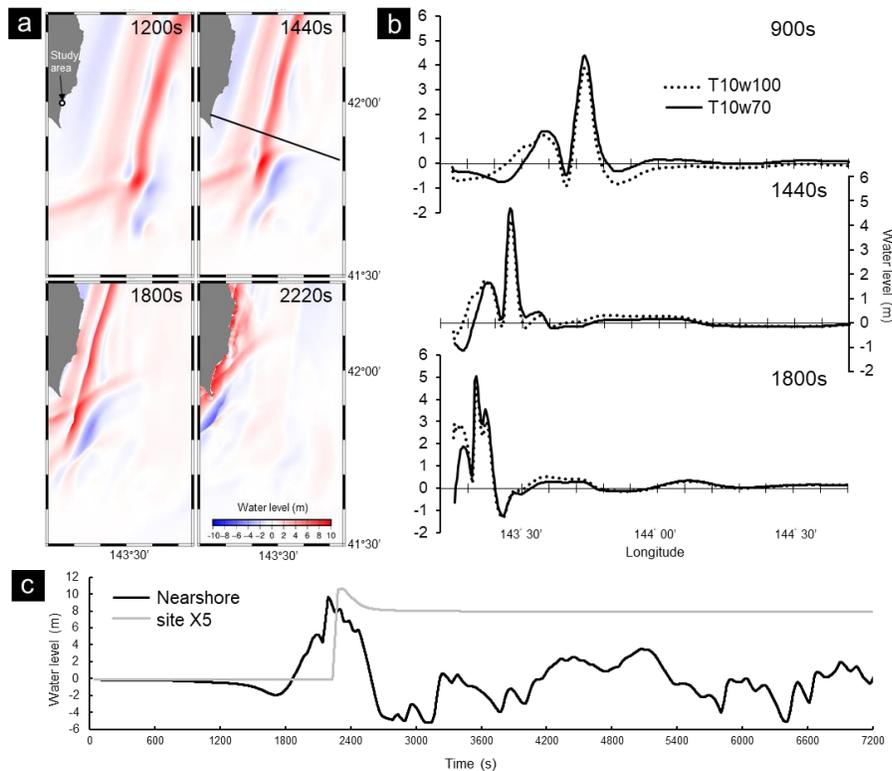
489
 490 Figure 9. Simulation results of tsunami run-ups on the Erimo coast. a: Tsunami inundation area using the loki
 491 & Tanioka (2016) model (black line) and T10w70 model (colors). b: Tsunami inundation depth profiles for
 492 each fault width value and the topographic profile on Line X. c: Tsunami inundation depth profiles and the
 493 topographic profile on Line Z. T10w100 is a conventional model (Ioki & Tanioka, 2016)

494 The T10w50–T10w70 models generated inundation areas that covered the entire
 495 distribution of layer S1 (Fig. 9). These models also changed the energy directivity to the Sanriku
 496 region, such that the maximum wave heights on the Sanriku coastline decreased as the fault
 497 widths narrowed (Fig. S5). The maximum wave heights along the coastline from the Tokachi to
 498 Nemuro regions also exhibited slight decreases when narrow fault width conditions were used.

499 The T10w70 model is more reasonable because it reproduces an inundation area that covers a
500 sufficient amount of tsunami deposits and minimizes the wave height changes from the Tokachi
501 to Nemuro regions.

502 In these numerical simulations, we used a simple modification method that consisted of
503 changing the fault width of the Tokachi-oki segments, and it can be explained that the inundation
504 area extended wider for several reasons. The crustal deformation of this model exhibits a
505 subsidence level of ~15 cm around Erimo (Fig. S6), while the model proposed by Ioki &
506 Tanioka (2016) exhibited a subsidence level of ~40 cm, which does not explain the differences in
507 the inundation extents. The locally high wave heights in the Erimo area likely reflected the
508 shortening of the wave periods that was caused by the narrowing of the fault width. When the
509 fault width of T10 is 100 km, there is a certain interval between the first and second waves, but
510 the T10w70 model shows that the first wave propagates slightly more slowly and excites the
511 second wave (Fig. 10). In addition, in the T10w70 model, the third wave appears earlier and
512 merges with the wave that is trapped by the shallow bathymetry on the Cape Erimo extension
513 and expands the inundation area subsequent to the second wave. The landward side of X5 was
514 inundated only by the first wave (Fig. 10). This is consistent with the fact that the observed
515 sedimentary structure of the sand layers consisted of single normal grading (Fig. 5). The T10w70
516 model is nearly equal to Mw 8.8, which is the same scale as the conventional model. This means
517 that the model based on the tsunami deposits from the Tokachi to Nemuro region can reproduce
518 the tsunami deposit distributions found on the eastern Erimo coast without major changes. Since
519 it is known that actual rupture zones are not rectangular but are heterogeneous (e.g., Liu et al.,
520 2013), it should be noted that the model does not necessarily recover the detailed parameters of
521 the source fault. By taking this information into account, a simulation could reproduce the
522 tsunami deposit distributions in each region with greater accuracy. On the other hand, it should
523 be noted that the distributions of the sand layers contain uncertainties because tsunami deposits
524 are known poor preservation by postdepositional processes (Richmond et al., 2012; Szczuciński,
525 2020), and tsunami inundations limited sand can only be transported (Abe et al., 2012). To
526 reduce such uncertainties, it is necessary to evaluate the tsunami inundation areas by using

527 chemical and biomarker methods that are independent of the sand layers (e.g., Szczuciński et al.,
528 2016; Chagué-Goff et al., 2017).



529

530 Figure 10. Tsunami waveforms in the time series from the numerical simulations of the 17th-century tsunami.
531 a: Temporal changes in water level for the T10w70 model. Snapshots taken at 1200, 1440, 1800, and 2220
532 seconds after the earthquake. The black line indicates the survey line shown in Figure 10b. b: The wave
533 profiles on the black line shown in Fig. 10a at 900, 1440, and 1800 seconds for the T10w100 and T10w70
534 models. c: Time series of the tsunami waveforms of the T10w70 model at site X5 and nearshore of Line X.
535 The DEM elevation at site X5 is 7.9 m asl.

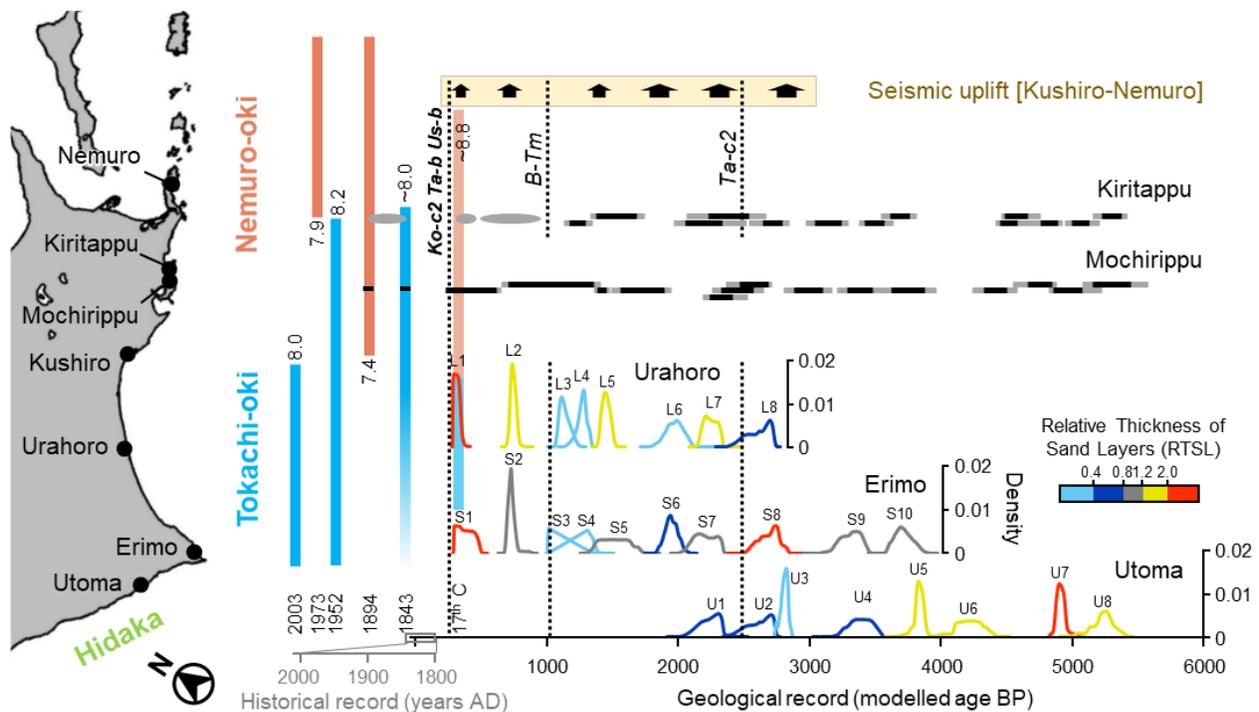
536

537 6 Correlation of tsunami deposits

538 Regional comparison of tsunami events was made by assessing the consistency of
539 tsunami ages in the surrounding areas. Tsunami ages have been reported in the Tokachi
540 (Urahoro) and Hidaka (Utoma) regions (Ishizawa et al., 2017; Nakanishi et al., 2020a). The
541 agreements of the probability density functions of events that were obtained by the sequencing
542 model for each region were calculated using the overlapping coefficient (OVL: Inman &

543 Bradley, 1989; Hutchinson & Clague, 2017: Table 2). Hutchinson & Clague (2017) adopted
 544 critical threshold values for probable coeval events when the OVLs were > 0.25 .

545 We compared the depositional ages of the sand layers with those reported for Urahoro
 546 and Utoma, which were measured using the same ^{14}C dating method (Fig. 11). In the Erimo and
 547 Urahoro areas facing the Tokachi-oki segment, the depositional ages were comparable in all of
 548 the observed sand layers, and the OVLs ranged from 0.39 to 0.78 (Table 2). In particular, S2 and
 549 L2 (layer 2 in Urahoro: Ishizawa et al., 2017) showed a narrow range from ~ 800 to 700 BP, with
 550 an error of 20 years. The results mean that Urahoro and Erimo recorded the same numbers of
 551 large tsunamis that occurred at the same times during the past 3000 years. In the Utoma area,
 552 which is located on the southern Hidaka coast, the depositional ages are comparable in the four
 553 sand layers, except for U3 (sand layer interpreted as a tsunami deposit in Utoma: Nakanishi et
 554 al., 2020a), and the OVLs range from 0.26 to 0.81 (Table 2). The geological record in Utoma is
 555 currently limited to the period before 2,000 BP, but tsunamis with recurrence intervals of ~ 430
 556 years are most likely to have been generated by the common $M_w > 8$ earthquakes in the Tokachi
 557 region. However, it is necessary to consider the possibility that tsunamis from different sources,
 558 such as the U3 event in Utoma, are also included in the data.



559
 560 Figure 11. Depositional ages of sand layers in the Hidaka and Tokachi regions (Ishizawa et al., 2017; Nakanishi
 561 et al., 2020a). The depositional ages are presented as probability density functions. The color scale of the
 562 histograms indicates the relative thicknesses of the sand layers, which were normalized using the median
 563 values and exclude any outliers. Outliers were defined as exceeding the interquartile range by 1.5 times
 564 based on the upper quartile. Gray and black bars indicate the estimated ages (2σ and 1σ , respectively) of the

565 tsunamis reported in the Mochirippu and Kiritappu areas (Sawai et al., 2009). Black arrows indicate the age
 566 ranges of the seismic uplifts in the Kushiro to Nemuro regions (Kelsey et al., 2006). Blue and red color bars
 567 indicate the fault lengths of earthquakes that occurred off the Tokachi and Nemuro coasts, respectively
 568 (Satake, 2015). The numbers beside each color bar indicate the moment magnitudes of the earthquakes.

569 Table 2. The depositional ages of the compared events obtained by the sequence model and the overlapping
 570 coefficients (OVL).

Urahoro*1	2 σ range modeled age (BP)		Erimo	2 σ range modeled age (BP)		OVL
L1	350	283	S1	474	287	0.42
L2	786	704	S2	796	680	0.72
L3	1214	1070	S3	1279	1003	0.53
L4	1340	1184	S4	1372	1068	0.63
L5	1511	1400	S5	1704	1348	0.39
L6	2097	1845	S6	2043	1851	0.78
L7	2336	2166	S7	2340	2069	0.65
L8	2735	2442	S8	2825	2546	0.63
Ave.						0.59

Erimo	2 σ range modeled age (BP)		Utoma*2	2 σ range modeled age (BP)		OVL
S7	2340	2069	U1	2356	2071	0.81
S8	2825	2546	U2	2743	2438	0.67
S9	3449	3170	U4	3542	3234	0.66
S10	3846	3590	U5	3905	3735	0.26
Ave.						0.60

Urahoro	2 σ range modeled age (BP)		Utoma	2 σ range modeled age (BP)		OVL
L7	2336	2166	U1	2356	2071	0.76
L8	2735	2442	U2	2743	2438	0.95
Ave.						0.86

*1: Ishizawa et al. (2017)

*2: Nakanishi et al. (2020a)

571

572 The relative thicknesses of the sand layers (RTSL), which are normalized by the median
 573 values of all sand layers within the core, suggest that Tokachi and Hidaka regions experienced
 574 tsunamis of similar magnitude (Fig. 11). In Erimo and Urahoro, it is very interesting to match the
 575 patterns of the high and low values, except for layer S8. This good agreement between the two
 576 regions, despite the 100 km distance between them, may support the assumption that both
 577 regions experienced tsunamis of similar magnitudes. This is because modern tsunami deposit
 578 studies infer that the thicknesses of tsunami deposits are most sensitive to the inundation depths
 579 (Goto et al., 2014; Naruse & Abe, 2017). However, it should be noted that the microtopography,
 580 accommodation space, sediment source, and postdepositional changes all significantly influence
 581 the thicknesses of tsunami deposits (Szczuciński, 2020). Actually, the RTSL in Utoma is thick in

582 the sand layers that were deposited at approximately 5000 BP, which was due to the influence of
583 the sea level highstand in the mid-Holocene (Nakanishi et al., 2020a).

584 Several tsunami deposits can be correlated along more than 150 km of the coast, which
585 suggests that they were generated by tsunamis that resulted from $M_w > 8$ earthquakes, with fault
586 lengths of several hundred kilometers. Therefore, few of the layer S1–10 events are expected to
587 be correlated with the geological evidence identified from the Kushiro to Nemuro regions for
588 paleo-earthquakes (Nanayama et al., 2003; Sawai et al., 2004, 2009; Kelsey et al., 2006). The
589 tsunami ages and seismic crustal movements in Kiritappu and Mochirippu have been estimated
590 based on ^{14}C dating (Sawai et al., 2009). However, direct correlations are difficult because of the
591 differences in the sample materials, dating methods, and dating depth points. The estimated ages
592 of the seismic uplifts in the Nemuro to Kushiro regions (Kelsey et al., 2006) appear to correlate
593 with the depositional ages of the relatively thick sand layers found in the Hidaka and Tokachi
594 regions (e.g., layers S1, S5, and S7). The distributions of these layers are most extensive inland
595 (Fig. 2). Thus, these tsunamis might have been generated by higher-magnitude earthquakes or by
596 wider rupture zones when compared with other events from the Tokachi to Hidaka regions.

597 The observed earthquakes along the Japan Trench caused 3–4 m tsunamis in the Erimo
598 area, and may have caused larger tsunamis in prehistoric periods (Nakanishi et al., 2020a). The
599 1611 Keicho tsunami is considered to be the largest tsunami among those of historical
600 earthquakes recorded in documents of the Sanriku region. Although it is not known how high
601 this tsunami reached along the Hokkaido coast (Sawai, 2020), Tetsuka et al. (2020) simulated a
602 tsunamis generated by $M_w 8.8$ earthquakes off the Sanriku coast and showed that a large tsunami
603 would not reach the Hokkaido coast. However, some paleo-tsunami deposits in the Sanriku
604 region show similar ages to those in the Erimo area (e.g., Ishizawa et al., 2022). Therefore, some
605 for events there is a need to consider a wave source in (or across) the Japan Trench.

606 The Nemuro and Tokachi-Hidaka regions likely experienced different paleotsunami
607 histories based on the numbers of sand layers recorded at well-studied sites. Thus, both regions
608 likely experienced at least some different paleotsunamis. Cataloging megathrust earthquakes
609 shows that the rupture segmentation and cyclicity can be complex (Philibosian & Meltzner,
610 2020), and the Kuril Trench is no exception, which suggests that the regions that are located
611 across the currently known segments have different rupture histories. Information on the
612 earthquake segmentation in this trench and their recurrence intervals will better explain the
613 megathrust earthquake mechanisms and hazard assessments.

614 **7 Conclusions**

615 We identified ten anomalous sand layers in the coastal wetland in Erimo, Hokkaido,
616 Japan. These sand layers share similarities with beach sands and can be distinguished from
617 aeolian and river sands. Moreover, they share common features with modern tsunami deposits,
618 such as normal grading structures and sharp basal contacts with rip-up clasts. The sand layers are

619 distributed at distances of several hundred meters from the shoreline (up to 7–11 m asl), which
620 cannot be explained by extreme storms or tsunamis caused by Mw~8 earthquakes. Numerical
621 simulations were performed for a 17th-century tsunami to examine whether the distributions of
622 the 17th-century sand layers could be reproduced by the Mw 8.8 earthquake model. The results
623 showed that tsunami inundation by conventional models was insufficient to cover the distribution
624 of tsunami deposits. We modified the fault width slightly without changing the tsunami wave
625 heights in the eastern area, and found that it was possible to reproduce the distribution of tsunami
626 deposits in the Erimo area. The correlations of tsunami events in the past 4000 years imply
627 widespread deposition of sand by large tsunamis that impacted the Tokachi and Hidaka regions
628 and indicate recurrence intervals of 110–620 years. The chronological correlations between most
629 of the tsunamis at these sites indicate that the paleotsunamis reported in eastern Hokkaido
630 reached the Hidaka region. On the other hand, a detailed study of the tsunami ages also revealed
631 the differences between the regions that face the Tokachi-oki and Nemuro-oki segments,
632 suggesting the diversity of the Kuril earthquakes every few hundred years. By extending the
633 correlations of the paleotsunamis from Hidaka to Tokachi to the past 4000 years, the spatial
634 extents and recurrence intervals of large tsunamis in the regions that face the Tokachi-oki
635 segment have been significantly updated. The common history of the tsunamis in the area facing
636 the Tokachi-oki segment that are revealed by this study will provide a framework for more
637 comprehensive correlations of the Mw>8 earthquakes in the entire Kuril Trench.

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645 **Data Availability Statement**

646 The new ¹⁴C datasets for this research are provided in Table 1 in this paper. We used the tsunami
647 calculation code JAGURS (Baba et al., 2015, 2017) at <https://doi.org/10.5281/zenodo.3737816>.
648 Generic mapping tools were used to draw the figures (Wessel et al., 2013).

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650 **References**

651 Abe, T., Goto, & K., Sugawara, D. (2012). Relationship between the maximum extent of tsunami sand and the inundation limit of
652 the 2011 Tohoku-oki tsunami on the Sendai Plain, Japan. *Sedimentary Geology*, 282, 142–150.
653 <https://doi.org/10.1016/j.sedgeo.2012.05.004>

- 654 Baba, T., Takahashi, N., Kaneda, Y., Ando, K., Matsuoka, D., & Kato, T. (2015). Parallel Implementation of Dispersive Tsunami
655 Wave Modeling with a Nesting Algorithm for the 2011 Tohoku Tsunami. *Pure and Applied Geophysics*, 172(12), 3455–3472.
656 <https://doi.org/10.1007/s00024-015-1049-2>
- 657 Baba, T., Allgeyer, S., Hossen, J., Cummins, P. R., Tsushima, H., Imai, K., et al. (2017). Accurate numerical simulation of the
658 far-field tsunami caused by the 2011 Tohoku earthquake, including the effects of Boussinesq dispersion, seawater density
659 stratification, elastic loading, and gravitational potential change. *Ocean Modelling*, 111, 46–54.
660 <https://doi.org/10.1016/j.ocemod.2017.01.002>
- 661 Brill, D., May, S. M., Engel, M., Reyes, M., Pint, A., Opitz, S., et al. (2016). Typhoon Haiyan’s sedimentary record in coastal
662 environments of the Philippines and its palaeotempestological implications. *Natural Hazards and Earth System Sciences*, 16(12),
663 2799–2822. <https://doi.org/10.5194/nhess-16-2799-2016>
- 664 Bronk Ramsey, C. (2008). Deposition models for chronological records. *Quaternary Science Reviews*, 27(1), 42–60.
665 <https://doi.org/10.1016/j.quascirev.2007.01.019>
- 666 Bronk Ramsey, C. (2009a). Bayesian Analysis of Radiocarbon Dates. *Radiocarbon*, 51(1), 337–360.
667 <https://doi.org/10.1017/S0033822200033865>
- 668 Bronk Ramsey, C. (2009b). Dealing with Outliers and Offsets in Radiocarbon Dating. *Radiocarbon*, 51(3), 1023–1045.
669 <https://doi.org/10.1017/S0033822200034093>
- 670 Chagué-Goff, C., Szczuciński, W., & Shinozaki, T. (2017). Applications of geochemistry in tsunami research: A review. *Earth-
671 Science Reviews*, 165, 203–244. <https://doi.org/10.1016/j.earscirev.2016.12.003>
- 672 Chiba, T., & Sawai, Y. (2014). Reexamination and updating of diatom species for paleoenvironmental reconstructions. *Diatom*,
673 30, 17–30 (in Japanese, with English abstract).
- 674 Chiba, T., & Nishimura, Y. (2018). Depositional processes of the storm surge deposits due to Typhoon 1523 in Horokayanto,
675 Hokkaido, Japan. *The Quaternary Research (Daiyonki kenkyu)*, 57, 53–63 (in Japanese, with English abstract).
- 676 Folk, R. L., & Ward, W. C. (1957). Brazos river bar: A study in the significance of grain size parameters. *Journal of Sedimentary
677 Petrology*, 27(1), 3–26. <https://doi.org/10.1306/74D70646-2B21-11D7-8648000102C1865D>
- 678 Ganzei, L. A., Razzhigaeva, N. G., Harlamov, A. A., & Ivel’skaya, T. N. (2010). Extreme storms in 2006–2007 on Shikotan
679 Island and their impact on the coastal relief and deposits. *Oceanology*, 50(3), 425–434.
680 <https://doi.org/10.1134/S0001437010030112>
- 681 Goldfinger, C., Ikeda, Y., Yeats, R. S., & Ren, J. (2013). Superquakes and Supercycles. *Seismological Research Letters*, 84(1),
682 24–32. <https://doi.org/10.1785/0220110135>
- 683 Goto, K., Hashimoto, K., Sugawara, D., Yanagisawa, H., & Abe, T. (2014). Spatial thickness variability of the 2011 Tohoku-oki
684 tsunami deposits along the coastline of Sendai Bay. *Marine Geology*, 358, 38–48. <https://doi.org/10.1016/j.margeo.2013.12.015>
- 685 Hashimoto, C., Noda, A., Sagiya, T., & Matsu’ura, M. (2009). Interplate seismogenic zones along the Kuril–Japan trench
686 inferred from GPS data inversion. *Nature Geoscience*, 2(2), 141–144. <https://doi.org/10.1038/ngeo421>
- 687 Hutchinson, I., & Clague, J. (2017). Were they all giants? Perspectives on late Holocene plate-boundary earthquakes at the
688 northern end of the Cascadia subduction zone. *Quaternary Science Reviews*, 169, 29–49.
689 <https://doi.org/10.1016/j.quascirev.2017.05.015>
- 690 Inman, H. F., & E. L. Bradly Jr. (1989). The overlapping coefficient as a measure of agreement between probability distributions
691 and point estimation of the overlap of two normal densities. *Communications in Statistics - Theory and Methods*, 18(10), 3851–
692 3874. <https://doi.org/10.1080/03610928908830127>
- 693 Ioki, K., & Tanioka, Y. (2016). Re-estimated fault model of the 17th century great earthquake off Hokkaido using tsunami
694 deposit data. *Earth and Planetary Science Letters*, 433, 133–138. <https://doi.org/10.1016/j.epsl.2015.10.009>

- 695 Ishizawa, T., Goto, K., Yokoyama, Y., Miyairi, Y., Sawada, C., Nishimura, Y., & Sugawara, D. (2017). Sequential radiocarbon
696 measurement of bulk peat for high-precision dating of tsunami deposits. *Quaternary Geochronology*, 41, 202–210.
697 <https://doi.org/10.1016/j.quageo.2017.05.003>
- 698 Ishizawa, T., Goto, K., Yokoyama, Y., & Goff, J. (2020). Dating tsunami deposits: Present knowledge and challenges. *Earth-
699 Science Reviews*, 200, 102971. <https://doi.org/10.1016/j.earscirev.2019.10297>
- 700 Ishizawa, T., Goto, K., Nishimura, Y., Miyairi, Y., Sawada, C., & Yokoyama, Y. (2022). Paleotsunami history along the northern
701 Japan trench based on sequential dating of the continuous geological record potentially inundated only by large tsunamis.
702 *Quaternary Science Reviews*, 279, 107381. <https://doi.org/10.1016/j.quascirev.2022.107381>
- 703 Jaffe, B. E., Goto, K., Sugawara, D., Richmond, B. M., Fujino, S., & Nishimura, Y. (2012). Flow speed estimated by inverse
704 modeling of sandy tsunami deposits: results from the 11 March 2011 tsunami on the coastal plain near the Sendai Airport,
705 Honshu, Japan. *Sedimentary Geology*, 282, 90–109. <https://doi.org/10.1016/j.sedgeo.2012.09.002>
- 706 Jagodziński, R., Sternal, B., Szczuciński, W., Chagué-Goff, C., & Sugawara, D. (2012). Heavy minerals in the 2011 Tohoku-oki
707 tsunami deposits—insights into sediment sources and hydrodynamics. *Sedimentary Geology*, 282, 57–64.
708 <https://doi.org/10.1016/j.sedgeo.2012.07.015>
- 709 Kelsey, H., Satake, K., Sawai, Y., Sherrod, B., Shimokawa, K., & Shishikura, M. (2006). Recurrence of postseismic coastal
710 uplift, Kuril subduction zone, Japan. *Geophysical Research Letters*, 33(13), L13315. <https://doi.org/10.1029/2006GL026052>
- 711 Koike, K. & Machida, H. (2001). *Atlas of Quaternary Marine Terraces in the Japanese Islands*: Tokyo, University of Tokyo Press
712 (in Japanese).
- 713 Lienkaemper, J. J., & Ramsey, C. B. (2009). OxCal: Versatile Tool for Developing Paleoseismic Chronologies –A Primer.
714 *Seismological Research Letters*, 80(3), 431–434. <https://doi.org/10.1785/gssrl.80.3.431>
- 715 Liu, X., Zhao, D., & Li, S. (2013). Seismic heterogeneity and anisotropy of the southern Kuril arc: insight into megathrust
716 earthquakes. *Geophysical Journal International*, 194(2), 1069–1090. <https://doi.org/10.1093/gji/ggt150>
- 717 Morton, R. A., Gelfenbaum, G., & Jaffe, B. E. (2007). Physical criteria for distinguishing sandy tsunami and storm deposits using
718 modern examples. *Sedimentary Geology*, 200(3–4), 184–207. <https://doi.org/10.1016/j.sedgeo.2007.01.003>
- 719 Nakamura, Y., Nishimura, Y., & Putra, P. S. (2012). Local variation of inundation, sedimentary characteristics, and mineral
720 assemblages of the 2011 Tohoku-oki tsunami on the Misawa coast, Aomori, Japan. *Sedimentary Geology*, 282, 216–227.
721 <https://doi.org/10.1016/j.sedgeo.2012.06.003>
- 722 Nakanishi, R., Okamura, S., Yokoyama, Y., Miyairi, Y., Sagayama, T., & Ashi, J. (2020a). Holocene tsunami, storm, and relative
723 sea level records obtained from the southern Hidaka coast, Hokkaido, Japan. *Quaternary Science Reviews*, 250, 106678.
724 <https://doi.org/10.1016/j.quascirev.2020.106678>
- 725 Nakanishi, R., Ashi, J., & Okamura, S. (2020b). A dataset for distribution and characteristics of Holocene pyroclastic fall
726 deposits along the Pacific coasts in western Hokkaido, Japan. *Data in Brief*, 33, 106565.
727 <https://doi.org/10.1016/j.dib.2020.106565>
- 728 Nanayama, F., Satake, K., Furukawa, R., Shimokawa, K., Atwater, B. F., Shigeno, K., & Yamaki, S. (2003). Unusually large
729 earthquakes inferred from tsunami deposits along the Kuril trench. *Nature*, 424(6949), 660–663.
730 <https://doi.org/10.1038/nature01864>
- 731 Nanayama, F., Furukawa, R., Shigeno, K., Makino, A., Soeda, Y., & Igarashi, Y. (2007). Nine unusually large tsunami deposits
732 from the past 4000 years at Kiritappu marsh along the southern Kuril Trench. *Sedimentary Geology*, 200(3–4), 275–294.
733 <https://doi.org/10.1016/j.sedgeo.2007.01.008>

- 734 Naruse, H., & Abe, T. (2017). Inverse Tsunami Flow Modeling Including Nonequilibrium Sediment Transport, With Application
735 to Deposits From the 2011 Tohoku-Oki Tsunami. *Journal of Geophysical Research: Earth Surface*, 122(11), 2159–2182.
736 <https://doi.org/10.1002/2017JF004226>
- 737 Nelson, A. R., Kelsey, H. M., & Witter, R. C. (2006). Great earthquakes of variable magnitude at the Cascadia subduction zone.
738 *Quaternary Research*, 65(3), 354–365. <https://doi.org/10.1016/j.yqres.2006.02.009>
- 739 Nishimura, Y., & Miyaji, N. (1996). Investigations of storm deposits caused by a typhoon of September 19th, 1994. *Bulletin of*
740 *the Natural Disaster Science Data Center, Hokkaido*, 10, 15–26 (in Japanese, with English abstract).
- 741 Nishimura Y., Tanioka Y., & Hirakawa K. (2004). Beachside Trace for the 2003 Tokachi-oki Tsunami Run-up. *Zisin (Journal of*
742 *the Seismological Society of Japan. 2nd ser.)*, 57(2), 135–138. https://doi.org/10.4294/zisin1948.57.2_135
- 743 Noda, A., & TuZino, T. (2010). Shelf–slope sedimentation during the late Quaternary on the southwestern Kuril forearc margin,
744 northern Japan. *Sedimentary Geology*, 232(1–2), 35–51. <https://doi.org/10.1016/j.sedgeo.2010.09.008>
- 745 Noda, A., & Katayama, H. (2011) *Sedimentological Map Offshore of Cape Erimo. Marine Geology Map Series, no. 72,*
746 *Geological Survey of Japan, AIST.*
- 747 Okada, Y. (1985). Surface deformation due to shear and tensile faults in a half-space, *Bulletin of the Seismological Society of*
748 *America*, 75, 1135–1154.
- 749 Okuno, J., Nakada, M., Ishii, M., & Miura, H. (2014). Vertical tectonic crustal movements along the Japanese coastlines inferred
750 from late Quaternary and recent relative sea-level changes. *Quaternary Science Reviews*, 91, 42–61.
751 <https://doi.org/10.1016/j.quascirev.2014.03.010>
- 752 Philibosian, B., & Meltzner, A. J. (2020). Segmentation and supercycles: A catalog of earthquake rupture patterns from the
753 Sumatran Sunda Megathrust and other well-studied faults worldwide. *Quaternary Science Reviews*, 241, 106390.
754 <https://doi.org/10.1016/j.quascirev.2020.106390>
- 755 Reimer, P. J., Austin, W. E. N., Bard, E., Bayliss, A., Blackwell, P. G., Ramsey, C. B., et al. (2020). The IntCal20 Northern
756 Hemisphere Radiocarbon Age Calibration Curve (0–55 cal kBP). *Radiocarbon*, 62(4), 725–757.
757 <https://doi.org/10.1017/RDC.2020.41>
- 758 Richmond, B., Szczuciński, W., Chagué-Goff, C., Goto, K., Sugawara, D., Witter, R., et al. (2012). Erosion, deposition and
759 landscape change on the Sendai coastal plain, Japan, resulting from the March 11, 2011 Tohoku-oki tsunami. *Sedimentary*
760 *Geology*, 282, 27–39. <https://doi.org/10.1016/j.sedgeo.2012.08.005>
- 761 Sakuraba, H. (2019). Cape Erimo National Forest Management Project (Erino Town, Hokkaido), *Suirikagaku*, 62, 119–124 (in
762 Japanese). https://doi.org/10.20820/suirikagaku.62.6_119
- 763 Salditch, L., Stein, S., Neely, J., Spencer, B. D., Brooks, E. M., Agnon, A., & Liu, M. (2020). Earthquake supercycles and Long-
764 Term Fault Memory. *Tectonophysics*, 774, 228–289. <https://doi.org/10.1016/j.tecto.2019.228289>
- 765 Satake, K. (2015). Geological and historical evidence of irregular recurrent earthquakes in Japan. *Philosophical Transactions of*
766 *the Royal Society A: Mathematical, Physical and Engineering Sciences*, 373(2053), 20140375.
767 <https://doi.org/10.1098/rsta.2014.0375>
- 768 Satake, K., Nanayama, F., & Yamaki, S. (2008). Fault models of unusual tsunami in the 17th century along the Kuril trench.
769 *Earth, Planets and Space*, 60(9), 925–935. <https://doi.org/10.1186/BF03352848>
- 770 Satake, K., Fujii, Y., Harada, T., Namegaya, N. (2013). Time and Space Distribution of Coseismic Slip of the 2011 Tohoku
771 Earthquake as Inferred from Tsunami Waveform Data. *Bulletin of the Seismological Society of America*, 103 (2B), 1473–1492.
772 <https://doi.org/10.1785/0120120122>
- 773 Sawai, Y., Satake, K., Kamataki, T., Nasu, H., Shishikura, M., Atwater, B. F., et al. (2004). Transient Uplift After a 17th-Century
774 Earthquake Along the Kuril Subduction Zone. *Science*, 306(5703), 1918–1920. <https://doi.org/10.1126/science.1104895>

- 775 Sawai, Y., Kamataki, T., Shishikura, M., Nasu, H., Okamura, Y., Satake, K., et al. (2009). Aperiodic recurrence of geologically
776 recorded tsunamis during the past 5500 years in eastern Hokkaido, Japan. *Journal of Geophysical Research: Solid Earth*, 114,
777 B01319. <https://doi.org/10.1029/2007JB005503>
- 778 Sawai, Y. (2020). Subduction zone paleoseismology along the Pacific coast of northeast Japan — progress and remaining
779 problems. *Earth-Science Reviews*, 208, 103261. <https://doi.org/10.1016/j.earscirev.2020.103261>
- 780 Sella, G. F., Dixon, T. H., & Mao, A. (2002). REVEL: A model for Recent plate velocities from space geodesy: REVEL-
781 RECENT PLATE VELOCITIES FROM SPACE GEODESY. *Journal of Geophysical Research: Solid Earth*, 107(B4), ETG 11-
782 1-ETG 11-30. <https://doi.org/10.1029/2000JB000033>
- 783 Shigeno, K., & Nanayama, F. (2016). Sedimentary Process of a Small Sandy Event Deposit due to a Storm Surge and Storm
784 Waves Generated by a Typhoon: An Example from the Hirahama Coastal Lowland along the Western Example from the
785 Hirahama Coastal Lowland along the Western Coast of the Oshima Peninsula between 09:00 September 18 and 01:00 September
786 19, 1959. *Journal of Geography (Chigaku Zasshi)*, 125, 747–762 (in Japanese, with English abstract).
787 <https://doi:10.5026/jgeography.125.747>
- 788 Sieh, K., Natawidjaja, D. H., Meltzner, A. J., Shen, C.-C., Cheng, H., Li, K.-S., et al. (2008). Earthquake Supercycles Inferred
789 from Sea-Level Changes Recorded in the Corals of West Sumatra. *Science*, 322(5908), 1674–1678.
790 <https://doi.org/10.1126/science.1163589>
- 791 Soria, J. L. A., Switzer, A. D., Pilarczyk, J. E., Siringan, F. P., Khan, N. S., & Fritz, H. M. (2017). Typhoon Haiyan overwash
792 sediments from Leyte Gulf coastlines show local spatial variations with hybrid storm and tsunami signatures. *Sedimentary
793 Geology*, 358, 121–138. <https://doi.org/10.1016/j.sedgeo.2017.06.006>
- 794 Soria, J. L. A., Switzer, A. D., Pilarczyk, J. E., Tang, H., Weiss, R., Siringan, F., et al. (2018). Surf beat-induced overwash during
795 Typhoon Haiyan deposited two distinct sediment assemblages on the carbonate coast of Hernani, Samar, central Philippines.
796 *Marine Geology*, 396, 215–230.
- 797 Szczuciński, W. (2020). Post-depositional changes to tsunami deposits and their preservation potential. In M. Engel, J. Pilarczyk,
798 S. Matthias, M. D. Brill, & Ed Garrett (Eds.), *Geological records of tsunamis and other extreme waves*. Elsevier.
799 <https://doi.org/10.1016/B978-0-12-815686-5.00021-3>
- 800 Szczuciński, W., Kokociński, M., Rzeszewski, M., Chagué-Goff, C., Cachão, M., Goto, K., & Sugawara, D. (2012). Sediment
801 sources and sedimentation processes of 2011 Tohoku-oki tsunami deposits on the Sendai Plain, Japan — Insights from diatoms,
802 nannoliths and grain size distribution. *Sedimentary Geology*, 282, 40–56. <https://doi.org/10.1016/j.sedgeo.2012.07.019>
- 803 Szczuciński, W., Pawłowska, J., Lejzerowicz, F., Nishimura, Y., Kokociński, M., Majewski, W., Nakamura, Y., Pawłowski, J.,
804 2016. Ancient sedimentary DNA reveals past tsunami deposits. *Marine Geology*, 381, 29–33. [https://doi.org/10.1016/j.
805 margeo.2016.08.00](https://doi.org/10.1016/j.margeo.2016.08.00)
- 806 Switzer, A. D., Felix, R. P., Soria, J. L. A., & Shaw, T. A. (2020). A comparative study of the 2013 typhoon Haiyan overwash
807 sediments from a coastal cave and beach system at Salcedo, Eastern Samar, central Philippines. *Marine Geology*, 419, 106083.
808 <https://doi.org/10.1016/j.margeo.2019.106083>
- 809 Takada, K., Nakata, T., Miyagi, T., Haraguchi, T. & Nishitani, Y. (2002). Handy Geoslicer—new soil sampler for Quaternary
810 geologist. *Chishitsu News*, 579, 12-18 (in Japanese).
- 811 Takashimizu, Y., Urabe, A., Suzuki, K., & Sato, Y. (2012). Deposition by the 2011 Tohoku-oki tsunami on coastal lowland
812 controlled by beach ridges near Sendai, Japan. *Sedimentary Geology*, 282, 124–141.
813 <https://doi.org/10.1016/j.sedgeo.2012.07.004>
- 814 Tanioka, Y., Nishimura, Y., Hirakawa, K., Imamura, F., Abe, I., Abe, Y., et al. (2004). Tsunami run-up heights of the 2003
815 Tokachi-oki earthquake. *Earth, Planets and Space*, 56(3), 359–365. <https://doi.org/10.1186/BF03353065>

- 816 Tetsuka, H., Goto, K., Ebina, Y., Sugawara, D., & Ishizawa, T. (2020). Historical and geological evidence for the 17th-century
817 tsunami(s) along Kuril and Japan trenches: Implications for the origin of the AD 1611 Keicho earthquake and tsunami, and for
818 the probable future risk potential. *Geological Society, London, Special Publications*, 501. <https://doi.org/10.1144/SP501-2019-60>
- 819 Watanabe, M., Goto, K., Bricker, J. D., & Imamura, F. (2018). Are inundation limit and maximum extent of sand useful for
820 differentiating tsunamis and storms? An example from sediment transport simulations on the Sendai Plain, Japan. *Sedimentary
821 Geology*, 364, 204–216. <https://doi.org/10.1016/j.sedgeo.2017.12.026>
- 822 Wessel, P., Smith, W. H. F., Scharroo, R., Luis, J., & Wobbe, F. (2013). Generic Mapping Tools: Improved version released.
823 *Eos, Transactions American Geophysical Union*, 94(45), 409–410. <https://doi.org/10.1002/2013EO450001>
- 824 Yoshii, T., Tanaka, S., & Matsuyama, M. (2017). Tsunami deposits in a super-large wave flume. *Marine Geology*, 391, 98–107.
825 <https://doi.org/10.1016/j.margeo.2017.07.020>
- 826 Yokoyama, Y., Miyairi, Y., Aze, T., Yamane, M., Sawada, C., Ando, Y., et al. (2019). A single stage Accelerator Mass
827 Spectrometry at the Atmosphere and Ocean Research Institute, The University of Tokyo. *Nuclear Instruments and Methods in
828 Physics Research Section B: Beam Interactions with Materials and Atoms*, 455, 311–316.
829 <https://doi.org/10.1016/j.nimb.2019.01.055>