1   Paleotsunami research along the Nankai Trough and Ryukyu Trench subduction zones
2   – Current achievements and future challenges –
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1. Introduction

Facing the convergent plate boundary along the northwestern Pacific margin, the Japanese islands have repeatedly suffered from great subduction zone earthquakes and related tsunamis. As it is a mountainous country, the large population and industrial activity are concentrated on small coastal plains, making them particularly vulnerable to these natural hazards. The most recent example is the 2011 Tōhoku earthquake and tsunami, which occurred along the northern Japan Trench, where the Pacific Plate subducts beneath the Japanese archipelago.

One area highly likely to generate a future magnitude (M) 8.5+ subduction zone earthquake and tsunami is the Nankai Trough-Ryukyu Trench subduction zone, where the Philippine Sea Plate descends beneath the Japanese archipelago (e.g., HERP, 2013). Forecasting of the next earthquake and tsunami in these regions is an urgent issue for the Japanese government's national disaster prevention measures. To address this issue, many research projects have sought to identify the locations, timing, and size of past great earthquakes, based on historical records along with geological evidence. Documented Nankai Trough earthquakes and related tsunamis date back to the 7th century CE and suggest that M8-class great earthquakes have occurred at an interval of 100-200 years in this region (HERP, 2013), with 75 years having passed since the last events, which occurred in 1944 and 1946 CE.

On the other hand, documented records of great earthquakes and tsunamis along the Ryukyu Trench are limited mainly to the 17th century CE and later. It was once thought that no great earthquakes occur along the Ryukyu Trench; however, results from seismology and tsunami deposit studies since 2011 CE have overturned this theory and have highlighted the importance of estimating the magnitude and location of possible future great earthquakes in this region (Ando et al., 2009, 2012a, 2018).

Both the historical and geological records are inevitably biased by age and region as archives for paleoseismic events. Furthermore, geological records tend to be biased towards larger events. However, the geological record has the greatest advantage of all: the length of time it covers. Geological and archeological studies have been conducted to supplement the imperfect documented earthquake and tsunami history and to extend the record back to earlier periods. The first line of these studies is "Earthquake archaeology", established in the late 1980s, which seeks to identify the past earthquakes based on evidence of strong ground shaking, such as liquefaction features, found in archeological sites (Sangawa, 1992). These studies suggested the possibility of subduction zone earthquakes, such as the 887 and 684 CE Tōkai earthquakes along the eastern Nankai trough, which had not been confirmed in the documentary record. Several earlier Nankai Trough earthquakes dating back to 1800-2000 years ago have been also proposed from the archaeological data (e.g., Sangawa, 2007). However, this method generally

60   does not provide information on tsunamis, which are a key characteristic of major subduction
61   zone earthquakes.

62   A second line of research, paleotsunami studies, commenced along the Ryukyu Trench in
63   the 1960s, and full-scale research along the Nankai Trough began in the 1990s. Tsunamis
64   following two M9-class earthquakes that occurred in close succession in the 21st century, the
65   2004 CE Indian Ocean tsunami and 2011 CE Tōhoku tsunami, have provided a clear incentive
66   to study the record of past tsunamis and incorporate paleotsunami data into forecasts of future
67   events. Results of paleotsunami studies along the Nankai and Sagami Troughs up to 2007 CE
68   are summarized in Komatsubara and Fujiwara (2007). Garrett et al. (2016) provides a
69   comprehensive review of the recurrence history of great Nankai Trough earthquakes in the
70   Middle to Late Holocene, including not only tsunami deposits and liquefaction features on land,
71   but also marine and lacustrine turbidites.

72   The Geological Society of Japan published a special issue summarizing the results of
73   paleotsunami research along the Japanese archipelago as one of the commemorative
74   publications celebrating its 125th anniversary (Journal of the Geological Society of Japan 123,
75   No. 10). Wallis et al. (2018) describes the changes in the Japanese government's policy
76   concerning tsunami countermeasures following the Tōhoku event, and the changes in the
77   paleotsunami community's research direction reflecting the new policy. The policy (CDMC,
78   2011) reflected on the overreliance on geodetic observation data (~130 year-long) and "recent
79   and reliable" historical documents (covering the last ~400 years) in the long-term earthquake
80   prediction for Japan, and stipulated that all information from various different sources including
81   geological data should be considered in the new earthquake and tsunami countermeasures (see
82   summary in Goto et al., in this issue). As a response to this situation, in Japan, many
83   paleotsunami studies have been conducted and some are in progress under the auspices of the
84   national and local governments.

85   Tsunami deposit studies along the Nankai and Ryukyu subduction zones have been
86   independently reviewed (e.g., Fujiwara and Tanigawa, 2017; Goto, 2017), but there is no
87   comprehensive review comparing the two or summarizing the latest research. In this review, we
88   intend to address this point. For example, differences in the climate and geology of the two
89   regions create differences in the occurrence and composition of typical tsunami deposits: sandy
90   tsunami deposits for the Nankai and tsunami boulders (mainly coral blocks) for the Ryukyu
91   subduction zone. Additionally, as the two regions comprise a continuous subduction zone, the
92   possibility of a giant earthquake spanning both regions must carefully be assessed (Furumoto
93   and Ando, 2009). For this reason, synthesizing the earthquake and tsunami histories of both
94   subduction zones is particularly important.

Earthquake and tsunami forecasts operate at various timescales ranging from earthquake early warnings seconds before shaking is felt to long-term probability predictions over decades where geological data is predominant (e.g., Satake and Atwater, 2007). Every case requires the specification of three factors: where, when, and how big. From this perspective, this paper reviews the results of research on onshore tsunami deposits, their limitations, and existing knowledge gaps, and discusses measures to address the remaining issues. When and where earthquakes occurred is reconstructed based mainly on tsunami deposits studies with the help of other geological evidence. A comprehensive earthquake recurrence model is still incomplete for these subduction zones. This paper therefore also addresses the role that tsunami deposit research plays in the development of this recurrence model. The magnitude of earthquakes that occur along a subduction zone varies widely (e.g., Satake and Atwater, 2007); this paper discusses the variation in the size of earthquakes inferred from tsunami deposit research.

The possibility of the occurrence of an unforeseen or unexpected large earthquake and tsunami is a major point of reflection following the 2011 Tōhoku event. Learning from this, an estimation of the "maximum possible earthquake and tsunami" that could affect the Japanese coast has been included in disaster prevention measures in Japan (CDMC, 2011; Cabinet Office, 2012a). The CDMC (2011) stated that future tsunami countermeasures will basically require the assumption of two possible levels of tsunamis. Level one reflects the largest tsunamis known to occur on centennial timescales and forms the basis for designing the height of seawalls and other features. Level two is the maximum possible tsunami, which occurs at a much lower frequency than the first level but is considered to be more destructive. The hypothetical size of a level two tsunami is based on studies of tsunami deposits and crustal movements. The 2011 Tōhoku tsunami is an example of a level two tsunami. Countermeasures against a level two tsunami could include relocating settlements to higher ground and building huge seawalls, but specific measures are difficult to implement. Complicating attempts to mitigate future impacts, it is unknown whether such extremely large earthquakes and tsunamis have occurred in the past along the Nankai and Ryukyu subduction zones. This paper refers to the plausibility of this scenario from the standpoint of the geological evidence.

2. Tectonic setting

2.1. Plate subduction and related geomorphology

The Nankai and Ryukyu subduction zones mark the northwestern margin of the Philippine Sea Plate (Fig. 1), but the nature of plate subduction, geological structures, and occurrence mode of earthquakes are largely different between the two. The Palau-Kyushu Ridge forms the boundary between the two subduction zones. The Nankai Trough extends for 700 km from the southwest of Shikoku Island in the west to Suruga Bay in the east, where it is also called the

131 Suruga Trough. The Ryukyu Islands consist of numerous small islands extending over 1000 km

132 along the Ryukyu Trench, which belong to the subtropical climate zone and hence the

133 depositional environment is very different from other parts of Japan. The convergence rate is

134 generally 40-50 mm a^{-1} along the Nankai subduction zone and 50-63 mm a^{-1} along the Ryukyu

135 subduction zone, respectively. The slope of the descending slab, with some local variations, is

136 generally steeper in the Ryukyu subduction zone.

137 The Amami Plateau, Daito Ridge, Oki-Daito Ridge and basins between them extend in

138 E-W or NW-SE directions on the Philippine Sea Plate. Western extensions of these topographic

139 irregularities subduct beneath the Ryukyu arc (e.g., Kato, 1993). The Ryukyu and Nankai

140 subduction zones are represented by old (40-49 Ma), cold slab that originated from the West

141 Philippine Basin and young (<20 Ma), hot slab that originated from the Shikoku Basin,

142 respectively (e.g., Ishizuka et al., 2011). A thick accretionary prism characterizes the Nankai

143 Trough, in which high-angle reverse faults (splay faults) branching off the plate boundary

144 megathrust may move together with the main thrust during a great earthquake and enhance the

145 tsunami (Baba et al., 2006; Moore et al., 2007).

146 The Nankai Trough fault region has been empirically divided into five fault planes (rupture

147 zones), which denoted as segments A to E from west to east, respectively (Fig. 1) (Ando, 1975a,

148 b; Ishibashi, 1976, 1981). This classification is mainly based on studies of the source

149 mechanisms of historical earthquakes and the forearc submarine topography, which reflects the

150 seismotectonic activities along this region. Each of the five segments has a length of 100-150

151 km along the trough axis and almost corresponds to the distribution of forearc basins (e.g.,

152 Awata and Sugiyama, 1989; Sugiyama, 1990). In general, plate boundary earthquakes

153 (M8-class) that are considered to have source areas in areas A to B and C to E are called Nankai

154 and Tōkai earthquakes, respectively.

155 Allocating the source areas of past earthquakes to any of segments A to E is based on the

156 implicit assumption that a large earthquake recurs on its own fault plane. The necessity to

157 criticize and review this assignment has recently been pointed out (e.g., Seno, 2012). The

158 Hyūga-nada area (Figs. 1, 2), where M7-class earthquakes occur separately from the Nankai and

159 Tōkai earthquakes, may be another seismic zone denoted as Z (e.g., Wells et al., 2003). This

160 empirical segmentation of rupture zones has not been applied to the Ryukyu Trench as historical

161 and seismological information is scarce.

162 Flight of marine terraces, mainly from mid-Pleistocene to Holocene in age, are distributed

163 on the peninsulas, capes and islands of the Nankai and Ryukyu subduction zones (Koike and

164 Machida, 2001), and there have been many discussions about their formation process in relation

165 to sea-level changes and crustal movements (e.g. Yoshikawa et al., 1964; Webster et al., 1998).

166 They show a wide regional variation in average uplift rates. For example, the terraces of marine

167 oxygen isotope stage (MIS) 5e are 40-50 m high in many areas, but reach nearly 190 m at Cape

168 Muroto, the boundary between segments A and B. Coseismic uplift and subsidence during

169 historical earthquakes characterize the tips of the peninsulas and their inland plains, respectively

170 (e.g., Shikoku; Geographical Survey Institute, 1952; Tokai; Ishibashi, 1981). In these area,

171 reverse sense crustal deformation is observed in the interseismic period; gradual subsidence of

172 the peninsulas and uplift of the lowland, respectively (Geographical Survey Institute, 2008).

173 　　　Kikai Island, located at the northern margin of the Amami Islands on the subducting

174 Amami Plateau, is the closest island to the Ryukyu Trench axis. It shows an extremely high

175 uplift rate, with the height of the MIS 5e marine terrace exceeding 200 m. On this island,

176 Holocene marine terraces consisting of coral reefs record several-meter-scale uplift events that

177 recurred with 1500-year intervals (Webster et al., 1998). The other islands along the Ryukyu

178 Trench do not show remarkable crustal deformation, but the coastal notch in Ishigaki Island

179 records a cumulative uplift of ~2 m in the last 2000 years (Kawana, 1989). Some of these

180 marine terraces distributed along the Nankai and Ryukyu coasts have provided the theoretical

181 basis for time-predictable recurrence models for subduction zone earthquakes (e.g., Shimazaki

182 and Nakata, 1980).

183

184 **2.2. Seismicity**

185 　　　Understanding the state of inter-plate coupling and seismogenesis in the subduction zones

186 provides basic information for estimating the region's potential of great earthquake and tsunami.

187 Especially, in the Hyūga-nada region, connecting the Nankai and Ryukyu subduction zones, this

188 information is the key to identifying the geological and seismological differences between the

189 two.

190

191 **Geodetic deformation and inferences of inter-plate coupling along the Nankai Trough and**

192 **Ryukyu Trench subduction zones**

193 　　　The source areas of large earthquakes are usually expected to present strong inter-plate

194 coupling during inter-seismic periods (e.g., Lay and Kanamori, 1981). The degree of the

195 coupling can vary over fault areas, reflecting differences in the fault's frictional properties. If a

196 fault creeps steadily, it has a low degree of coupling, while high coupling results from a

197 seismogenic fault that is stuck during inter-seismic periods. The degree of the coupling is

198 quantified in geodesy (Savage, 1983) by the slip-deficit rate (SDR) (or back-slip rate), which is

199 the rate of upper plate displacement along with the plate interface driven by the subduction

200 motion of the lower plate. The coupling coefficient is also determined as the SDR divided by

201 the long-term average rate of the relative motion between the two plates. When the fault is

202 completely coupled, the SDR is equal to the long-term average relative plate motion rate and the

203 coupling coefficient is unity. SDRs are estimated from geodetic data by applying slip inversion
204 analysis (Savage, 1983).

205   Several geodetic inferences of inter-plate coupling have been recorded along the Nankai
206 Trough subduction zone (Miyazaki and Heki, 2001; Ito and Hashimoto, 2004; Nishimura and
207 Hashimoto, 2006; Wallace et al., 2009; Loveless and Meade, 2010; Yokota et al., 2016), while
208 only a few studies have been conducted along the Ryukyu Trench (Ando et al., 2009; Tadokoro
209 et al., 2018) due to limited spatial coverage of onshore GNSS observations in this island area.
210 Along the Nankai Trough, spatial distributions of SDR have been inferred and certain
211 heterogeneous structures have been identified (Fig. 1), owing to a dense onshore GNSS network
212 called GEONET and the recent development of seafloor geodetic observation networks by the
213 Japan Coast Guard (Yokota et al., 2016). The seafloor observation sites deployed near the
214 trough axis contribute greatly to resolving the previously unknown coupling at shallow depths
215 of the plate interface (Sagiya and Thatcher, 1999), which caused an ambiguity in the estimation
216 of the maximum earthquake size in the area. Lack of knowledge of shallow coupling was clearly
217 illustrated along the Japan Trench subduction zone, which hosted the 2011 Tōhoku-oki
218 earthquake (Nishimura et al., 2000; Loveless and Meade, 2011).

219   All of the SDR estimates indicate that the Nankai Trough subduction interface is strongly
220 coupled overall, with Yokota et al. (2016) further resolving regional heterogeneity, with three
221 areas of relatively stronger coupling with higher SDRs inferred offshore Shikoku Island, the Kii
222 Peninsula and the region from the Atsumi Peninsula to the Omaezaki Peninsula (Fig. 1). The
223 first two areas seen to coincide with the source areas of the 1946 CE Nankai (M8.0) and the
224 1944 CE Tōnankai (M7.9) earthquakes, and the third area seems to be included in the source
225 area of the 1854 CE Ansei-Tōkai earthquake (Ando, 1975b). Given such correlations, the
226 observed heterogeneous distribution of SDR might have played a role in defining the
227 segmentation of historical earthquakes. However, some earthquakes can involve multiple
228 segments, as in the 1707 Hōei (M8.6) earthquake, which is considered to have ruptured all three
229 of the high SDR segments (Ando, 1975b), and the earthquake recurrence pattern is not always
230 characteristic (Seno, 2012).

231   Near the southwestern end of the Nankai Trough subduction zone, the degree of coupling
232 offshore of southern Kyushu Island becomes rather ambiguous due to poorer observational
233 coverage; however, geodetic inferences still tend to suggest weaker coupling in the southern
234 half of the Hyūga-nada region (Nishimura and Hashimoto, 2006; Wallace et al., 2009; Loveless
235 and Meade, 2010) (see Fig.1), reflecting the rotation of the displacement vectors observed over
236 southern Kyushu. The location of this weak coupling is suggested to be correlated with the
237 subducting Kyushu-Palau ridge (Wallace et al., 2009; Yamamoto et al., 2013). To the south,

238 inferences of inter-plate coupling become more difficult due to the much poorer observational
239 coverage.

240   Along the Ryukyu Trench subduction zone, inter-plate coupling has been evaluated
241 qualitatively due to the sparseness of onshore observational stations, with generally weak
242 coupling inferred (Nishimura et al., 2004). However, more quantitative geodetic inferences have
243 been conducted with the addition of ocean bottom geodetic observations. Seafloor displacement
244 data has recently been retrieved from the area offshore the Okinawa Islands, in the center of the
245 Ryukyu Trench, and almost complete coupling of the plate interface is inferred by Tadokoro et
246 al. (2018), in contrast to the onshore-based estimation. It is also suggested that this highly
247 coupled area corresponds to the source area of the 1791 CE M 8 inter-plate earthquake.

248

249 **Seismological inferences of inter-plate coupling using repeating earthquakes**

250   Repeating earthquakes are moderate-sized earthquakes that are considered to occur
251 repeatedly at the same points on faults due to the similarity in their seismic waveforms. They
252 are used to infer the degree of inter-plate coupling in a similar fashion to creep meters (Nadeau
253 and Johnson, 1998). While the absence of repeating earthquakes characterizes both the end
254 members of the fully coupled and the fully decoupled fault, once repeating earthquakes are
255 observed, inter-plate coupling can be estimated, with higher production rates reflecting weaker
256 coupling and more frequent occurrence of fault slip. The use of repeating earthquakes can
257 supplement geodetic inferences due to its higher detection capability, reflecting the weaker
258 geometrical spreading nature of the seismic waves than the geodetically-observed static
259 displacement. Nevertheless, both observations are inevitably limited in island areas.

260   Igarashi (2010) inferred inter-plate coupling by using repeating earthquakes and mapped its
261 spatial variation on the plate interfaces off the Japanese islands. The background seismic
262 activity of the Nankai Trough is generally low reflecting the strong coupling and repeating
263 earthquakes are rarely found except the shallowest portion of the plate interface near the trough
264 axis. However, it is evident that inter-plate coupling becomes weaker from the Hyūga-nada
265 region towards the Ryukyu Trench to the south, matching to the inference from geodetic
266 observations. Yamashita et *al.* (2012) increased the detection of repeating earthquakes, focusing
267 on the Hyūga-nada region, and mapped a more detailed spatial distribution of inter-plate
268 coupling (Fig. 1). In this region, the coupling appears to be weakest along the band where the
269 Kyushu-Palau ridge is being subducted (Yamamoto et al., 2013), with the repeating earthquakes
270 accommodating most of the long-term relative plate motion at shallower (<~20 km) and deeper
271 (>40 km) depths. However, to the south, the coupling increases in the area off Tanegashima
272 Island, corresponding to the northern end of the Ryukyu Trench, where repeating earthquakes
273 accommodate approximately half the relative plate motion.

274   In the Hyūga-nada region, the history of moderate- and large-sized earthquakes appears to
275 be matched with the inferred coupling distribution (Yamashita et al., 2012). In the northern area
276 of the Hyūga-nada, M 6 and 7-class earthquakes have repeatedly occurred, while in the southern
277 part, the 1996 CE M 6.5 earthquake occurred. Between these regions, major earthquakes are not
278 known, while moderate-sized earthquakes (5<M<6.5) have been observed in the relatively
279 strongly coupled area.

280   Inter-plate coupling along the Ryukyu Trench seems to show a systematic variation where
281 the northern region – from Tanegashima Island to Amami Island – exhibits relatively strong
282 coupling and the southern region around the Sakishima Islands exhibits weaker coupling.
283 However, the coupling state may not be definitively identified solely based on repeating
284 earthquakes in these island areas; in the central region, while repeating earthquakes suggest
285 moderate coupling, seafloor observations suggest almost full coupling (Tadokoro et al., 2018).

286

287 **Spatial distribution of shallow and deep slow earthquakes; implications for the state of**
288 **inter-plate coupling**

289   Weak inter-plate coupling does not necessarily mean that faults creep steadily. Weak
290 coupling can reflect intermittent slip events on the plate interfaces. Such slip events are
291 generally called slow earthquakes and include slow slip events (SSEs), low-frequency
292 earthquakes and (non-volcanic) tremor. Slow earthquakes are thought to occur along the
293 transition zones of the seismogenic and aseismic zones of plate interfaces (Ando et al., 2012b;
294 Obara and Kato, 2016) both at the up-dip (~5 km) and down-dip ends (~30 km) of the
295 seismogenic zones that generate the huge earthquakes.

296   The family of slow earthquakes has been discovered over the past twenty years along the
297 Nankai Trough by seismological (e.g., Obara, 2002) and geodetic (e.g., Hirose and Obara, 2005)
298 observations, and their correlation with inferred inter-plate coupling states are confirmed in
299 great detail. Along the plate interface at depths from ~30 to ~40 km, tremor and short-term
300 SSEs (S-SSEs) are generally observed with recurrence periods of ~6 months (Obara, 2002;
301 Hirose and Obara, 2005), while long-term SSEs with M 7 have recurred at ~10-year intervals
302 particularly beneath Lake Hamana in Tōkai region (Suito and Ozawa, 2009 and references
303 therein). For the Ryukyu Trench, seismological observations suggest that slow earthquakes,
304 called very low frequency earthquakes, probably occur on the plate interface over the entire
305 Ryukyu Trench (Ando et al., 2012a), and geodetic observations identify seemingly
306 corresponding short-term SSEs (S-SSEs) and long-term SSEs (L-SSEs) (Heki and Kataoka,
307 2008; Nishimura, 2014). Nishimura (2014) suggested that the occurrence of S-SSEs at shallow
308 depths may be a signature of incomplete inter-plate coupling. Tadokoro et al. (2018) suggested

309   the inferred strong coupling area does not overlap with the SSE areas, and heterogeneity in the

310   areas of SSE occurrence should exist along the Ryukyu Trench.

311   The Hyūga-nada region appears to present a transitional character in terms of the spatial

312   distributions of slow earthquakes. In the other areas of the Nankai Trough, shallow slow

313   earthquakes and deep S-SSEs and tremors are both observed, while deep S-SSE/tremor is not

314   observed in the Hyūga-nada region (Obara and Kato, 2016). The occurrence of shallow slow

315   earthquakes has been characterized in great detail owing to the recent deployment of ocean

316   bottom seismometers (Yamashita et al., 2015). Tremor was observed to migrate in the shallow

317   depths of the plate interface, and the lateral extent of the migration was delimited by the

318   possible area of the subducting Kyushu-Palau ridge (Yamamoto et al., 2013). The down-dip

319   limit of the migration was not reached until ~ 20 km, corresponding to the hypocentral depths of

320   M 7-class earthquakes in this region, suggested that it stopped at the shallower end of the

321   seismogenic zone (Yamashita et al., 2015). This implies that the intermediate depths of the plate

322   interface are still coupled and have the potential to nucleate a major earthquake. The

323   geophysical observations clearly show that the inter-plate coupling of this area is relatively low

324   overall, but these observations still leave the possibility that the rupture of this coupled zone

325   could be triggered by neighboring large earthquakes, or the rupture of this area could facilitate

326   interactions between neighboring large earthquakes, if their timings are synchronized.

327   While the above-mentioned geophysical observations have clarified the spatial

328   characteristics of the inter-plate properties, we should still remember that our experience is

329   limited by the short history of geophysical observations. The states of inter-plate coupling can

330   slowly fluctuate even over the course of several years (Uchida et al., 2016), and limited

331   snapshots can fail to reveal critical properties. Integration of paleoseismological evidence is still

332   critically important to capture phenomena spanning long periods of time.

333

334   **3. Record of tsunamis and tsunami deposits**

335   **3.1. Nankai Trough region**

336   **3.1.1. Historical Tōkai and Nankai earthquakes and tsunamis**

337   The Nankai Trough region, which is close to the political and cultural center of Japan, the

338   Kyoto-Nara-Osaka area, has the longest and most complete documented record for the

339   recurrence of subduction zone earthquakes and tsunamis in the world. According to this

340   1300-year long record, earthquakes have occurred in a variety of modes. Some earthquakes

341   have ruptured almost the entire subduction zone at the same time, while at other times, two

342   earthquakes, referred to as Tōkai and Nankai earthquakes, have occurred consecutively,

343   separated by short intervals (Fig. 2). Nankai earthquakes that are historically confirmed include

344   the 684 CE Hakuho (M8.0-8.5), 887 CE Nin-na (M8.0-8.5), 1361 CE Kōan (M8.2-8.5), 1707

345 CE Hōei, 1854 CE Ansei-Nankai (M8.4; only 32 hours after the 1854 Ansei-Tōkai earthquake),
346 and 1946 CE Showa-Nankai (M8.0) earthquakes, with further unconfirmed events in 1099 and
347 1614 CE. Tōkai earthquakes confirmed by documented records include the 1096 CE Eichō
348 (M8.0-8.5), 1498 CE Meiō (M8.2-8.4), 1707 CE Hōei (M8.6), 1854 Ansei-Tōkai (M8.4), and
349 1944 CE Tōnankai (M7.9) earthquakes, and a further rupture may have occurred in 1614 CE.
350 The 1707 earthquake was a simultaneous rupture of the Tōkai and Nankai fault segments
351 (segments A to D, and perhaps also including the southern part of E) and was the largest
352 earthquake in Japanese history until the occurrence of the 2011 Tōhoku earthquake.
353   There are two possibilities for a great earthquake that occurred along the Nankai Trough in
354 the early 17th century. The first is the 1605 CE Keichō earthquake, which generated a large
355 tsunami from Kanto to Kyushu. This event has been considered as a "tsunami earthquake",
356 because only weak shaking accompanied the large tsunami (e.g., Ishibashi and Satake, 1998).
357 Ando and Nakamura (2013) suggested that the rupture occurred along a shallow portion of the
358 plate interface. An alternative hypothesis is that the 1605 earthquake occurred along the
359 Izu-Ogasawara Trench, to the east of Izu-Bonin Arc (Harada et al., 2013). Instead of 1605 CE
360 event, Ishibashi (2014) proposed the 1614 CE earthquake as a candidate for a 17th century
361 Nankai Trough earthquake, based on records of strong ground shaking in Kyoto and tsunamis at
362 the Kii Peninsula. A Nankai earthquake paired with the 1498 CE Meiō Tōkai earthquake has not
363 been confirmed; nevertheless, liquefaction features in western Japan suggest the occurrence of a
364 Nankai earthquake around 1498 CE (Sangawa, 2001, 2007). Ishibashi (2014) proposed four
365 earthquakes that occurred between 1498 and 1512 CE as candidates, but there is no solid record
366 of tsunami inundation during this period.
367   Documentary records confirming the occurrence of a Tōkai earthquake paired with the
368 1361 CE Nankai earthquake have not yet been discovered. As a candidate for this earthquake,
369 Ishibashi (1998, 2014) proposed an earthquake that featured strong shaking in Kyoto, Nara, and
370 Kumano (north of Shiono-misaki; boundary between the segments B and C) on July 24th 1361
371 (2 days before the Nankai earthquake). Late 14th century coseismic uplift around Cape
372 Omaezaki (Fig. 2, site 27) shown by emerged boring bivalves (Kitamura et al., 2018c) and
373 marine terrace (Fujiwara et al., 2010) support Ishibashi's hypothesis.
374   The occurrence of the 1099 CE earthquake was originally inferred from documents
375 suggesting strong ground shaking in the Kyoto, Nara, and Osaka region (but less than 1096 CE
376 earthquake) and coastal subsidence in Kochi (around site 10 in Fig. 2); which matches general
377 characteristics of other Nankai earthquakes (e.g., Ishibashi, 1999). However, the occurrence of
378 this earthquake is uncertain due to the unreliability of underlying evidence, especially the lack
379 of reports of strong ground shaking and tsunami inundation in the area facing segments A and B.
380 Ishibashi (2016) suggested that the 1099 CE earthquake did not actually exist and that the 1096

381 CE earthquake was instead a full-length rupture of the Tōkai and Nankai segments. Ishibashi
382 (2014) also suggested the possibility that the 1099 CE Nankai earthquake did occur but was
383 smaller in magnitude than other Nankai earthquakes.

384 The possibility that a Tōkai earthquake occurred at the same time as the Nankai earthquake
385 on 22nd August, 887 CE has been suggested from a description in the Nihon Sandai Jitsuroku
386 (日本三代実録, "The true history of three reigns of Japan"; completed in 901 CE) (e.g.,
387 Ishibashi, 1999, 2014). The chronicle's record for that day shows that strong ground shaking
388 was felt in wide area including the Tōkai region along with the occurrence of an earthquake and
389 tsunami in the Nankai region. The occurrence of the 684 CE Nankai earthquake is documented
390 in the Nihon Shoki (日本書紀, The Chronicles of Japan, completed in 720 CE), with a report of
391 tsunami and coastal submergence in Kōchi, but there are no records of a corresponding
392 earthquake in the Tōkai region. Only liquefaction features in archaeological sites in the Tōkai
393 region have suggested as possible evidence of a Tōkai earthquake corresponding to the 684 CE
394 Nankai earthquake (e.g., Sangawa, 2001, 2007). However, given the lack of evidence for
395 tsunami to determine the occurrence of the Tōkai earthquake in either case, the possibility of an
396 inland earthquake from an active fault cannot be ruled out. These imperfections in the historical
397 earthquake record have been a hindrance in discussing the occurrence time of Tōkai and Nankai
398 earthquakes and their linkage.

399 The 1498 CE Meiō tsunami might have been somewhat higher than 1707 and 1854 CE
400 tsunamis, ~ 8 m along the Enshu-nada coast, which includes the sites 19 to 27 in Fig. 2 (Hatori,
401 1975). The height of the 1605 CE tsunami was generally 4 - 6 m and up to 13 m in Shikoku
402 (Murakami et al., 1996). The 1707 CE tsunami generally reached 5 - 8 m-high along the Tōkai
403 and Nankai coasts, over 10 m in some locations on the Kii Peninsula and Shikoku, and 2 - 4 m
404 along the Kyushu coast (Murakami et al., 1996, Watanabe, 1998). The 1854 CE Ansei-Tōkai
405 tsunami had a height of 5 - 8 m, with some isolated cases of ~20 m at the tips of promontories
406 (Watanabe, 1998). The 1854 CE Ansei-Nankai tsunami generally reached 5 - 6 m-high,
407 occasionally over 8 m along the Kii Peninsula and Shikoku (Murakami et al., 1996, Watanabe,
408 1998), and 2 - 4 m along the Kyushu coast (Cabinet Office, 2012b). The 1944 CE tsunami was
409 3 - 7 m-high along the eastern coast of Kii Peninsula and 0.9 - 2 m-high along the Enshu-nada
410 coast (Watanabe, 1998). The height of the 1946 CE Showa-Nankai tsunami was 2 - 4 m in
411 general, up to 6 m on the Kii and Sikoku coasts and 1 - 1.6 m on the Kyushu coast (Murakami et
412 al., 1996, Watanabe, 1998). These tsunamis generally caused greater human and economic
413 damages than resulted from ground shaking during historical Nankai Trough earthquakes (e.g.,
414 Yata, 2009, 2018; Ishibashi, 2014).

415

416 **3.1.2. Tsunami deposits**

417   Nankai Trough coasts have few extensive wetlands suitable for reconstructing the tsunami
418   inundation area, which characterize the major paleotsunami study fields in the world, such as
419   Hokkaido (e.g., Nanayama et al., 2003), Cascadia (e.g., Peters et al., 2007) and Chile (e.g.
420   Cisternas et al., 2005). Artificial disturbance of surface sediments reduces the chances of
421   finding tsunami deposits from this region, along with the difficulties of distinguishing tsunami
422   deposits from the other washover (mainly storm) deposits and river-flooding deposits.
423   Consequently, coastal back-barrier lakes and ponds, and infilled valleys have mainly been
424   targeted as paleotsunami study sites. The barriers between these sites and the sea are mainly
425   composed of sand dunes and beach ridges formed during the last seven to six thousand years,
426   after the mid Holocene sea level high stand in Japan (e.g., Sato et al., 2016a). Many of these
427   sites have gradually subsided through the Holocene, and stably accumulated fine-grained
428   sediments, sometimes rich in organic matter (e.g., Okamura and Matsuoka, 2012). They have a
429   high potential for preserving marine overwash event deposits. Coring surveys using hand corers,
430   piston corers, and geoslicers (Nakata and Shimazaki, 1997) have been the main methods to
431   reveal the tsunami deposits, while their depositional ages have mainly been determined by
432   radiocarbon dating. Due to the very limited distribution of widespread Holocene tephras along
433   the Nankai and Ryukyu subduction zones, it is difficult to estimate the age of tsunami deposits
434   based on volcanic ash stratigraphy and to compare them between regions, as has been done in
435   the Hokkaido (e.g., Nanayama et al., 2003) and Tōhoku (e.g., Sawai et al., 2012) regions.

436   Tsunami and possible tsunami deposits have been reported from a total of 23 study sites
437   along the Nankai Trough coast (Fig. 2, sites 7 to 20, 22 to 26, 28, 29, 31 and 32). Adjacent
438   points that cannot be distinguished on the map, such as some points in the southern part of the
439   Kii Peninsula, are counted as one. Garett et al. (2016) recalibrated all available published
440   radiocarbon ages to take advantage of the latest radiocarbon calibration curves, updated
441   estimates of local marine radiocarbon reservoir effects, and Bayesian age modeling approaches.
442   As a result, some of the originally reported correlations between sedimentary layers and
443   historical earthquakes have been revised. We follow Garrett et al. (2016) for the ages of tsunami
444   deposits summarized here.

445   Most of the reported tsunami and possible deposits are washover sand deposits intercalated
446   in muddy or peaty deposits, with two exceptions of the tsunami boulders found from the
447   southern Kii peninsula (Fig. 2, site 14) and the southern Izu peninsula (Fig. 2, site 32). Sandy
448   tsunami deposits generally show a tapering shape with a fining landward trend and sometimes
449   include remains of marine organism. They may be a single layer, but often consist of multiple
450   layers, each of which shows a graded structure. Sedimentary structures indicating deposition
451   from a tractive flow, such as basal erosion surface, cross-lamination, and inverse grading also
452   characterize many cases.

Historical tsunami deposits

Examples of tsunami deposits that can be accurately correlated to historical earthquakes are still limited in the Nankai Trough region (Fig. 2). Limited numbers of measurements and chronological uncertainties inherent in the radiocarbon approach make it difficult to distinguish individual historical tsunami deposits that formed at short intervals, often of less than 100 years, and compare them between distant sites.

The easternmost examples are reported by Kitamura and Kobayashi (2014a) and Kitamura et al. (2014) from Shimoda, southern Izu Peninsula (Fig. 2, Site 32). The former is a laminated sand bed about 10 cm thick, with rip-up clasts that is estimated to have been deposited between 430-290 cal. BP and 1950 CE, possibly due to the 1854 CE Ansei-Tōkai or 1707 CE Hōei tsunami. The latter is a 32-tonne boulder located on a coastal plateau, which is attributed to the 1854 CE Ansei-Tōkai earthquake based on the radiocarbon ages of sessile marine fossils attached to the boulder.

On the Ita Lowlands (Fig. 2, site 31), one possible sandy tsunami deposit – with diatoms suggesting a marine incursion – was found from a sedimentary sequence covering the last 2500 years (Sawai et al., 2016). Radiocarbon dating limited the age of the tsunami bed to 1200-1320 CE or 1150-1330 CE and Sawai et al. (2016) attributed it to the 1096, 1099, or 1361 CE Nankai Trough earthquake or the 1293 CE Kantō earthquake, centered on the Sagami Trough, east of Izu Peninsula. On the Shimizu Plain, Kitamura and Kobayashi (2014b) reported the deposition of gravelly sand beds associated with the erosion of the muddy bay bottom between 670–600 and 250–50 cal. BP (Fig. 2, Site 29). They inferred that the sand beds may be attributed to the tsunamis from 1498 CE Meiō or 1707 CE Hōei earthquakes.

Komatsubara et al. (2008) conducted paleotsunami research in a back-marsh at Shirasuka (Fig. 2, site 20) using a 4 m-long geoslicer and found exotic sand beds within the muddy sedimentary sequence. Comparing the radiocarbon ages obtained from the cores and historical documents, five sand beds, each of which is 10 - 20 cm thick, were attributed to tsunamis associated with the 1498, 1605, 1707, 1854 CE earthquakes and the 1680 or 1699 CE typhoon. This hypothesis was verified with new age models developed using radiocarbon ages (Garrett et al., 2018) and optically stimulated luminescence (OSL) ages (Riedesel et al., 2018) from subsequently excavated sediment cores from the same site as the original study. The sand unit A of Komatsubara et al., (2008), located beneath the 1498 CE tsunami deposit and originally interpreted as a slope failure deposit, was recognized as a possible tsunami deposit based on the occurrence of marine and brackish diatoms (Garrett et al. 2018). Both radiocarbon and OSL age models suggest an age for this tsunami deposit of the late 14th century and the layer may be attributed to the 1361 CE Tōkai earthquake (Garrett et al., 2018; Riedesel et al., 2018).

489   Fujino et al. (2018) conducted an array coring survey on the Shima Lowland (Fig. 2, site
490   17), a Holocene drowned valley, and found 10 exotic sand beds within the muddy marsh
491   sequence. A tsunami origin for these beds is inferred from sedimentary features common in
492   modern tsunami deposits, including marine bioclast-rich sand, alternating sub-layers of sand and
493   silt, rip-up clasts, and upward-fining structures. The youngest three tsunami beds were
494   tentatively attributed to 1498, 1096, and 684 CE Nankai Trough earthquakes.

495   A group of possible tsunami boulders, consisting of hundreds of angular boulders weighing
496   up to 100 tons (Namegaya et al., 2011), are found on a wave-cut bench near Cape
497   Shiono-misaki (Fig. 2, site 14). These boulders were derived from the Middle Miocene
498   quartz-porphyry dyke penetrating the basement sedimentary rock, which is located seaward of
499   the boulder group. The dyke and boulders are famous as the tourist spot of "Hashigui-iwa",
500   which means large rocks (iwa) look like bridge (Hashi) piles (gui or kui) in Japanese. Excluding
501   smaller ones < 1m in diameter, these boulders were not moved by the 1946 CE Showa-Nankai
502   tsunami or subsequent storms (Shishikura, 2013). Shishikura (2013) recognized two major
503   boulder transport events based on the radiocarbon ages of sessile marine fossils, the 12-14th
504   century and the 17-18th century CE, and attributed them to the 1361 and 1707 CE tsunamis,
505   respectively.

506   Tōkai earthquakes corresponding to the 887 and 684 CE Nankai earthquakes have
507   remained unconfirmed for a long time, but have recently been substantiated by the discovery of
508   tsunami deposits on the Otagawa Lowland (Fig. 2, site 25) in western Shizuoka Prefecture
509   (Fujiwara et al., 2020). Excavation walls exposed by river improvement work in the lowland
510   revealed a ~1-km long coast-normal and 4 to 5 m-deep cross section of the strand plain (Fig.
511   3A). Fujiwara et al. (2020) mapped four sandy tsunami deposits along the excavation walls,
512   each of which shows inundation over 2 km inland from the coast at the time of tsunami, with
513   landward thinning and fining trends. Radiocarbon ages link the younger two tsunami deposits to
514   the 1498 and 1096 CE Tōkai earthquakes. The older two deposits indicate the occurrence of
515   Tōkai earthquakes near to the end of the 7th and 9th centuries CE. The late 9th century tsunami
516   deposit, integrated with the description of strong ground shaking in the Nihon Sandai Jitsuroku,
517   demonstrates that the 887 CE earthquake was a full-length rupture of the Tōkai and Nankai
518   segments (Fujiwara et al., 2020). As the late 7th century tsunami deposit lacks data to identify
519   the precise date of its occurrence, it is not known whether a Tōkai earthquake is coincident with
520   the 684 CE Nankai earthquake.

521   For Shikoku and northern Kyushu, in addition to the summary provided by Okamura and
522   Matsuoka (2012) from 26 ponds, Baranes et al. (2016) conducted paleotsunami research in three
523   ponds. Among these 29 sites, historical tsunami deposits have been found in four locations. At
524   Ryujin-ike in northern Kyushu (Fig.2, site 7), separated by a 17 m high and 100 m wide sand

525 dune, a narrow channel is the only access to the Pacific. According to historical documents, the

526 1707 CE Hōei tsunami, over 10 m-high, was much higher than that of 1854 and 1946 CE

527 tsunamis around Ryujin-ike (e.g., Chida and Nakaue, 2007). Among 40 sand beds intercalated

528 within the 5.4 m long cored humic mud sequence, eight prominent beds, each 5 - 20 cm-thick,

529 are interpreted as washover deposits due to their thinning landward trend and the inclusion of

530 beach sand and marine bioclasts (Okamura and Matsuoka, 2012). Based on radiocarbon dating,

531 Okamura and Matsuoka (2012) linked the youngest three beds to the 1707, 1361 and 684 CE

532 tsunamis.

533 Baranes et al. (2016) reported a total of nine washover sand beds from the back-barrier

534 lakes facing the Bungo Channel in western Shikoku (Fig. 2, site 8). A 9.3 m-deep sediment core

535 from Lake Ryuuō, consisting of mainly clay beds and covering ~2500-year-long record of

536 deposition, shows an environmental change from a brackish lake to freshwater conditions that

537 occurred around 1000 CE (about 6 m below the lake bottom). Nine washover sand beds occur in

538 the freshwater clay between the depths of 595 and 271 cm, accumulated from 1000 to 1800 CE.

539 Although with substantial chronological uncertainties, their depth-to-age model suggests the

540 possibility that four washover sand beds represent the 1099, 1361, 1605, and 1707 CE tsunamis.

541 There are two sand layers deposited around 1500 CE, one of which may be due to a Nankai

542 earthquake that is unknown from historical records and paired with the 1498 Meiō Tōkai

543 earthquake. Baranes et al. (2016) concluded that the 1707 CE tsunami, which left the thickest

544 (25 cm) coarse-grained sand bed, was one of the most significant floods over the last

545 millennium in the region.

546 Fourteen sandy tsunami deposits ranging from 4800 to 1300 cal BP in age are reported

547 from Tadasu-ike (Fig.2, site 9), located 800 m inland from the modern coast and behind a 5 m

548 high beach ridge (Okamura and Matsuoka, 2012). The sand beds range from 5 to 27 cm in

549 thickness. The pond was inundated by the 1707 and 1946 CE tsunamis, however the record of

550 the last 1300 years is missing in this pond due to artificial disturbance; the youngest bed was

551 linked to the 684 CE tsunami (Okamura and Matsuoka, 2012). Kaniga-ike in Shikoku (Fig.2,

552 site 10), located about 400 m inland from the modern coast and fringed by 6 m-high coastal

553 dunes, records six sandy tsunami deposits in the last 2000 years (Okamura and Matsuoka, 2012).

554 The tsunami beds range from 5 to 75 cm in thickness. Based on historical documents and

555 limited radiocarbon ages (Cabinet Office, 2011a), the youngest two can be correlated with the

556 1854 CE Ansei-Nankai and 1707 CE Hōei earthquakes, while the third and fourth ones from the

557 top may be attributed to the 1361 and 684 CE Nankai earthquakes, respectively.

558

559 **Prehistoric tsunami deposits**

560   Tsunami deposits before the oldest documented Nankai Trough earthquake, the 684 CE
561   Hakuho earthquake, have been recorded as far back as 6000 years ago. Kitamura and Kobayashi
562   (2014a) reported four possible tsunami sand beds from a muddy shallow bay sequence in the
563   Shimizu Plain (Fig. 2, site 29), which are dated to 6180–6010 to 5700–5580, 5700–5580 to
564   5520–5320, 4335–4125 to 4250–4067, and 3670–3540 to 3500–3360 cal. BP, respectively.
565   Kitamura et al. (2013) identified two possible tsunami sand beds from a back-marsh sequence in
566   the Shizuoka Plain covering over the past 4000 years (Fig. 2, site 28).

567   Drowned valleys on the Shima Peninsula archive a long and numerous record of possible
568   paleotsunamis. Okahashi et al. (2005) reported up to 12 washover beds deposited from 6,000 to
569   1,600 cal BP at Osatsu (Fig. 2, site 18). At Shima (Fig. 2, site 17), Fujino et al. (2018) reported
570   seven washover sand beds beneath the three historical tsunami beds. Their depth-to-age model,
571   based on radiocarbon dates, indicates that these ten tsunami beds were deposited from about
572   4,500 to 500 cal BP, with an interval of 100 to 600 years.

573   Lake Hamana, a large brackish lagoon on the Enshu-nada coast (Fig. 2, site 23), has also
574   been targeted for paleotsunami studies by several research groups. Tsuji et al. (1998) analyzed
575   cores of up to 2 m in length from the flood-tide delta and interpreted eight exotic coarse-grained
576   beds with gravel and marine shells as possible tsunami deposits. The four older deposits formed
577   between about 4,800 and 3,200 cal BP; storm surges and tidal channel migration also remain
578   possible explanations for them. The four youngest deposits are linked with tsunami inundation
579   in 1854 or 1707, 1498 CE, the 13th century and 1096 CE by Tsuji et al. (1998); however, more
580   recent recalibration of radiocarbon data highlights difficulties with ascribing particular historical
581   tsunamis to these deposits (Garrett et al., 2016). Sato et al. (2016b) identified a marine incursion
582   event around 4790 – 4420 cal. BP based on fossil diatom analyses of a sediment core obtained
583   from the central Lake Hamana.

584   The western Hamamatsu Plain (Fig. 2, site 24) has a maximum width of about 4 km and
585   six rows of beach ridges (Fig. 4A). Possible tsunami sand beds intercalated with mud and peat
586   were reported from two separate swales (Sato et al., 2016a) and an incised valley (Fujiwara et
587   al., 2013a). Sato et al. (2016a) does not state that the sand beds are tsunami deposits, but
588   additional evidence suggests a possible tsunami origin for them. Each sand bed, generally less
589   than 20 cm thick, shows an erosive base and upward fining trend, and records the occurrence of
590   sediment flow in the back-marsh. A rapid increase in marine and brackish diatoms recognized
591   within and just after the deposition of some of the sand beds and suggests coastal subsidence
592   coincident with the marine incursion (Fig. 4D). The uppermost tsunami deposits are
593   progressively younger with decreasing distance from the present coastline. In a
594   landward swale (Fig. 4B) and in the Rokkengawa lowland (Fig. 4E), 4 km inland from the
595   modern coast, the uppermost deposit dates to just before the fall of the Kawagodaira pumice

596 (1210-1187 cal BCE, Tani et al., 2013). In a more seaward swale, 2 km from the present coast

597 (Fig. 4C), the uppermost deposit dates to 775-895 CE.

598   In Shikoku and Kyushu, five tsunami deposits have been recognized from Ryujin-ike (Fig.

599 2, site 7) between 3300 and 1600 years ago, thirteen from Tadasu-ike (Fig. 1, site 9) between

600 3300 and 1600 years ago, and two from Kaniga-ike (Fig. 2, site 10) between 1000 BCE and 600

601 CE (Okamura and Matsuoka, 2012). Tanigawa et al. (2018) reported four sandy washover

602 deposits tapering landward from sediment cores drilled in a small coastal lowland in Nankoku,

603 Kochi Prefecture (Fig. 2, site 11). The lowland is separated from the Pacific by dunes 450 m

604 wide and 13 m high. The sand beds are laterally extensive through marine-brackish clay and

605 overlying freshwater clay and peat. Tanigawa et al. (2018) interpreted the sand beds as having

606 been deposited by tsunamis or unusually large storm surges. The possible tsunami sand beds

607 were found in sediments with ages between 5970 and 2440 cal BP. They attributed the lack of

608 marine incursions after 2440 cal BP to the development of beach ridges that protected the site

609 from tsunamis and storm surges.

610   Shimada et al. (2019) reported five tsunami inundations between 5581 and 3640 cal BP

611 from a small drowned valley in Mugi (Fig. 2, site 12), eastern Shikoku. Array coring along

612 the valley axis and diatom analyses of the cored sediments revealed the deposition of at least

613 nine exotic sand and gravel beds in the muddy brackish-to-fresh wetland sequence.

614 Environmental changes inferred from diatom assemblages suggest the occurrence of coastal

615 subsidence associated with the deposition of three of the sand beds. Subsidence of 0.3 m or

616 more was reported around the study site during the 1946 CE Showa-Nankai earthquake (Japan

617 Coast Guard, 1948). Shimada et al. (2019) tentatively attributed two other sand beds to tsunami

618 events, while acknowledging the possibility of deposition from a storm surge or far-field

619 tsunami. Reported tsunami deposits can be traced up to 240 m or more inland from the current

620 shoreline, and they range in thickness from 5 to 20 cm with some exceptions up to 75 cm.

621

622 **3.2. Transitional boundary of Nankai Trough and Ryukyu Trench (Hyūga-nada)**

623 **3.2.1. Historical earthquakes and tsunamis**

624   The Hyūga-nada region covers the western end of Nankai Trough and the northern end of

625 Ryukyu Trench. Seismic activity in this region is relatively high and small tsunamis were

626 detected following the 1941, 1961, 1968, and 1984 CE earthquakes (Hatori, 1985). Historically,

627 large tsunamis occurred in 1662 and 1769. The 1662 CE earthquake was the largest historical

628 event, with a magnitude of 7.6 (HERP, 2004) and a maximum tsunami height of <5 m at

629 Miyazaki Plain (Hatori, 1985). Hatori (1985) estimated that the 1662 CE tsunami could have

630 inundated the lowland of Miyazaki Plain via the large rivers, while tsunamis might not have

631 been able to overtop the coastal sand dunes because of their height (~10 to 15 m, Nagaoka et al.,
632 1991).
633
3.2.2. Paleotsunami deposits

635   Paleotsunami studies, both historical and prehistoric, are very scarce in this region; a major
636 reason is probably the lack of suitable lowlands. Although the Miyazaki Plain is one candidate
637 site for paleotsunami research, the coastal dunes are too high and the rivers and floodplains are
638 large. Therefore, it is not straightforward to locate tsunami deposits in such a depositional
639 environment. Nevertheless, Ichihara et al. (2015) reported a possible tsunami deposit at around
640 2000 years ago on this plain.

641   Further south at Kushima City (Fig. 2, site 6), Yamada et al. (2020) conducted paleotsunami
642 research and found only one tsunami deposit at 4600 years ago during the period between 2000
643 and 5000 years ago. Based on this result, they suggested that tsunamigenic earthquakes
644 probably only occur at low frequencies in the Hyūga-nada region.
645
3.3. Paleotsunamis on the Ryukyu Islands
3.3.1. Historical earthquakes and tsunamis

648   Most islands in the Ryukyu Islands are surrounded by fringing reefs; these are typically
649 wider in the south of the island chain (~1.5 km wide) and narrower in the north (a few hundred
650 meters wide). Based on the geographical distribution and paleotsunami records, we divide these
651 islands into three groups: the Amami, Okinawa and Sakishima Islands. The three groups face
652 the northern, central, and southern Ryukyu Trench, respectively (Fig. 1). In between Kyushu
653 and the Amami Islands, there are several islands which we group as the Ōsumi and Tokara
654 Islands. Since no paleotsunami generated by subduction zone earthquake have been reported,
655 we do not discuss this region in this study. The Ryukyu Islands align parallel to the trench and
656 hence each island has an approximately equal chance of being affected by tsunamis if
657 earthquakes are generated along the trench.

658   In contrast to Nankai Trough region, historical records from the Ryukyu Islands are only
659 available for the last ~400 years. Historical and instrumental data show that earthquake and
660 tsunami occurrence in this region is remarkably low in frequency. Indeed, the only earthquakes
661 with magnitudes >8 during the past ~400 years were in 1911 CE in the northern Ryukyu Trench
662 (Mw=8.0, e.g., Tsuji, 1997), in 1791 CE in the central Ryukyu Trench (Mw=8.0 - 8.3,
663 Nakamura and Kinjou, 2013; Tadokoro et al., 2018), and in 1771 CE in the southern Ryukyu
664 Trench (Mw>8.0, e.g., Nakamura, 2009). However, because there is no record of ground
665 shaking accompanying the 1791 CE tsunami, it is uncertain whether this resulted from a
666 near-field tsunamigenic earthquake (Nakamura and Kinjou, 2013; Tadokoro et al. 2018) or was

667   potentially a far-field tsunami generated somewhere in the Pacific Ocean (Matu'ura, in press).

668   The 1771 CE tsunami was exceedingly large with maximum run-up heights of approx. 30 m

669   (e.g., Goto et al., 2010a). However, based on detailed historical records, it is also well known

670   that the tsunami-affected area was spatially limited to a narrow range of the Sakishima Islands

671   (Goto et al., 2010a).

672   Lowland plains suitable for paleotsunami research using sandy tsunami deposits are

673   extremely scarce because of the fringing reef environment. On the other hand, numerous coral

674   boulders are scattered on the reefs surrounding many islands and they are considered as useful

675   geological evidence of past extreme wave events such as tsunamis and storm waves (Fig. 3B,

676   e.g., Kawana and Nakata, 1994; Goto et al., 2010a).

677   In this section, we summarize paleotsunami research on the Ryukyu Islands based on the

678   comprehensive review by Goto (2017) with additional recent works. As a brief summary, the

679   paleotsunami history of the Amami and Okinawa Islands can be characterized by scarce (or

680   even no) evidence of large tsunamis over the last few thousand years, while the Sakishima

681   Islands can be characterized by frequent large tsunamis with intervals of a few hundred years.

682

683   **3.3.2. Paleotsunami deposits on the Amami and Okinawa Islands**

684   On the Amami Islands, Kawana and Nakata (2003) estimated the possible occurrence of

685   tsunamis in 360-510 cal BP, 1390-1770 cal BP, and 2100-2110 cal BP, based on dating coral

686   boulders on the reef. Similarly, Kawana (2006) and Iwai and Kawana (2008) suggested possible

687   large tsunamis around 3400 years ago on the Okinawa Islands, based on the dating of beach

688   rocks and coral boulders. However, following studies by Goto et al. (2009, 2013), field surveys

689   and remote sensing analyses revealed that all of the coral boulders deposited on the reefs

690   fringing the Amami and Okinawa Islands can be explained by transport during storms despite

691   their heavy weight (~160 tons). The boulders are deposited close to the source area (i.e. around

692   the reef edge) and have not been transported a long distance. Since the reef was formed and the

693   boulders were emplaced on the reef at least 2300 years ago (Kawana and Nakata, 2003), the

694   present clast size and spatial distribution of boulders reflects the cumulative impacts of many

695   storm waves since at least 2300 years ago. If large tsunamis had occurred during this period,

696   boulders originally deposited on the reef by storm waves should be transported further inland, as

697   is observed in the Sakishima Islands (discussed below). Therefore, Goto et al. (2013) concluded

698   that no tsunamis large enough to significantly modify the characteristic distribution of storm

699   wave boulders have affected these islands in the last 2300 years. Indeed, using numerical

700   modeling, Minamidate et al. (2020) confirmed that boulders in the Okinawa Islands can be

701   explained by the realistic size of storm waves generated by past typhoons.

702   Another important paleoseismological aspect, especially in the Amami Islands, is a rapid

703 uplift of Kikai Island (e.g., Webster et al., 1998). At Kikai Island, the terrace of MIS 5e is at an

704 elevation of 224 m (e.g., Ota and Omura, 2000). At least 4 emerged Holocene coral terraces are

705 observed and each terrace is associated with emergence of 1 to 4 m (Webster et al., 1998;

706 Sugihara et al., 2003; Hongo, 2010). Because of these geomorphological features, it is thought

707 that Kikai Island was intermittently uplifted by large earthquakes (Nakata et al., 1978). On the

708 other hand, considering the absence of tsunami boulders but presence of storm boulders both on

709 Kikai and Amami-Oshima Islands (Goto et al., 2013), earthquakes that might have caused the

710 uplift of the island may not have generated large tsunamis. The uplift might be explained by slip

711 on an intra-plate fault, such as a splay fault, rather than slip on the subduction interface (Goto,

712 2017). This idea may better fit with the fact that nearby Amami-Oshima Island (~40 km to the

713 west) has not been uplifted significantly compared with Kikai Island (Ikeda, 1977) and hence

714 these islands are tilted within a narrow area. Alternatively, Shikakura (2014) numerically

715 modeled crustal movements and the formation of the coral terraces at Kikai Island and

716 suggested that large earthquakes are not mandatory requirements, but steady uplift with some

717 intermittent earthquakes may alternatively be enough to explain the uplift as well as the

718 formation of multiple coral terraces.

719

720 **3.3.3. Paleotsunamis on the Sakishima Islands**

721   Because of historical records that describe the movement of huge boulders during the 1771

722 CE tsunami and the presence of numerous boulders deposited on the reef and on land,

723 identification of historically described 1771 CE tsunami boulders has proved an attractive topic.

724 With this objective, geological and historical studies of the 1771 CE tsunami and boulders

725 began earlier than tsunami studies in other regions, with the earliest work by Makino (1968).

726 Makino (1968, 1981) interpreted that all boulders deposited on land were transported by the

727 1771 CE tsunami. However, following studies cautioned the need for careful identification

728 because many of the boulders are composed of the Pleistocene Ryukyu Limestone that forms

729 the island itself and may have been sourced from higher elevations and moved by gravity (Kato

730 and Kimura, 1983; Kawana and Nakata, 1994). Nevertheless, there are numerous coral boulders

731 with fresh Holocene coral skeletons that must have been transported from the sea possibly by

732 large tsunamis during Holocene (Kato and Kimura, 1983; Kawana and Nakata, 1994).

733   Radiocarbon dating of coral boulders has been performed since early 1980's and boulders

734 with ages around 1771 CE have been discovered, suggesting their origin from this tsunami (e.g.,

735 Kato and Kimura, 1983; Kawana and Nakata, 1994). However, many studies have revealed that

736 there are boulders with much older ages than 1771 CE, suggesting that these boulders are likely

737   to have been deposited by repeated tsunamis during the few thousand years before 1771 CE

738   (Kawana and Nakata, 1994; Omoto, 2012; Araoka et al., 2013).

739   Problems associated with pre-2010 CE studies on these boulders include 1) discrimination

740   of tsunami boulders from storm boulders, and 2) dating accuracy and reliability of tsunami age

741   estimation. Regarding the first issue, Goto et al. (2010b, 2013) investigated clast size and the

742   spatial distribution of boulders on the Sakishima Islands and suggested that coral and reef

743   boulders can be classified into two groups. One group has a characteristic distribution of storm

744   wave boulders such as a narrow range in spatial distribution from the reef edge and an

745   exponentially fining landward trend in clast-size distribution. The boulders of this group can be

746   considered as of storm wave origin (Goto et al., 2010b, 2013). On the other hand, the other

747   group can be characterized by boulders that are deposited along the coast and on land. They are

748   deposited far beyond the possible landward limit of storm wave boulders (i.e. ~300 m from the

749   reef edge, Goto et al., 2009). Also, there is no clear clast size distribution indicative of storm

750   deposition. Based on these features, Goto et al. (2010b, 2013) identified the boulders in this

751   group as of tsunami origin. Later, Watanabe et al. (2016) numerically confirmed the tsunami

752   origin of the boulders in the latter group, identifying that extremely long wavelengths are

753   required to explain the deposition of boulders along the coast.

754   Regarding the second issue, Omoto (2012, 2019) (Fig. 2, site 3) and Araoka et al. (2013)

755   (Fig. 2, site 2) carefully selected samples for dating. They limited measurements to the boulders

756   that meet tsunami boulder identification criteria proposed by Goto et al. (2010b, 2013) and that

757   were composed of single massive corals (i.e. *Porites* sp., Fig. 3B). Based on this approach,

758   Araoka et al. (2013) estimated a tsunami recurrence interval of about 150-400 years on the

759   Sakishima Islands. It should be noted, however, that they measured ages of various sizes of

760   boulders so it is uncertain whether the paleotsunamis that cast the boulders ashore were large or

761   small.

762   There are some studies that focus on sandy tsunami deposits from the Sakishima Islands.

763   Possible sandy tsunami deposits, especially relating to 1771 CE, were previously reported by

764   archaeologists (e.g., Yamamoto, 2008; Kugai, 2011; Nakaza, 2017) and coastal engineers

765   (Nakaza et al., 2013) following the work done by Kawana and Nakata (1994) but their detail

766   history had remained uncertain. Recently, Ando et al. (2018) performed a 120 m-long trench

767   survey and found that paleotsunamis may have occurred at approximately 600-year intervals

768   over the last 2000 years (Fig. 2, site 1). This observation is consistent with numerical modeling

769   of the movement of a very large boulder at southeastern Ishigaki Island by Hisamatsu et al.

770   (2014) who suggested at least 2 large tsunamis equivalent to or even larger than the 1771

771   tsunami occurred after 2000 years ago but before 1771 CE. Interestingly, the recurrence interval

772   of paleotsunamis estimated from sandy tsunami deposits (Ando et al., 2018; Kitamura et al.,

773 2018a, b) is longer than that estimated from boulders (Araoka et al., 2013). This discrepancy
774 may be explained by coastal boulders being more sensitive in recording tsunamis because even
775 small tsunamis can transport small boulders to the coast. Therefore, evidence of small to
776 medium tsunamis are included in the estimation of tsunami recurrence intervals from boulders.
777 Meanwhile, sandy tsunami deposits that extended far inland may only be suitable to recognize
778 large events. Alternatively, this discrepancy might also reflect the small number of studies
779 carried out on sandy deposits in this region so further research is required. Although some
780 researchers have suggested the absence of a predecessor to the 1771 CE tsunami (Nakaza et al.,
781 2013, 2015), geological evidence of both boulders and sandy deposits provides crucial evidence
782 that either small or large tsunamis have repeatedly occurred over the last few thousand years
783 around the Sakishima Islands.
784
785 **4. When and where: the history of great earthquakes**
786   This section summarizes the spatial and temporal distribution of paleotsunami evidence
787 along the Nankai and Ryukyu subduction zones. Table 1 summarizes when the historical
788 earthquakes occurred in these regions.
789
790 **4.1. Nankai subduction zone**
791   Great earthquakes occurred in the Tōkai region nine times in total, including the
792 controversial 1614 CE event, from the end of the 7th century until 1944 CE. On the other hand,
793 great earthquakes occurred in the Nankai region eight times from 684 to 1946 CE. The 1099 CE
794 earthquake was recorded with an incorrect date, and may have actually occurred in 1096 CE
795 (Ishibashi, 2016). The Tōkai and Nankai earthquakes always occur in pairs within a few days to
796 several years, except for the 1498 CE Meiō earthquake. A Nankai earthquake paired with the
797 1498 CE Meiō Tōkai earthquake is likely to have occurred, but we have not been able to
798 definitively ascribe any of the multiple candidates from the historical record. The 1707 CE Hōei
799 and 887 CE Nin-na earthquakes were full-length ruptures of both the Tōkai and Nankai fault
800 segments.
801   Some questions remain about the recurrence pattern of Nankai Trough earthquakes. First,
802 the recurrence intervals of earthquakes in the Tōkai and Nankai regions appears to be different
803 before and after the 1361 CE earthquake (Table 1). Even if the 1614 CE earthquake did exist, it
804 seems unsuitable to include it within a series of great Nankai Trough earthquakes because the
805 seismic intensity and tsunami during the 1614 CE event were much smaller than other
806 earthquakes. If this earthquake is included, the recurrence interval from the 1361 CE earthquake
807 to the 1854 CE earthquake were 137 years (1361 - 1498 CE), 116 years (1498 - 1614 CE), 93
808 years (1614 - 1707 CE), and 147 years (1707 - 1854 CE). On the other hand, before 1361CE,

809 recurrence intervals were much longer; 203 years (684 - 887CE), 209 years (887 - 1096 CE),
810 and 265 years (1361 - 1096 CE). It is unlikely that the process of plate boundary earthquakes
811 suddenly changed at some point and, while it is possible that the apparent change is an artifact
812 of the small sample number, other explanations may also be considered.

813   It is possible that other great earthquakes may be missing from the documented record due
814 to a range of circumstances (Koyama, 1999). However, if M8-class earthquakes repeated at
815 intervals of ~100 years before 1361 CE, due to the large Medieval population, it would be
816 surprising if no historical documents recorded their occurrence. Reflecting the sensitivity with
817 which historical records from this period document earthquakes, in the period between 1200 and
818 1260 CE, there are over 100 earthquakes detailed in documents from Kyoto (Ishibashi, 2014).
819 Nevertheless, liquefaction features suggest the occurrence of a Tōkai earthquake between 1096
820 and 1361 CE and evidence for a Nankai earthquake between 1099 and 1361 CE has been found
821 from archaeological sites in southern Osaka and the southern Kii Peninsula (Sangawa, 2007).
822 Further searches for tsunami deposits are necessary to identify whether there are additional
823 historical Nankai Trough earthquakes other than those shown in Table 1.

824

825 **4. 2. Ryukyu subduction zone**

826   In contrast to the Nankai Trough region, there seems to be little evidence of large
827 earthquakes (Mw>8) and tsunamis from the Hyūga-nada or the northern and central Ryukyu
828 Trench regions over the last few thousand years. However, more data needs to be collected,
829 especially from the Hyūga-nada region.

830   Large tsunamis have repeatedly occurred around the southern Ryukyu Islands. If
831 submarine landslides have contributed to enhancing tsunami magnitudes (Miyazawa et al.,
832 2012; Okamura et al., 2018), it is difficult to estimate earthquake magnitudes. However, it is
833 probably reasonable to infer from the paleotsunami record that Mw>8 earthquakes probably
834 occurred, with some potentially triggering submarine landslides (Hisamatsu et al., 2014).

835

836 **5. Variation in size and recurrence time**

837   The size of earthquakes and tsunamis occurring on the Nankai and Ryukyu subduction
838 zones shows a wide variation. However, lesser tsunamis may not leave a trace in the geological
839 record. Of course, the creation and preservation potential of tsunami deposits can change
840 depending on the geological and geomorphological conditions of the place where the tsunami
841 hit. For example, intervals between 10 tsunami deposits left in the Shima Lowlands (Fig. 2, site
842 17) between 4500 and 500 years ago show variations from 100 to 600 years (Fujino et al., 2018),
843 which are longer than expected from the historical Nankai Trough record (Fig. 2). The lack of
844 intervening tsunami deposits may simply be due to the lesser size of those tsunamis. Only the

845 larger tsunamis that reached the study site with sufficient stream power would be recorded in
846 the sediment record.

847   In that sense, it is interesting that traces of the 1361 CE tsunami were not found between
848 the 1096 and 1498 CE tsunami deposits in the Ota River Lowland, where a careful investigation
849 was conducted by Fujiwara et al. (2020) on a large outcrop over a length of 1 km (see chapter
850 3.1.2). This probably indicates that the 1361 CE tsunami was too small to inundate the study
851 area, or that, if the tsunami reached the area, its stream power was too weak to form an
852 identifiable tsunami deposit.

853   To enhance the contribution of the geological record, it is necessary to expand the search
854 for paleotsunami deposits to sediments that have not been targeted so far. For example, the 2011
855 Tōhoku tsunami widely left fine-grained deposits, mainly clay beds, inland from the area where
856 sandy deposits are distributed (e.g., Goto et al., 2011; Fujiwara, 2015). If the mud supply is
857 sufficient, such a mud layer may be left behind after the tsunami. However, muddy
858 paleotsunami deposits are likely to have been overlooked in previous studies. Development of
859 criteria and techniques for identifying such muddy tsunami deposits in the geological record
860 will help to expand the range of available sites.

861   Analyses of seawater components (Chagué-Goff et al., 2017) and marine biomarkers in the
862 sediment (e.g., Shinozaki et al., 2015) are among several promising new ways to solve this
863 problem. While efforts should be made to develop approaches for identifying tsunami deposits,
864 geological methods alone have their limitations. For example, identification of the difference
865 between a tsunami and a storm surge by sediment transport modeling is also expected
866 (Watanabe et al., 2018). Although many studies have performed paleotsunami surveys on
867 low-lying plains or coastal lakes, it is important to expand the survey area to the other regions
868 such as narrow valleys (e.g., Fujiwara, 2015; Abe et al., 2020) in order to increase number of
869 potential sites in the region.

870   There are only a few "treasure troves" where many tsunami deposits are recorded. Methods
871 for accurately dating the sediments and estimating the tsunami size from such records may have
872 not been sufficiently developed. Recent advances in coring, non-destructive testing technology
873 (Falvard and Paris, 2017, Paris et al., 2020), chemical analysis methods (e.g. Chagué-Goff et al.,
874 2017) and geophysical sensing techniques (e.g., Obrocki et al., 2020) will drive the
875 development of more complete tsunami records from such key sites. Advances in radiocarbon
876 dating and calibration techniques will enable more accurate estimation of the occurrence timing
877 of great earthquakes and tsunamis (Ishizawa et al., 2020).

878   Progress in methods for quantifying the tsunami size from deposits are needed to elucidate
879 the variations in earthquake size. Most lowlands along the Nankai Trough coast are small, with
880 mountains and hills close to the coast. In such a situation, tsunamis reach the landward margin

881 of the lowland irrespective of their size, and it is not possible to compare the differences in
882 inundation distances. The tsunami height could, however, be estimated using sediment transport
883 modeling, using parameters such as grain size, the thickness of tsunami deposits, and the height
884 of coastal barrier (cf. Baranes et al., 2016). Tsunami boulders may be a useful tool for
885 estimating local tsunami size especially in the Ryukyu Islands (e.g., Watanabe et al., 2016).
886 However, the future improvement of models and development of methodologies to collect field
887 data about the boulder source and shape are required for further high accuracy modeling.

888

889 **6. How big?**

890   As the 2004 Sumatra-Andaman and 2011 Tōhoku earthquakes demonstrated, an
891 earthquake much larger than any in the region's instrumental or documentary history may occur
892 in the future. Paleoseismological evidence shows that these extraordinary events have occurred
893 repeatedly along subduction zones around the world, including the Kuril Trench (e.g.,
894 Nanayama et al., 2003), northern Japan Trench (e.g., Sawai et al., 2012; Sawai, in press),
895 Peru-Chile Trench (e.g. Cisternas et al., 2005), and Sunda-Java Trench (e,g., Malik et al., 2011).
896 These giant earthquakes have much longer (~300-500 years or longer) recurrence intervals than
897 those of great earthquakes (M8-class) that have occurred in each subduction zone. According to
898 the available paleotsunami evidence, the Nankai and Ryukyu subduction zones may not be
899 exceptions.

900

901 **6.1. Nankai subduction zone**

902   An updated fault model for the 1707 CE Hōei earthquake based on crustal deformation
903 inferred from the GPS network and evidence of tsunami deposits at Ryujin Pond (Fig. 2, site 7)
904 suggest that its source area extended 70 km further to the west of the 1854 earthquake
905 (Furumura et al., 2011). Okamura and Matsuoka (2012) suggested that earthquakes of similar
906 magnitude to the 1707 event could occur every 300-500 years along this subduction zone. The
907 number of tsunami deposits varies from site to site and they are generally less frequent than
908 expected from the frequency of historical earthquakes. This may be the result of only larger
909 tsunamis overwashing the coastal barriers and leaving deposits in back-barrier ponds. The 684,
910 1361 and 1707 CE tsunamis selectively recorded at the Ryujin, Ryuuō, and Kaniga-ike Pond
911 (Fig. 2, sites 7, 8 and 10) may therefore be relatively larger than other historical tsunamis
912 (Okamura and Matsuoka, 2012; Baranes et al., 2016). If larger tsunamis result from larger
913 earthquakes, this indicates that larger earthquakes with 300 - 500-year recurrence intervals and
914 smaller earthquakes with shorter recurrence intervals coexist in the Nankai Trough region,
915 suggesting the recurrence pattern of this region is not periodic, but may still be time predictable
916 or slip predictable (Fig. 5A).

917   Okamura and Matsuoka (2012) also pointed out that the historical period of 1300 years
918   may not be enough to consider the maximum-size earthquake and tsunami of this region. The
919   discovery of tsunami deposits thicker than those from the tsunami that followed the 1707 CE
920   Hōei earthquake, the largest in the Nankai region's history, from Tadasu-ike (Fig. 2, site 9),
921   Kaniga-ike (Fig. 2, site 10), Kamoda-oike (Fig.2, site 13) and Sugari-oike (Fig. 2, site 15), all of
922   which were aged around 2000 years ago, led Okamura and Matsuoka (2012) to hypothesize the
923   occurrence of a giant tsunami. However, it is unreasonable to discuss the size of a tsunami
924   based on the thickness of tsunami deposits alone. The thickness of different tsunami deposits
925   within a site may be influenced by the state of the tide and the time of year, which controls
926   vegetation cover and therefore sediment transport. Furthermore, changes in coastal
927   geomorphology over time may change a site's susceptibility to accumulating and preserving
928   tsunami deposits (Garrett et al., 2016). Underdeveloped low dunes and beach ridges can allow
929   large amounts of overflow, even during lesser tsunamis, resulting in the formation of thicker
930   tsunami deposits in back-barrier ponds (Fujiwara, 2015). Additionally, the estimated ages for
931   the thick tsunami bed in each of Okamura and Matsuoka's (2012) ponds has substantial
932   uncertainties of hundreds of years. Consequently, it is ambiguous whether the thick tsunami bed
933   in each of the ponds was formed by one extremely large tsunami or series of closely spaced
934   large tsunamis.

935   Deposition of a thick tsunami deposits around 2000 years ago is not seen at Nankoku (Fig.
936   2, site 11) (Tanigawa et al., 2018) or Ryuuō Pond (Fig. 2, site 8) (Baranes et al., 2016). In
937   addition, at Ryūjin Pond in northern Kyushu, a sand bed deposited around 3300 years ago is the
938   thickest, with only a minor sand bed deposited around 2000 years ago. Thick tsunami deposits
939   formed around 2000 years ago have not been found from the Tōkai region (Fujiwara, 2013;
940   Fujiwara and Tanigawa, 2017). The late 7th century tsunami deposits in the Otagawa Lowland
941   is thick, over 40 cm more than 2 km inland from the coast at the time of tsunami occurrence, but
942   its distribution is limited to the topographic lows in the lowland (Fujiwara et al., 2020).
943   Fujiwara et al. (2020) interpreted that the tsunami was not so large and inundation and sand bed
944   deposition was limited to the topographic depressions.

945   Kitamura (2016) summarized the spatio-temporal distribution of reported tsunami
946   deposits from the eastern coast of the Nankai Trough, including sites 23, 24, 26, 28, 29, and 32
947   in Figure 2, and concluded that no tsunami deposits with a wide regional distribution that might
948   suggest the occurrence of "level 2" tsunami over the past 4000 years have been found in this
949   area. However, due to the limitation in the age control of the tsunami deposits, it would be
950   difficult to deny the existence of a giant tsunami on this basis alone.

951

952   **6.2. Hyūga-nada and the Ryukyu subduction zone**

953　　　Historical and geological evidence is too scarce to evaluate whether large tsunamigenic
954　earthquakes have been generated in the Hyūga-nada region. Historical records suggest no large
955　earthquakes with magnitude >8 have occurred over the last 400 years in this region. Geological
956　evidence from the Pacific coast of central and southern Kyushu is probably still insufficient to
957　discuss paleoseismicity in this region, although recent works suggest the rare occurrence of
958　tsunamigenic earthquakes (Ichihara et al., 2015; Yamada et al., 2020).

959　　　M8-class earthquakes can be generated elsewhere along the Ryukyu Trench based on
960　historical and instrumental records; however, the Ryukyu Trench is generally considered as an
961　aseismic zone in terms of the potential occurrence of giant earthquakes of magnitude ~9,
962　although potential coupling of the plate boundary and the consequent occurrence of extremely
963　large earthquakes along the Ryukyu Trench has been suspected by some researchers (e.g., Ando
964　et al., 2009, 2012a; Aydan and Tokashiki, 2018; Hsu et al., 2012). Geological evidence shows
965　that tsunamigenic earthquakes may be rare or even nonexistent in the northern and central
966　Ryukyu Trench (Goto et al., 2013). Even though Kikai Island in the northern Ryukyu Trench
967　has been uplifted quickly and the repeated occurrence of large earthquakes has been suspected
968　(e.g., Nakata et al., 1978), relatively small earthquakes just beneath the island (Goto, 2017) or
969　stable uplift without large earthquakes (Shikakura, 2014) may also explain this record.

970　　　Nakamura and Sunagawa (2015) studied the distribution of very low frequency
971　earthquakes (VLFEs) along entire the Ryukyu Trench. They suggested that occurrence of
972　VLFEs is high in the northern and central Ryukyu Trench but low in the southern Ryukyu
973　Trench. Therefore, it is likely that the southern Ryukyu Trench has potential to generate large
974　earthquakes, unlike the northern and central regions (Nakamura and Sunagawa, 2015).
975　Nishimura (2014) also suggested that short-term slow slip events (SSEs) cluster near the areas
976　where large earthquakes have occurred historically (e.g., the 1911 and 1771 CE events).
977　Nevertheless, the sources of these large earthquakes do not overlap with the clusters of
978　short-term SSEs and the reason remains uncertain.

979　　　Arai et al. (2016) and Nakamura (2017) suggested there is a strongly coupled zone at
980　shallower plate boundary depths in the southern Ryukyu Trench. Therefore, localized
981　tsunamigenic earthquakes at the southern Ryukyu Trench inferred from tsunami boulders (Goto
982　et al., 2013) may be explained by the heterogeneity of the coupling along the Ryukyu Trench.
983　Nevertheless, the source model of the 1771 CE tsunami is still controversial with proposed
984　models including an intra-plate earthquake plus a submarine landslide (Imamura et al., 2008;
985　Miyazawa et al., 2012), a tsunami earthquake (Nakamura, 2009), splay faulting (Hsu et al.,
986　2013), and a large submarine landslide near the trench axis (Okamura et al., 2018). By assuming
987　a large subduction zone earthquake along the Ryukyu Trench, Ando et al. (2018) estimated a
988　seismic coupling coefficient of 20%. Alternatively, if a submarine landslide played an important

989 role in the generation of the 1771 CE tsunami and predecessors, then a special tectonic state
990 capable of generating large earthquakes only at the southern Ryukyu Trench may not need to be
991 considered. Relationships among VLFEs, short-term SSEs, and large earthquakes (Mw ~ 8) are
992 still uncertain, so further seismological research with reference to paleotsunami results is
993 required along the Ryukyu Trench.

994

995 **7. Knowledge gaps and future work**

996

997 **7.1. Earthquake recurrence model**

998 Since Kanamori and McNally (1982) explained a series of different size earthquakes off
999 the Colombian coast as the rupture of different combinations of multiple asperities, similar
1000 phenomena have been reported from subduction zones around the world. Some examples show
1001 that giant (M9-class) and large (M7-8 class) earthquakes may share fault plane asperities, and
1002 the former have much longer recurrence intervals than the latter (Cisternas et al., 2005, 2017).
1003 The Nankai and Ryukyu subduction zones may be no exception. This may indicate that the
1004 recurrence pattern of subduction zone earthquakes is "nonpredictable" (Fig.5A).

1005 The recurrence of Nankai Trough earthquakes has generally been explained as the repeated
1006 rupture of all or part of characteristic fault planes (segments) A, B, C, D and E (Fig. 1 and Table
1007 1). Observations that the amount of uplift during the 1707, 1854, and 1946 CE Nankai
1008 earthquakes at Cape Muroto was proportional to the length of the interval preceding each
1009 earthquake led to the proposal of the time-predictable recurrence model (Fig. 5A) by Shimazaki
1010 and Nakata (1980). Kumagai (1996) applied this model to Nankai Trough earthquakes using the
1011 time series of great earthquakes since 684 CE. However, according to current knowledge,
1012 Kumagai's series of historical earthquakes contains some uncertainties, and the reliability of the
1013 recurrence model is ambiguous. For example, the 1230 CE earthquake and the 1498 CE Meiō
1014 earthquake in the Nankai region (as opposed to the undisputed Tōkai event) in his series are still
1015 undetermined. Results of paleotsunami research in the Otagawa Lowland (Fujiwara et al., 2020)
1016 does not support the time-predictable recurrence model. According to this study, the 1361 CE
1017 Tōkai earthquake has the longest pre-earthquake interval (265 years following the 1096 CE
1018 earthquake), but its tsunami seems to have been less extensive than others, suggesting a smaller
1019 magnitude of earthquake (see chapter 5).

1020 Seno (2012) proposed a new earthquake recurrence model for Nankai Trough earthquakes
1021 that challenged the traditional idea of the characteristic earthquake model, which partitions the
1022 Nankai Trough seismogenic zone into ruptures A-E and assign the epicenters of historical
1023 earthquakes to them. He assumed two complementary earthquakes that have different
1024 non-overlapping asperities mainly based on the spatial distribution of seismic intensities, while

1025 also considering tsunami heights and coastal deformation estimated from documented records
1026 and liquefaction features in archaeological sites (Fig. 5B). According to his idea, historical great
1027 Nankai Trough earthquakes are classified into two types; "Hōei-type" earthquakes (887, 1361,
1028 1707 and 1944+1946 CE) and the "Ansei-type" earthquakes (684, 1096+1099, 1489,
1029 1854+1854 CE), which alternately occur with recurrence intervals of around 350 and 400 years,
1030 respectively. Seno's (2012) model seems to explain the variations in the recurrence intervals
1031 and extent of source area of each earthquake in Nankai Trough; however, the reliability of
1032 model is reduced by the decrease in the number of reliable documentary records with increasing
1033 age, especially before the 1096 CE earthquake.

1034 In contrast to the Nankai Trough region, the establishment of an earthquake recurrence
1035 model along the Ryukyu Trench is not straightforward because of limited historical and
1036 geological data. Presently available data suggest that earthquakes with magnitudes of around 8
1037 have occurred patchily in space and time from the northern to the central Ryukyu Trench. The
1038 repeated occurrence of these earthquakes is uncertain. On the other hand, in the southern
1039 Ryukyu Trench, the repeated occurrence of large tsunamis is suggested (Omoto, 2012; Araoka
1040 et al., 2013; Ando et al., 2018). Therefore, the future occurrence of large earthquakes in this
1041 region is likely.

1042 The amount of reported paleo-tsunami data is still insufficient to update the earthquake
1043 recurrence model in these regions (Figure 2). Updating the model also requires improved
1044 accuracy in the dating and the identification of tsunami deposits, which is a particularly
1045 important issue with sites surveyed in the early stages of the development of the field.
1046 Re-examination of these sites with current knowledge and technology, as discussed in chapter 5,
1047 may reveal some rules in the recurrence mode of the earthquakes in these regions.

1048 Coastal deformation data will enhance the revision of earthquake recurrence models (e.g.,
1049 Garrett et al, 2016). In particular, uplift along the western Suruga Bay coast, including Cape
1050 Omaezaki (representing the rupture of segment E), and the southern Kii Peninsula (representing
1051 the rupture of segment C) is key to proving the hypothesis of Ansei- and Hōei-type earthquakes
1052 (Fig. 5B). Kitamura et al. (2019) suggested a possible seismic uplift occurred around 400 CE
1053 from the Shimizu Plain (Fig. 2, site 29).

1054 Integration of offshore paleoseismic features, such as turbidite sequences, with onshore
1055 tsunami deposits also will help us to consider which recurrence model is plausible. In fact, the
1056 integrated study of coastal deformation and onshore tsunami deposits, and offshore
1057 seismo-turbidite sequences in the Cascadia subduction zone and northern San Andreas fault
1058 systems has revealed the recurrence mode of great earthquakes in this region (e.g., Goldfinger et
1059 al., 2003 a, b). These studies also contribute to regional earthquake and tsunami
1060 countermeasures through underpinning tsunami simulations (e.g., Priest et al., 2017). Although

there are some initial studies on seismo-turbidites in the Nankai (Garrett et al., 2016 and references therein) and the Ryukyu subduction zone (e.g., Ujiie et. al. 1997; Ikehara et al., 2017), at present, coordination of onshore and offshore paleoseismic research is an area where future progress is expected.

7.2. Maximum possible earthquake and tsunami

As a response to the occurrence of unexpectedly huge 2011 Tōhoku earthquake, the Cabinet Office (2011b) attempted to quantify the maximum possible earthquake and tsunami for the Nankai Trough subduction zone ("maximum scenario" in Fig. 1). The earthquake source consists of the strongly coupled zone of 10 to 20 km depth, which has been considered to be the main seismogenic zone, with the addition of down-dip weakly coupled zones at ~30 to 40 km depth, which are the location of the occurrence of SSE and deep low-frequency tremor. Its western end is delimited by the Kyushu-Palau ridge. The up-dip zone from 10 km to the trough axis, which may also generate large tsunamis, is assumed as an additional tsunami source area. According to the worst scenario by the Cabinet Office (2012a), the height of the tsunami occurring from this M9 earthquake could exceed 10 m over a wide area along the Pacific coast of western Japan, and reach in excess of 30 m in some places. However, no geological evidence, such as tsunami deposits, was taken into account for this catastrophic earthquake and tsunami scenario (e.g., Wallis et al., 2018). Determining when the most recent outsize earthquake and tsunami occurred and how large it was will help to address this concern.

Largest tsunami in the area estimated from geological data

The "2 ka giant tsunami hypothesis" by Okamura and Matsuoka (2012), derived from the discovery of extremely thick tsunami deposits, played an important role in making people aware of the possibility of a giant earthquake and tsunami in the Nankai Trough region and raising awareness of disaster management. Nevertheless, it has not been tested. First, it is necessary to ensure that multiple seismic events are not confused with a single major event. To solve this issue, the extremely thick tsunami deposits should be re-examined by increasing the number of reliable radiocarbon ages and using new age modeling methods (e.g., Bronk Ramsey, 2009; Lougheed and Obrochta, 2016, 2019).

Other causes of the thick tsunami deposit in individual ponds should also be considered. Earthquake-induced submarine landslides are a candidate source for locally thick tsunami deposits. They generate much larger, but regionally limited, tsunamis than would be expected based on the size of concurrent earthquakes (e.g., Tappin et al., 2008; Harbitz et al., 2013; Kawamura et al., 2017). In fact, many submarine landslides have been found on the submarine slopes along the Nankai Trough and some of them have areas of $\geqq 10$ km^2 (Moriki et al., 2017).

1097   The study of the evolutionary history of both coastal and submarine landforms may be the key
1098   to solving this problem.

1099   Reconstruction of the inundation area is the most important indicator for determining the
1100   size of a tsunami from its deposit (e.g., Fujiwara, 2015). In the case of eastern Hokkaido, the
1101   discovery of prehistoric tsunami deposits showing inundation areas several times larger than
1102   those of historical M8-class earthquakes revealed that unusually large tsunamis repeatedly
1103   occurred on the Kuril Trench (e.g., Nanayama et al., 2003). Similarly, the 869 CE Jōgan
1104   tsunami deposit showed much larger inundation areas than those of other historical earthquakes
1105   that occurred in the northern Japan Trench, suggesting an unprecedentedly large earthquake
1106   (Satake et al., 2008; Namegaya et al., 2010, Sawai et al., 2012).

1107   The Hamamatsu Plain (Fig. 2, site 24) is the first candidate site where we can suggest that
1108   no giant tsunami has occurred in the eastern Nankai Trough over the latter half of the Holocene
1109   based on the tsunami inundation area shown by the tsunami deposits. The strand plain has
1110   prograded seaward over the last 6000 years following the Holocene sea-level high stand (e.g.,
1111   Sato et al., 2011, 2016a), as evidenced by preservation of rows of sandy beach ridges. This
1112   progradation process can be seen in the vertical change of diatom assemblages in swale deposits,
1113   which shows a change from brackish-marine to freshwater conditions (Fig. 4B). In harmony
1114   with the progradation of the coast, older washover deposits, probably including tsunami
1115   deposits, are distributed inland and younger ones are found in more seaward locations (Fig. 4 B,
1116   C). Additionally, at each site, younger tsunami deposits tend to be thinner (Fig. 4 B, C).
1117   Assuming tsunamis (or other washover events) of similar size over time, the progradation of the
1118   coast would make it less likely that each successive tsunami would reach the inland areas, and
1119   the observed sequence of tsunami deposits would be left on the Hamamatsu Plain (Fig. 6). If an
1120   extremely large tsunami occurred over the past few thousand years, its deposit would have
1121   extended from the coast to further inland, as shown at the top of the Fig. 6. However, no such
1122   trace has yet been found (Fujiwara, 2013; Fujiwara and Tanigawa, 2017).

1123   Along the Ryukyu subduction zone, the 1771 CE tsunami at the Sakishima Islands is the
1124   largest known event. Based on numerical modeling of boulder transport Hisamatsu et al. (2014)
1125   suggested that at least one prehistoric tsunami event was even larger than the 1771 CE tsunami.
1126   However, since information is limited, further geological data is required to clarify the size of
1127   prehistoric tsunamis.

1128

1129   **Crustal movements suggesting outsized earthquakes along the Nankai Trough**
1130   In discussing the possibility of a giant earthquake, it is necessary to assess a wide range of
1131   crustal deformation patterns occurring at different temporal and spatial scales. In the case of
1132   eastern Hokkaido, where the recurrence of giant tsunami along the Kuril Trench is shown by

1133 tsunami deposits, there is a contradiction between the coastal deformation pattern shown by
1134 geological and geodetic data. In this area, 60 to 80 km above the top of the subducting plate,
1135 marine terraces showing 20 to 50 m of net uplift in the past 125 ka (Okumura, 1996) coexist
1136 with tide gauge and bench mark-inferred subsidence of up to 1 m in the past 100 years
1137 (Shimazaki, 1974a; Ozawa et al., 1997). Little, if any, uplift is known from the M8-class
1138 earthquakes in this region's written history during the past 200 years (Shimazaki, 1974b;
1139 Kasahara, 1975; Kasahara and Kato, 1980; Yamanaka and Kikuchi, 2003).

1140 Micropaleontological research on marsh deposits in this area (Sawai et al., 2004) revealed
1141 post-seismic uplift associated with unusually large earthquakes, which was induced by greater
1142 creep further down the plate boundary than during any of the Kuril subduction zone's historical
1143 events. The last uplift event of up to 2 m, which might resolve some of the contradiction
1144 between geologic and geodetic data, occurred over the first few decades following an outsized
1145 17th-century earthquake.

1146 A similar phenomenon can be seen around Lake Hamana. Middle Pleistocene (MIS 9 to
1147 MIS 5) terraces, mainly of marine and partly of fan origin, are found in this area (Sugiyama,
1148 1991; Nakashima et al., 2008), showing that uplift has been dominant over 10^4 to 10^5-year
1149 scales. The MIS-9 marine terrace reaches 78 m in elevation and is highest on the west coast of
1150 Lake Hamana, decreasing in height to the north and west. Subsidence of 0.1-0.6 m during the
1151 1854 CE Ansei-Tōkai earthquake is estimated around the Lake Hamana area (Ishibashi, 1981;
1152 Sato and Fujiwara, 2017). The northern coast of the lake is also known to have subsided during
1153 the 1707 CE Hōei earthquake (Yata, 2013). Coseismic subsidence between 7000 and 5700 years
1154 ago, similar to the 1707 and 1854 CE events, was reported from Shinjo Lowland (Fig. 2, site 21)
1155 on the west coast of Lake Hamana (Sato and Fujiwara, 2017). Multiple active reverse faults
1156 (The research group for active submarine faults off Tokai, 1999) and normal faults (Arai et al.,
1157 2006) parallel to the trough axis are distributed off the Hamana area. However, they cannot
1158 explain the coastal deformation in the area.

1159 Major differences between this area and eastern Hokkaido include the inter-seismic uplift
1160 after the 1854 CE Ansei-Tōkai earthquake and the short-term uplift associated with SSEs.
1161 Inter-seismic uplift, averaging ~6 mm/y between 1901 and 2008 CE (Geographical Survey
1162 Institute, 2008), may compensate for the coseismic subsidence during "ordinal" Tōkai
1163 earthquakes; however, other factors are needed to explain the formation of high and tilted
1164 terraces. This average uplift rate includes the effect of SSEs (see chapter 2.2). SSEs seem to
1165 occur with decadal recurrence intervals and are high frequency uplift events on a geological
1166 time scale, but associated uplift seems too small to form the terraces. For example, a Tōkai slow
1167 slip event (Ozawa et al., 2002; Miyazaki et al., 2006) occurred under Lake Hamana from
1168 January 2000 to July 2005, with a cumulative moment magnitude of ~7.1, and generated ~5 cm

1169 uplift in this area (Suito and Ozawa, 2009). At present the importance of SSEs for terrace
1170 formation remains unclear and their contribution may need to considered further.

1171 To elucidate the possibility of a giant earthquake, a unified explanation is required for
1172 geodetic and geologic phenomena on different timescales, including SSEs with 10^0 to 10^1-year
1173 scales, inter-seismic uplift with a 10^2-year scales, and tilting uplift of the Pleistocene terraces
1174 over 10^4 to 10^5-year scales. This requires further investigation of Holocene crustal movements
1175 on a 10^3-year scales, which connects the two.

1176 The possibility of eastward rupture propagation from the Nankai Trough to the
1177 Fujikawa-kako fault zone (FKZ) must also be assessed in the context of defining the maximum
1178 possible earthquake size. The FKZ is an eastern extension of the Nankai or Suruga Trough and
1179 might rupture in combination with future great Nankai Trough earthquakes (HERP, 2010).
1180 Nevertheless, the recurrence history is still unclear at present. Marsh deposits on the
1181 Ukishima-gahara Lowlands (Fig. 2, site 30) record a 1500-year history of repeated coastal
1182 subsidence suggesting the activity of the FKZ (Fujiwara et al., 2016). Further research is needed
1183 to elucidate the relationship between episodes of coastal subsidence and historical earthquakes.

1184

1185 **Possibility of full-length rupturing event of the Nankai and Ryukyu subduction zone**
1186 The Nankai and the Ryukyu subduction zones make up one plate boundary through the
1187 Hyūga-nada region, and both may have the potential to generate M9-class earthquakes.
1188 Therefore, it is important to understand the possible risk of the occurrence of extremely large
1189 earthquakes crossing this boundary, as suggested by some researchers (e.g., Furumoto and Ando,
1190 2009). As reviewed in this paper, the occurrence of such an extremely large earthquake and
1191 tsunami in the past few thousand years is not supported by the presently available historical and
1192 geological data in this region. However, the spatio-temporal correlation of earthquakes that
1193 occurred on the Nankai and Ryukyu subduction zones or the possibility of a full-length
1194 mega-earthquake linking both seismogenic zones has not been evaluated well. From this point
1195 of view, research integrating various scientific fields such as seismology, paleoseismology,
1196 geology, and geodesy is also an issue for the future.

1197

1198 8. **Conclusions**
1199 This review illustrates the difference in the type of tsunami deposits and the progress of
1200 research between the Nankai and Ryukyu subduction zones. We also summarize the recurrence
1201 mode of great earthquakes and tsunamis in these regions based on geological evidence. The
1202 volume of information currently available about paleoearthquakes and tsunamis varies greatly
1203 different between the two subduction zones, despite the fact they comprise a contiguous
1204 northwestern boundary of the subducting Philippine Sea Plate beneath the Japanese archipelago.

1205 Information from the Ryukyu subduction zone is very limited compared to the Nankai
1206 subduction zone.

1207 The model describing the recurrence mode of great earthquakes in this region is still
1208 incomplete, and updating the model requires future studies to gather more paleoseismic data
1209 from both on land and offshore. Reconstruction of the largest event ever in the region based on
1210 geological evidence will help discussions about the "real" size of the maximum possible
1211 earthquake in this region. This requires a new conceptual model that can unify geological
1212 processes that occur on various time scales. Differentiation of extremely large earthquakes and
1213 piecemeal ruptures of lesser magnitude in the geological record is also important for the
1214 accurate reconstruction of past events and corresponding forecasts of future earthquake.
1215 Progress in age estimation techniques will help to diminish this enduring problem. The potential
1216 of giant earthquakes that rupture both the Nankai and Ryukyu subduction zones remains
1217 unknown. Paleoseismological studies in Hyūga-nada region that connects the two subduction
1218 zones are key to solving this question.

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1939

1940   *English translation from the original written in Japanese.

1941

1942

1943   **Figure captions**
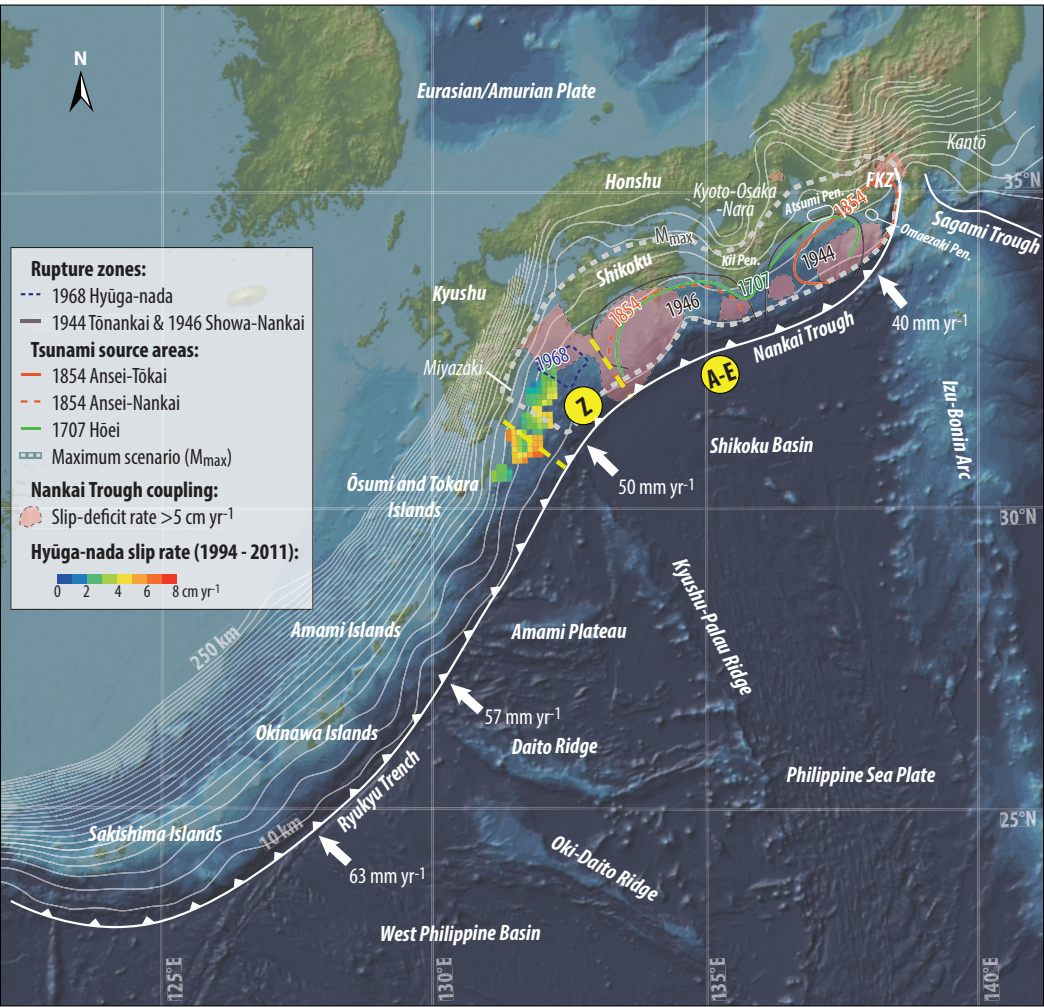
1944

1945   **Fig. 1. Map showing the outline of the seismotectonic setting around the Nankai and**
1946   **Ryukyu subduction zones.**

1947   Tsunami source areas for the 1707 CE (green line) and two 1854 CE earthquakes (orange
1948   solid and dashed lines), as modified from Hatori (1974, 1976). Earthquake rupture areas of the
1949   1944 CE Tōnankai and 1946 CE Showa-Nankai earthquakes (blue shaded areas) are simplified
1950   from Tanioka and Satake (2001a, b). Plate movement vector from Zang et al. (2002).
1951   Classification of rupture zones, A to E and Z, follows Ando, 1975 (a, b), Ishibashi, 1976, and
1952   Wells et al. (2003). Contour lines showing the depth of the Philippine Sea plate surface, 10
1953   km-interval, modified from Saito (2017). Distribution of high interpolate slip-deficit rate area
1954   (>5 cm yr^{-1}) along the Nankai Trough is simplified from Yokota et al. (2016). Spatial
1955   distribution of the averaged quasi-static slip rate in the Hyūga-nada region (from May 1994 to
1956   May 2011) is reproduced from Fig. 2 b of Yamashita et al. (2012) with the permission of
1957   publisher (John Wiley and Sons; license number 4775371178223). Source area of "the

maximum possible large earthquake" in Nankai subduction zone modified from Cabinet Office (2011b).
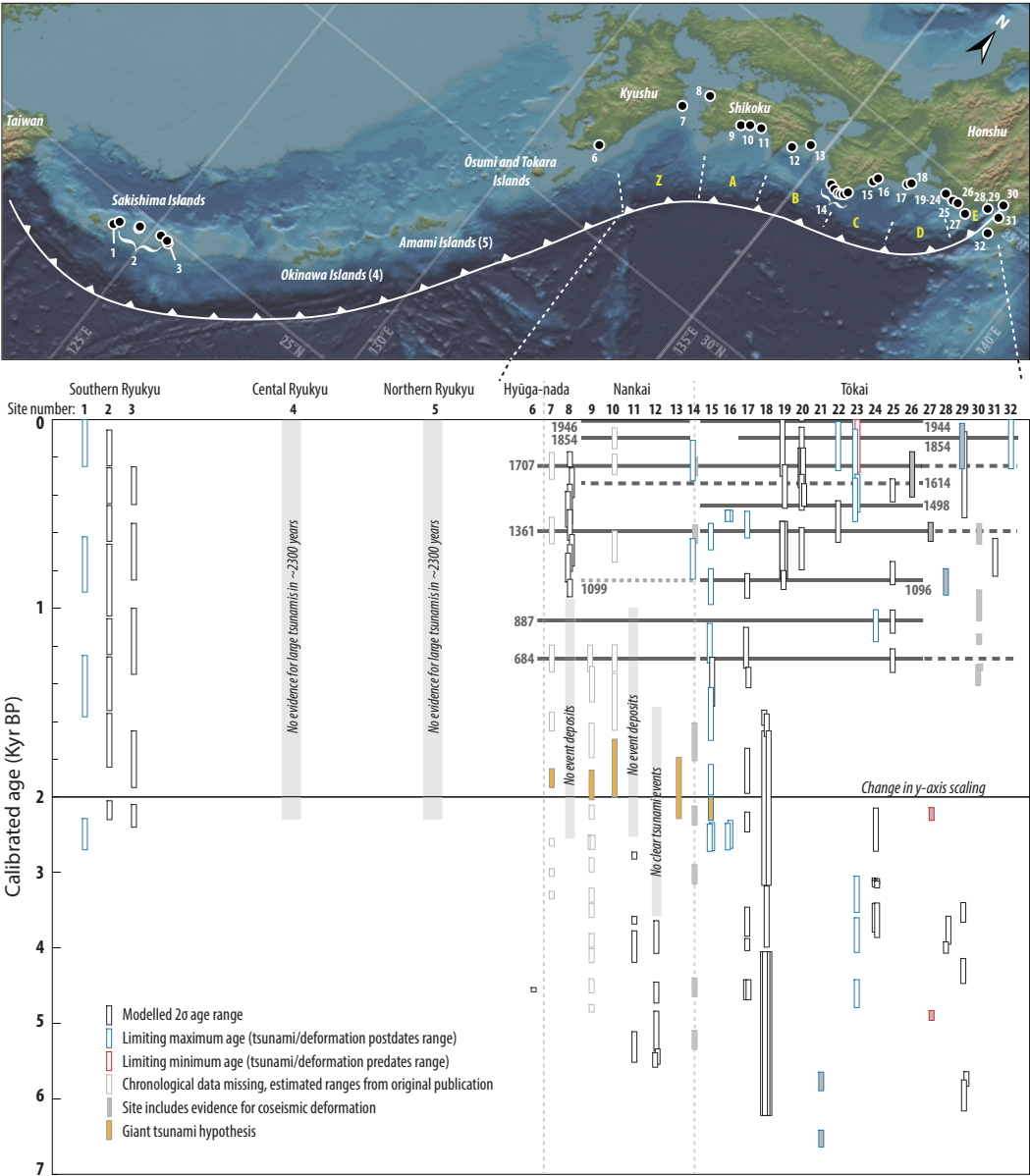
Fig. 2. Spatiotemporal compilation of representative paleotsunami evidence along the Ryukyu and Nankai subduction zones

Horizontal lines indicate Nankai megathrust earthquake rupture zones, following Fujiwara et al. (2020); dashed where uncertain, dotted to indicate the debated 1099 CE earthquake. We include four sites that record evidence for coseismic deformation (numbers 14, 21, 27 and 30).

Site numbers: 1. Ibaruma, Ishigaki Island; 2. Sakishima Islands; 3. Miyako Islands[†]; 4. Okinawa Islands; 5. Amami Islands; 6. Kushima City[†]; 7. Ryūjin-ike; 8. Ryuuoo-ike; 9.

1969 Tadasu-ike*; 10. Kaniga-ike; 11. Nankoku†; 12. Mugi Town†; 13. Kamoda-oike; 14. Kii
1970 Peninsula* (including Kuchiwabuka, Ameshima, Shionomisaki, Izumozaki, Arafunezaki,
1971 Ikeshima, Yamamibana, Taiji, and Suzushima); 15. Ōike (Sagari) Pond*; 16. Suwa Pond*; 17.
1972 Shima Lowlands†; 18. Ōsatsu Town*; 19. Nagaya Moto-Yashiki*; 20. Shirasuka; 21. Shinjo
1973 Lowlands†; 22. Arai*; 23. Lake Hamana*; 24. Western Hamamatsu strand plain†; 25. Ōtagawa
1974 Lowlands†; 26. Yokosuka Lowlands*; 27. Omaezaki†; 28. Oya Lowlands*; 29. Shimizu Plain*;
1975 30. Ukishima-ga-hara*; 31. Ita Lowlands*; 32. Shimoda*.

1976 Sites with age recalibrated or modelled in Garrett et al. (2016) marked *, sites with age
1977 ranges recalibrated or modelled in this publication marked †. We modeled age ranges using the
1978 OxCal program v.4.2 (Bronk Ramsay, 2009) using P_Sequence and Sequence models (Bronk
1979 Ramsay, 1995, 2008, 2009), following the approach detailed in Garrett et al. (2016). For the
1980 Miyako Islands, we interpreted peaks in a cumulative probability distribution as indicating the
1981 timing of boulder movement events, following the approach of Araoka et al. (2013).

1982 Basemap uses Global Multi-Resolution Topography (Ryan et al., 2009) through
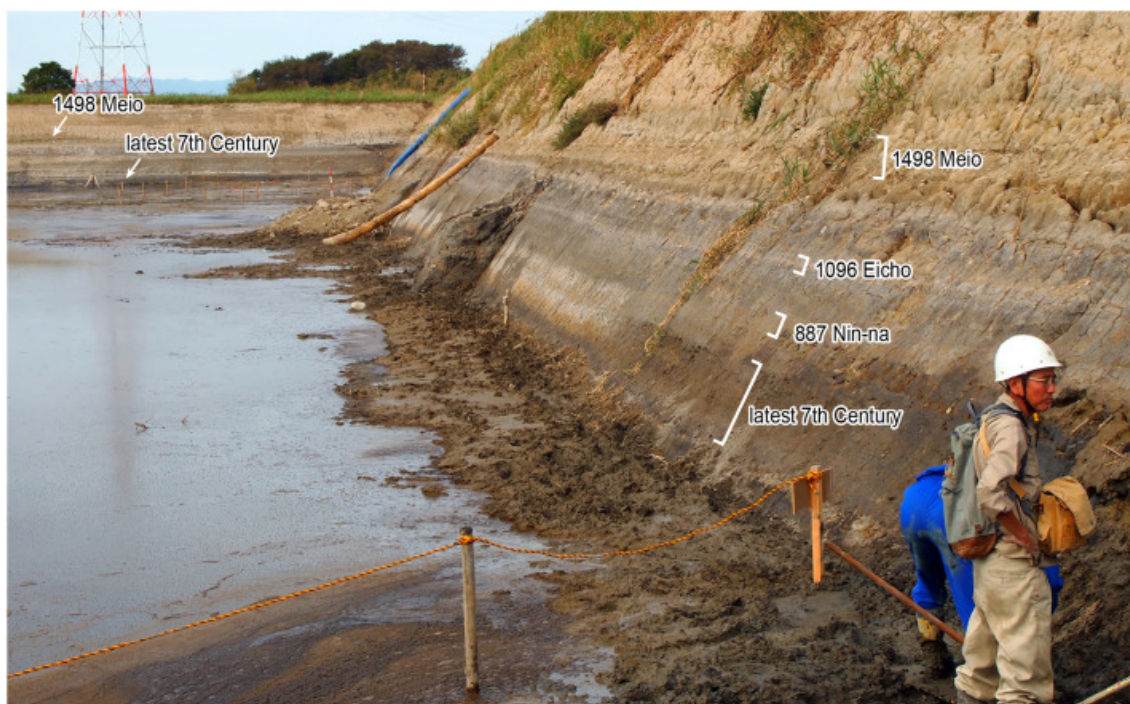1983 www.GeoMapApp.org.

1984

1985 Main references cited in Garrett et al. (2016) but not discussed further in this paper are as
1986 follows.
1987 Site 14: Shishikura et al. (2008a), Site 15, 16: Tsuji et al., (2002), Site 19: Takada et al., 2002,
1988 Site 22: Fujiwara et al. (2013b), Site 26: Fujiwara et al. 2007, Site 28: Kitamura et al. 2013),
1989 Site 32: Kitamura et al. (2014).
1990

1991

1992

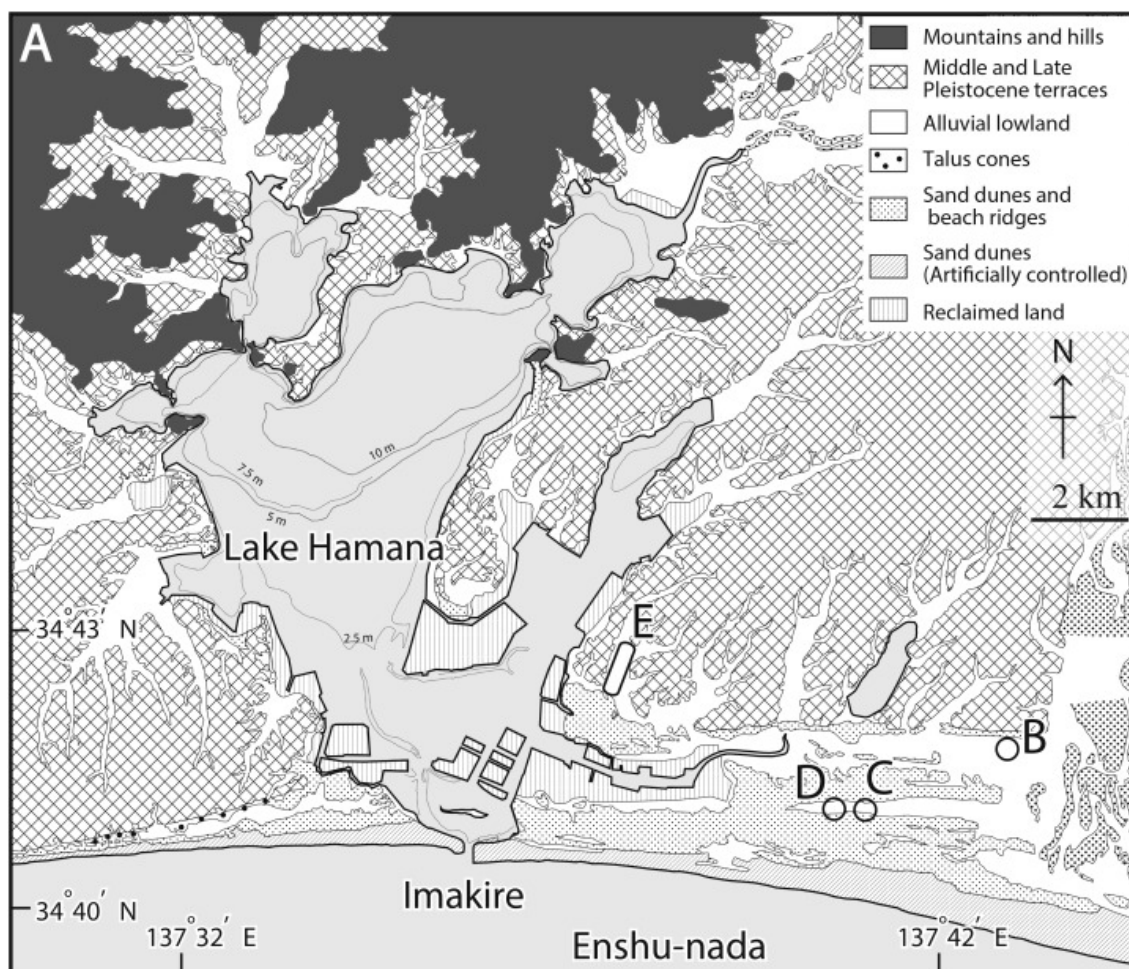1993 **Fig. 3. Photographs showing examples of tsunami deposits**

1994 A) Overview of four historical tsunami deposits at the Ōtagawa Lowlands, facing the Tōkai

1995 segment of the Nankai Trough. Stratified sandy tsunami deposits are distinguished as

1996 horizontal color bands, which serve as an aquifer, in back-marsh mud.
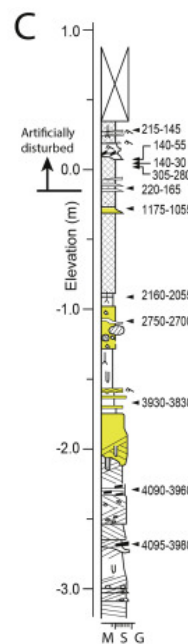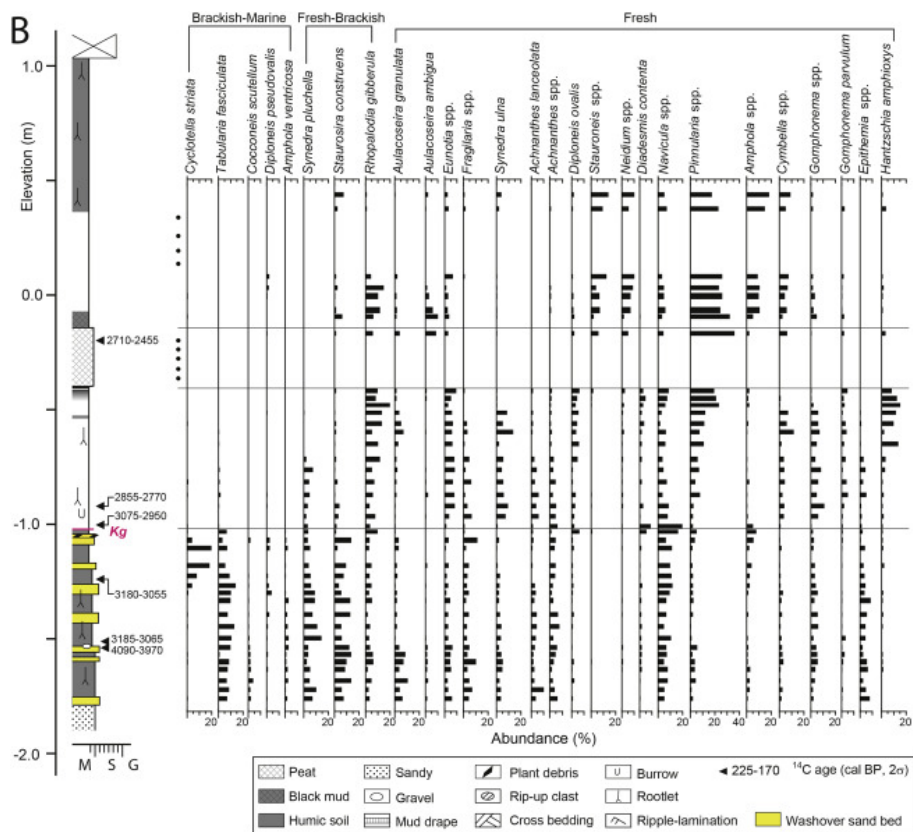
1997

1998   B) A *Porites* tsunami boulder with microatoll shape deposited on the eastern coast of Ishigaki
1999   Island, Japan. The scale at lower right is 30 cm.
2000

2001

Legend:
- Mountains and hills
- Middle and Late Pleistocene terraces
- Alluvial lowland
- Talus cones
- Sand dunes and beach ridges
- Sand dunes (Artificially controlled)
- Reclaimed land

N

2 km

Lake Hamana

10 m
7.5 m
5 m
2.5 m

34° 43′ N

E

B
D C

Imakire

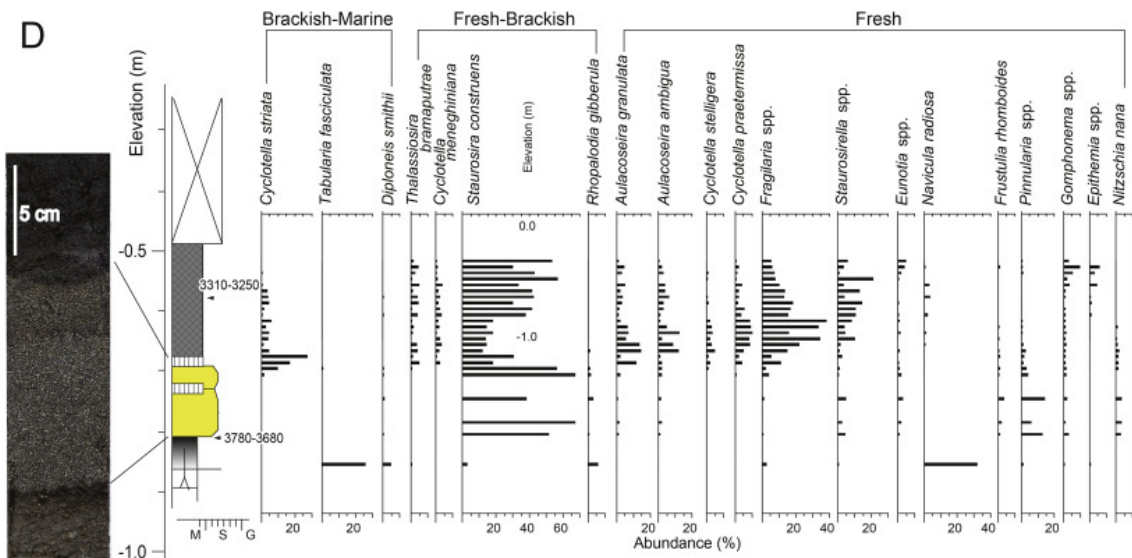34° 40′ N
137° 32′ E
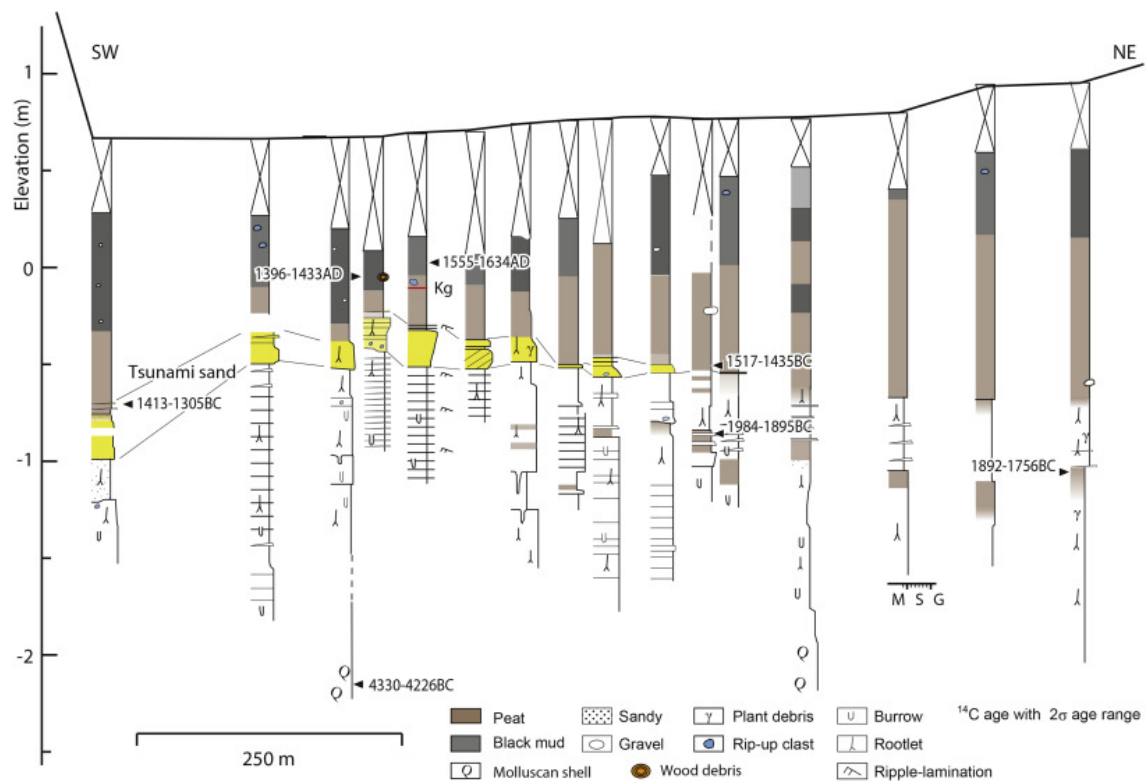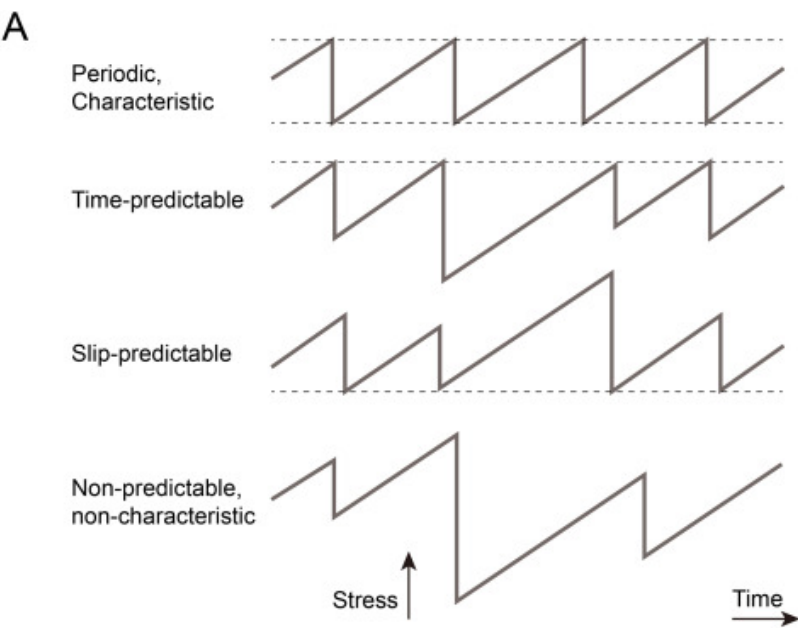137° 42′ E

Enshu-nada

2002

2003

2004

Fig 4. Paleotsunami deposits in the western Hamamatsu Plain

A) Geomorphic classification map of Hamamatsu-Lake Hamana area showing the main coring sites B to E. Map modified from Sato et al. (2011) with additional air photo analyses and land survey data.

B) Geological columnar section and results of diatom analyses from Site B. Occurrence of possible tsunami sand beds is limited beneath the Kg tephra (1210-1187 cal BCE, Tani et al., 2013). Up-core changes in diatom assemblages suggest increasingly freshwater conditions in the back marsh. Increased occurrences of marine-brackish diatoms accompany the deposition of some sand beds. Figure captions are common for Fig. B to Fig. D.

C) Geological columnar section from Site C. The occurrence of possible washover sand beds is limited to the period older than 1175-1055 cal BP (775-895 CE).

D) Geological columnar section and results of diatom analyses from Site D. An increase of brackish-marine diatom species suggests the deposition of a tsunami sand bed ~3700-3300 cal BP following the coastal subsidence.

E) Geological cross-section at the site E. A tsunami sand bed showing a landward fining and thinning trend over 500 m along the incised valley axis. The occurrence of tsunami beds and other sand beds are limited below the Kg tephra.

A

B

	Ansei-type						Hoei-type					
	Z	A	B	C	D	E	Z	A	B	C	D	E
Showa								1946		1944		
Ansei		1854			1854							
Hoei									1707			
Meio		?			1498							
Shohei								1361		1361		
Kowa/Eicho		1099			1096							
Ninna									887			
Hakuho		684			?							

2026

2027 Fig. 5 Schematic illustration of proposed recurrence models for subduction zone earthquakes

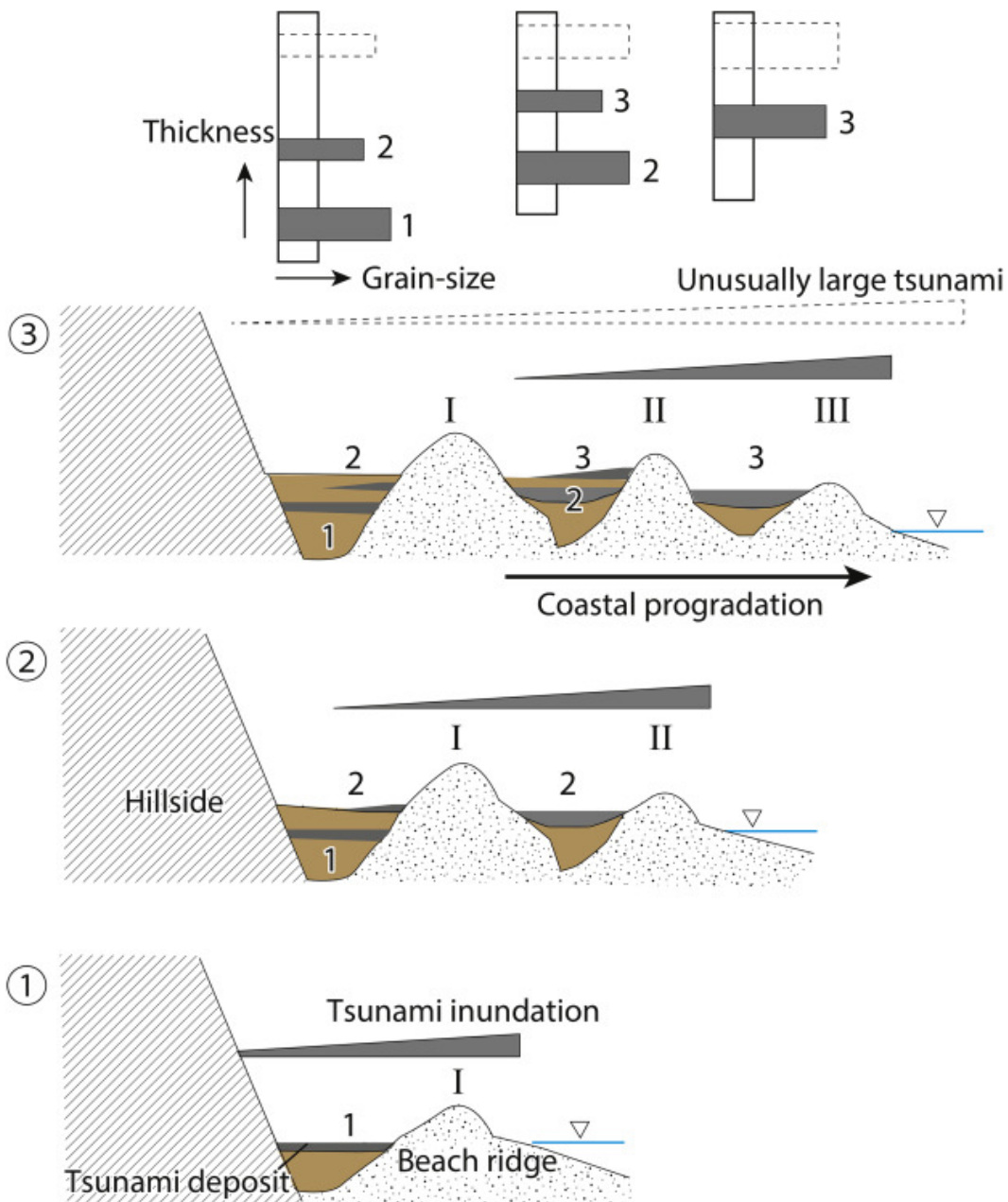2028 A) modified from Satake and Atwater (2007), B) simplified from Seno (2012).

Fig. 6. Conceptual model illustrating the reconstruction of paleotsunami size from the spatial distribution of tsunami deposits considering the geomorphic evolution of a strand plain

This figure was originally shown as Fig. 3 in Fujiwara (2013) and quoted in Fujiwara and Tanigawa (2017).

Table 1. Paleoearthquake and tsunami evidence and related seismotectonic data along the Nankai and Ryukyu subduction zones.

		Rupture zone									
		Sakishima	Okinawa	Amami	Osumi and Tokara	Z	A	B	C	D	E
		Ryukyu Trench				Hyuga-nada	Nankai Trough				
Length of history		~400 years				?	~1300 years				
historical event (Mw>8.0)	1946 Showa						orange	orange			
	1944 Showa								orange	orange	
	1911 Kikai			grey							
	1854 Ansei						orange				
	1854 Ansei								orange	orange	orange
	1791 event		dark red								
	1771 Meiwa	grey									
	1707 Hoei						orange	orange	orange	orange	
	1614 Keicho						hatched	hatched	hatched	hatched	
	1498 Meio						△	△	grey	grey	
	1361 Shohei										
	1361 Shohei										orange
	1099 Kowa						blue	blue			
	1096 Eicho								grey	grey	
	887 Ninna						grey	grey			
	684 Hakuho						orange	orange			
	latest 7th century								△	△	
Seismological evidence	slow slip	not studied	not studied	not studied	not studied	○	○	○	○	○	○
	coupled or decoupled	◎	○	not studied	not studied	not studied	◎	○	○	◎	○
Paleotsunami evidence	late Holocene paleotsunami evidence	Y	N	N	not studied	very rare?	Y	Y	Y	Y	Y
	paleotsunami interval (years)	150–400	–	–	not studied		90–265				
Geomorphological evidence	co-seismic crustal deformation	–	–	Y	–	–	Y	Y	Y	Y	Rare
	...										

Legend:
- orange: Shaking+tsunami+coastal deformation
- grey: Shaking+tsunami
- blue: Shaking+coastal deformation
- △: Liquefaction
- dark red: tsunami