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

Authors:

Osamu Fujiwara[1*], Kazuhisa Goto[2], Ryosuke Ando[2] and Ed Garrett[3,4]

Affiliations:

1. Geological Survey of Japan, National Institute of Advanced Industrial Science and Technology, AIST Central 7, 1-1-1 Higashi, Tsukuba, 305-8567, Japan.
2. Department of Earth and Planetary Science, The University of Tokyo, 7-3-1 Hongo, Bunkyo-ku, Tokyo, 113-0033, Japan.
3. Department of Environment and Geography, University of York, Wentworth Way, Heslington, York, YO10 5NG, United Kingdom
4. Geological Survey of Belgium, Royal Belgian Institute of Natural Sciences, Brussels, Belgium

[*]Corresponding Author: Osamu Fujiwara; Email: o.fujiwara@aist.go.jp.

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

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

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

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

Suruga Trough. The Ryukyu Islands consist of numerous small islands extending over 1000 km along the Ryukyu Trench, which belong to the subtropical climate zone and hence the depositional environment is very different from other parts of Japan. The convergence rate is generally 40-50 mm a^{-1} along the Nankai subduction zone and 50-63 mm a^{-1} along the Ryukyu subduction zone, respectively. The slope of the descending slab, with some local variations, is generally steeper in the Ryukyu subduction zone.

The Amami Plateau, Daito Ridge, Oki-Daito Ridge and basins between them extend in E-W or NW-SE directions on the Philippine Sea Plate. Western extensions of these topographic irregularities subduct beneath the Ryukyu arc (e.g., Kato, 1993). The Ryukyu and Nankai subduction zones are represented by old (40-49 Ma), cold slab that originated from the West Philippine Basin and young (<20 Ma), hot slab that originated from the Shikoku Basin, respectively (e.g., Ishizuka et al., 2011). A thick accretionary prism characterizes the Nankai Trough, in which high-angle reverse faults (splay faults) branching off the plate boundary megathrust may move together with the main thrust during a great earthquake and enhance the tsunami (Baba et al., 2006; Moore et al., 2007).

The Nankai Trough fault region has been empirically divided into five fault planes (rupture zones), which denoted as segments A to E from west to east, respectively (Fig. 1) (Ando, 1975a, b; Ishibashi, 1976, 1981). This classification is mainly based on studies of the source mechanisms of historical earthquakes and the forearc submarine topography, which reflects the seismotectonic activities along this region. Each of the five segments has a length of 100-150 km along the trough axis and almost corresponds to the distribution of forearc basins (e.g., Awata and Sugiyama, 1989; Sugiyama, 1990). In general, plate boundary earthquakes (M8-class) that are considered to have source areas in areas A to B and C to E are called Nankai and Tōkai earthquakes, respectively.

Allocating the source areas of past earthquakes to any of segments A to E is based on the implicit assumption that a large earthquake recurs on its own fault plane. The necessity to criticize and review this assignment has recently been pointed out (e.g., Seno, 2012). The Hyūga-nada area (Figs. 1, 2), where M7-class earthquakes occur separately from the Nankai and Tōkai earthquakes, may be another seismic zone denoted as Z (e.g., Wells et al., 2003). This empirical segmentation of rupture zones has not been applied to the Ryukyu Trench as historical and seismological information is scarce.

Flight of marine terraces, mainly from mid-Pleistocene to Holocene in age, are distributed on the peninsulas, capes and islands of the Nankai and Ryukyu subduction zones (Koike and Machida, 2001), and there have been many discussions about their formation process in relation to sea-level changes and crustal movements (e.g. Yoshikawa et al., 1964; Webster et al., 1998). They show a wide regional variation in average uplift rates. For example, the terraces of marine oxygen isotope stage (MIS) 5e are 40-50 m high in many areas, but reach nearly 190 m at Cape Muroto, the boundary between segments A and B. Coseismic uplift and subsidence during historical earthquakes characterize the tips of the peninsulas and their inland plains, respectively (e.g., Shikoku; Geographical Survey Institute, 1952; Tokai; Ishibashi, 1981). In these area, reverse sense crustal deformation is observed in the interseismic period; gradual subsidence of the peninsulas and uplift of the lowland, respectively (Geographical Survey Institute, 2008).

Kikai Island, located at the northern margin of the Amami Islands on the subducting Amami Plateau, is the closest island to the Ryukyu Trench axis. It shows an extremely high uplift rate, with the height of the MIS 5e marine terrace exceeding 200 m. On this island, Holocene marine terraces consisting of coral reefs record several-meter-scale uplift events that recurred with 1500-year intervals (Webster et al., 1998). The other islands along the Ryukyu Trench do not show remarkable crustal deformation, but the coastal notch in Ishigaki Island records a cumulative uplift of ~2 m in the last 2000 years (Kawana, 1989). Some of these marine terraces distributed along the Nankai and Ryukyu coasts have provided the theoretical basis for time-predictable recurrence models for subduction zone earthquakes (e.g., Shimazaki and Nakata, 1980).

2.2. Seismicity

Understanding the state of inter-plate coupling and seismogenesis in the subduction zones provides basic information for estimating the region's potential of great earthquake and tsunami. Especially, in the Hyūga-nada region, connecting the Nankai and Ryukyu subduction zones, this information is the key to identifying the geological and seismological differences between the two.

Geodetic deformation and inferences of inter-plate coupling along the Nankai Trough and Ryukyu Trench subduction zones

The source areas of large earthquakes are usually expected to present strong inter-plate coupling during inter-seismic periods (e.g., Lay and Kanamori, 1981). The degree of the coupling can vary over fault areas, reflecting differences in the fault's frictional properties. If a fault creeps steadily, it has a low degree of coupling, while high coupling results from a seismogenic fault that is stuck during inter-seismic periods. The degree of the coupling is quantified in geodesy (Savage, 1983) by the slip-deficit rate (SDR) (or back-slip rate), which is the rate of upper plate displacement along with the plate interface driven by the subduction motion of the lower plate. The coupling coefficient is also determined as the SDR divided by the long-term average rate of the relative motion between the two plates. When the fault is completely coupled, the SDR is equal to the long-term average relative plate motion rate and the coupling coefficient is unity. SDRs are estimated from geodetic data by applying slip inversion analysis (Savage, 1983).

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

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

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

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

Seismological inferences of inter-plate coupling using repeating earthquakes

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

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

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

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

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

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

The family of slow earthquakes has been discovered over the past twenty years along the Nankai Trough by seismological (e.g., Obara, 2002) and geodetic (e.g., Hirose and Obara, 2005) observations, and their correlation with inferred inter-plate coupling states are confirmed in great detail. Along the plate interface at depths from ~30 to ~40 km, tremor and short-term SSEs (S-SSEs) are generally observed with recurrence periods of ~6 months (Obara, 2002; Hirose and Obara, 2005), while long-term SSEs with M 7 have recurred at ~10-year intervals particularly beneath Lake Hamana in Tōkai region (Suito and Ozawa, 2009 and references therein). For the Ryukyu Trench, seismological observations suggest that slow earthquakes, called very low frequency earthquakes, probably occur on the plate interface over the entire Ryukyu Trench (Ando et al., 2012a), and geodetic observations identify seemingly corresponding short-term SSEs (S-SSEs) and long-term SSEs (L-SSEs) (Heki and Kataoka, 2008; Nishimura, 2014). Nishimura (2014) suggested that the occurrence of S-SSEs at shallow depths may be a signature of incomplete inter-plate coupling. Tadokoro et al. (2018) suggested the inferred strong coupling area does not overlap with the SSE areas, and heterogeneity in the areas of SSE occurrence should exist along the Ryukyu Trench.

The Hyūga-nada region appears to present a transitional character in terms of the spatial distributions of slow earthquakes. In the other areas of the Nankai Trough, shallow slow earthquakes and deep S-SSEs and tremors are both observed, while deep S-SSE/tremor is not observed in the Hyūga-nada region (Obara and Kato, 2016). The occurrence of shallow slow earthquakes has been characterized in great detail owing to the recent deployment of ocean bottom seismometers (Yamashita et al., 2015). Tremor was observed to migrate in the shallow depths of the plate interface, and the lateral extent of the migration was delimited by the possible area of the subducting Kyushu-Palau ridge (Yamamoto et al., 2013). The down-dip limit of the migration was not reached until ~ 20 km, corresponding to the hypocentral depths of M 7-class earthquakes in this region, suggested that it stopped at the shallower end of the seismogenic zone (Yamashita et al., 2015). This implies that the intermediate depths of the plate interface are still coupled and have the potential to nucleate a major earthquake. The geophysical observations clearly show that the inter-plate coupling of this area is relatively low overall, but these observations still leave the possibility that the rupture of this coupled zone could be triggered by neighboring large earthquakes, or the rupture of this area could facilitate interactions between neighboring large earthquakes, if their timings are synchronized.

While the above-mentioned geophysical observations have clarified the spatial characteristics of the inter-plate properties, we should still remember that our experience is limited by the short history of geophysical observations. The states of inter-plate coupling can slowly fluctuate even over the course of several years (Uchida et al., 2016), and limited snapshots can fail to reveal critical properties. Integration of paleoseismological evidence is still critically important to capture phenomena spanning long periods of time.

3. Record of tsunamis and tsunami deposits

3.1. Nankai Trough region

3.1.1. Historical Tōkai and Nankai earthquakes and tsunamis

The Nankai Trough region, which is close to the political and cultural center of Japan, the Kyoto-Nara-Osaka area, has the longest and most complete documented record for the recurrence of subduction zone earthquakes and tsunamis in the world. According to this 1300-year long record, earthquakes have occurred in a variety of modes. Some earthquakes have ruptured almost the entire subduction zone at the same time, while at other times, two earthquakes, referred to as Tōkai and Nankai earthquakes, have occurred consecutively, separated by short intervals (Fig. 2). Nankai earthquakes that are historically confirmed include the 684 CE Hakuho (M8.0-8.5), 887 CE Nin-na (M8.0-8.5), 1361 CE Kōan (M8.2-8.5), 1707

CE Hōei, 1854 CE Ansei-Nankai (M8.4; only 32 hours after the 1854 Ansei-Tōkai earthquake), and 1946 CE Showa-Nankai (M8.0) earthquakes, with further unconfirmed events in 1099 and 1614 CE. Tōkai earthquakes confirmed by documented records include the 1096 CE Eichō (M8.0-8.5), 1498 CE Meiō (M8.2-8.4), 1707 CE Hōei (M8.6), 1854 Ansei-Tōkai (M8.4), and 1944 CE Tōnankai (M7.9) earthquakes, and a further rupture may have occurred in 1614 CE. The 1707 earthquake was a simultaneous rupture of the Tōkai and Nankai fault segments (segments A to D, and perhaps also including the southern part of E) and was the largest earthquake in Japanese history until the occurrence of the 2011 Tōhoku earthquake.

There are two possibilities for a great earthquake that occurred along the Nankai Trough in the early 17th century. The first is the 1605 CE Keichō earthquake, which generated a large tsunami from Kanto to Kyushu. This event has been considered as a "tsunami earthquake", because only weak shaking accompanied the large tsunami (e.g., Ishibashi and Satake, 1998). Ando and Nakamura (2013) suggested that the rupture occurred along a shallow portion of the plate interface. An alternative hypothesis is that the 1605 earthquake occurred along the Izu-Ogasawara Trench, to the east of Izu-Bonin Arc (Harada et al., 2013). Instead of 1605 CE event, Ishibashi (2014) proposed the 1614 CE earthquake as a candidate for a 17th century Nankai Trough earthquake, based on records of strong ground shaking in Kyoto and tsunamis at the Kii Peninsula. A Nankai earthquake paired with the 1498 CE Meiō Tōkai earthquake has not been confirmed; nevertheless, liquefaction features in western Japan suggest the occurrence of a Nankai earthquake around 1498 CE (Sangawa, 2001, 2007). Ishibashi (2014) proposed four earthquakes that occurred between 1498 and 1512 CE as candidates, but there is no solid record of tsunami inundation during this period.

Documentary records confirming the occurrence of a Tōkai earthquake paired with the 1361 CE Nankai earthquake have not yet been discovered. As a candidate for this earthquake, Ishibashi (1998, 2014) proposed an earthquake that featured strong shaking in Kyoto, Nara, and Kumano (north of Shiono-misaki; boundary between the segments B and C) on July 24th 1361 (2 days before the Nankai earthquake). Late 14th century coseismic uplift around Cape Omaezaki (Fig. 2, site 27) shown by emerged boring bivalves (Kitamura et al., 2018c) and marine terrace (Fujiwara et al., 2010) support Ishibashi's hypothesis.

The occurrence of the 1099 CE earthquake was originally inferred from documents suggesting strong ground shaking in the Kyoto, Nara, and Osaka region (but less than 1096 CE earthquake) and coastal subsidence in Kochi (around site 10 in Fig. 2); which matches general characteristics of other Nankai earthquakes (e.g., Ishibashi, 1999). However, the occurrence of this earthquake is uncertain due to the unreliability of underlying evidence, especially the lack of reports of strong ground shaking and tsunami inundation in the area facing segments A and B. Ishibashi (2016) suggested that the 1099 CE earthquake did not actually exist and that the 1096

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

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

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

3.1.2. Tsunami deposits

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

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

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

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

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

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

For Shikoku and northern Kyushu, in addition to the summary provided by Okamura and Matsuoka (2012) from 26 ponds, Baranes et al. (2016) conducted paleotsunami research in three ponds. Among these 29 sites, historical tsunami deposits have been found in four locations. At Ryujin-ike in northern Kyushu (Fig.2, site 7), separated by a 17 m high and 100 m wide sand dune, a narrow channel is the only access to the Pacific. According to historical documents, the 1707 CE Hōei tsunami, over 10 m-high, was much higher than that of 1854 and 1946 CE tsunamis around Ryujin-ike (e.g., Chida and Nakaue, 2007). Among 40 sand beds intercalated within the 5.4 m long cored humic mud sequence, eight prominent beds, each 5 - 20 cm-thick, are interpreted as washover deposits due to their thinning landward trend and the inclusion of beach sand and marine bioclasts (Okamura and Matsuoka, 2012). Based on radiocarbon dating, Okamura and Matsuoka (2012) linked the youngest three beds to the 1707, 1361 and 684 CE tsunamis.

Baranes et al. (2016) reported a total of nine washover sand beds from the back-barrier lakes facing the Bungo Channel in western Shikoku (Fig. 2, site 8). A 9.3 m-deep sediment core from Lake Ryuuō, consisting of mainly clay beds and covering ~2500-year-long record of deposition, shows an environmental change from a brackish lake to freshwater conditions that occurred around 1000 CE (about 6 m below the lake bottom). Nine washover sand beds occur in the freshwater clay between the depths of 595 and 271 cm, accumulated from 1000 to 1800 CE. Although with substantial chronological uncertainties, their depth-to-age model suggests the possibility that four washover sand beds represent the 1099, 1361, 1605, and 1707 CE tsunamis. There are two sand layers deposited around 1500 CE, one of which may be due to a Nankai earthquake that is unknown from historical records and paired with the 1498 Meiō Tōkai earthquake. Baranes et al. (2016) concluded that the 1707 CE tsunami, which left the thickest (25 cm) coarse-grained sand bed, was one of the most significant floods over the last millennium in the region.

Fourteen sandy tsunami deposits ranging from 4800 to 1300 cal BP in age are reported from Tadasu-ike (Fig.2, site 9), located 800 m inland from the modern coast and behind a 5 m high beach ridge (Okamura and Matsuoka, 2012). The sand beds range from 5 to 27 cm in thickness. The pond was inundated by the 1707 and 1946 CE tsunamis, however the record of the last 1300 years is missing in this pond due to artificial disturbance; the youngest bed was linked to the 684 CE tsunami (Okamura and Matsuoka, 2012). Kaniga-ike in Shikoku (Fig.2, site 10), located about 400 m inland from the modern coast and fringed by 6 m-high coastal dunes, records six sandy tsunami deposits in the last 2000 years (Okamura and Matsuoka, 2012). The tsunami beds range from 5 to 75 cm in thickness. Based on historical documents and limited radiocarbon ages (Cabinet Office, 2011a), the youngest two can be correlated with the 1854 CE Ansei-Nankai and 1707 CE Hōei earthquakes, while the third and fourth ones from the top may be attributed to the 1361 and 684 CE Nankai earthquakes, respectively.

Prehistoric tsunami deposits

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

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

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

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

(1210-1187 cal BCE, Tani et al., 2013). In a more seaward swale, 2 km from the present coast (Fig. 4C), the uppermost deposit dates to 775-895 CE.

In Shikoku and Kyushu, five tsunami deposits have been recognized from Ryujin-ike (Fig. 2, site 7) between 3300 and 1600 years ago, thirteen from Tadasu-ike (Fig. 1, site 9) between 3300 and 1600 years ago, and two from Kaniga-ike (Fig. 2, site 10) between 1000 BCE and 600 CE (Okamura and Matsuoka, 2012). Tanigawa et al. (2018) reported four sandy washover deposits tapering landward from sediment cores drilled in a small coastal lowland in Nankoku, Kochi Prefecture (Fig. 2, site 11). The lowland is separated from the Pacific by dunes 450 m wide and 13 m high. The sand beds are laterally extensive through marine-brackish clay and overlying freshwater clay and peat. Tanigawa et al. (2018) interpreted the sand beds as having been deposited by tsunamis or unusually large storm surges. The possible tsunami sand beds were found in sediments with ages between 5970 and 2440 cal BP. They attributed the lack of marine incursions after 2440 cal BP to the development of beach ridges that protected the site from tsunamis and storm surges.

Shimada et al. (2019) reported five tsunami inundations between 5581 and 3640 cal BP from a small drowned valley in Mugi (Fig. 2, site 12), eastern Shikoku. Array coring along the valley axis and diatom analyses of the cored sediments revealed the deposition of at least nine exotic sand and gravel beds in the muddy brackish-to-fresh wetland sequence. Environmental changes inferred from diatom assemblages suggest the occurrence of coastal subsidence associated with the deposition of three of the sand beds. Subsidence of 0.3 m or more was reported around the study site during the 1946 CE Showa-Nankai earthquake (Japan Coast Guard, 1948). Shimada et al. (2019) tentatively attributed two other sand beds to tsunami events, while acknowledging the possibility of deposition from a storm surge or far-field tsunami. Reported tsunami deposits can be traced up to 240 m or more inland from the current shoreline, and they range in thickness from 5 to 20 cm with some exceptions up to 75 cm.

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

3.2.1. Historical earthquakes and tsunamis

The Hyūga-nada region covers the western end of Nankai Trough and the northern end of Ryukyu Trench. Seismic activity in this region is relatively high and small tsunamis were detected following the 1941, 1961, 1968, and 1984 CE earthquakes (Hatori, 1985). Historically, large tsunamis occurred in 1662 and 1769. The 1662 CE earthquake was the largest historical event, with a magnitude of 7.6 (HERP, 2004) and a maximum tsunami height of <5 m at Miyazaki Plain (Hatori, 1985). Hatori (1985) estimated that the 1662 CE tsunami could have inundated the lowland of Miyazaki Plain via the large rivers, while tsunamis might not have been able to overtop the coastal sand dunes because of their height (~10 to 15 m, Nagaoka et al., 1991).

3.2.2. Paleotsunami deposits

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

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

3.3. Paleotsunamis on the Ryukyu Islands

3.3.1. Historical earthquakes and tsunamis

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

In contrast to Nankai Trough region, historical records from the Ryukyu Islands are only available for the last ~400 years. Historical and instrumental data show that earthquake and tsunami occurrence in this region is remarkably low in frequency. Indeed, the only earthquakes with magnitudes >8 during the past ~400 years were in 1911 CE in the northern Ryukyu Trench (Mw=8.0, e.g., Tsuji, 1997), in 1791 CE in the central Ryukyu Trench (Mw=8.0 - 8.3, Nakamura and Kinjou, 2013; Tadokoro et al., 2018), and in 1771 CE in the southern Ryukyu Trench (Mw>8.0, e.g., Nakamura, 2009). However, because there is no record of ground shaking accompanying the 1791 CE tsunami, it is uncertain whether this resulted from a near-field tsunamigenic earthquake (Nakamura and Kinjou, 2013; Tadokoro et al. 2018) or was potentially a far-field tsunami generated somewhere in the Pacific Ocean (Matu'ura, in press). The 1771 CE tsunami was exceedingly large with maximum run-up heights of approx. 30 m (e.g., Goto et al., 2010a). However, based on detailed historical records, it is also well known that the tsunami-affected area was spatially limited to a narrow range of the Sakishima Islands (Goto et al., 2010a).

Lowland plains suitable for paleotsunami research using sandy tsunami deposits are extremely scarce because of the fringing reef environment. On the other hand, numerous coral boulders are scattered on the reefs surrounding many islands and they are considered as useful geological evidence of past extreme wave events such as tsunamis and storm waves (Fig. 3B, e.g., Kawana and Nakata, 1994; Goto et al., 2010a).

In this section, we summarize paleotsunami research on the Ryukyu Islands based on the comprehensive review by Goto (2017) with additional recent works. As a brief summary, the paleotsunami history of the Amami and Okinawa Islands can be characterized by scarce (or even no) evidence of large tsunamis over the last few thousand years, while the Sakishima Islands can be characterized by frequent large tsunamis with intervals of a few hundred years.

3.3.2. Paleotsunami deposits on the Amami and Okinawa Islands

On the Amami Islands, Kawana and Nakata (2003) estimated the possible occurrence of tsunamis in 360-510 cal BP, 1390-1770 cal BP, and 2100-2110 cal BP, based on dating coral boulders on the reef. Similarly, Kawana (2006) and Iwai and Kawana (2008) suggested possible large tsunamis around 3400 years ago on the Okinawa Islands, based on the dating of beach rocks and coral boulders. However, following studies by Goto et al. (2009, 2013), field surveys and remote sensing analyses revealed that all of the coral boulders deposited on the reefs fringing the Amami and Okinawa Islands can be explained by transport during storms despite their heavy weight (~160 tons). The boulders are deposited close to the source area (i.e. around the reef edge) and have not been transported a long distance. Since the reef was formed and the boulders were emplaced on the reef at least 2300 years ago (Kawana and Nakata, 2003), the present clast size and spatial distribution of boulders reflects the cumulative impacts of many storm waves since at least 2300 years ago. If large tsunamis had occurred during this period, boulders originally deposited on the reef by storm waves should be transported further inland, as is observed in the Sakishima Islands (discussed below). Therefore, Goto et al. (2013) concluded that no tsunamis large enough to significantly modify the characteristic distribution of storm wave boulders have affected these islands in the last 2300 years. Indeed, using numerical modeling, Minamidate et al. (2020) confirmed that boulders in the Okinawa Islands can be explained by the realistic size of storm waves generated by past typhoons.

Another important paleoseismological aspect, especially in the Amami Islands, is a rapid uplift of Kikai Island (e.g., Webster et al., 1998). At Kikai Island, the terrace of MIS 5e is at an elevation of 224 m (e.g., Ota and Omura, 2000). At least 4 emerged Holocene coral terraces are observed and each terrace is associated with emergence of 1 to 4 m (Webster et al., 1998; Sugihara et al., 2003; Hongo, 2010). Because of these geomorphological features, it is thought that Kikai Island was intermittently uplifted by large earthquakes (Nakata et al., 1978). On the other hand, considering the absence of tsunami boulders but presence of storm boulders both on Kikai and Amami-Oshima Islands (Goto et al., 2013), earthquakes that might have caused the uplift of the island may not have generated large tsunamis. The uplift might be explained by slip on an intra-plate fault, such as a splay fault, rather than slip on the subduction interface (Goto, 2017). This idea may better fit with the fact that nearby Amami-Oshima Island (~40 km to the west) has not been uplifted significantly compared with Kikai Island (Ikeda, 1977) and hence these islands are tilted within a narrow area. Alternatively, Shikakura (2014) numerically modeled crustal movements and the formation of the coral terraces at Kikai Island and suggested that large earthquakes are not mandatory requirements, but steady uplift with some intermittent earthquakes may alternatively be enough to explain the uplift as well as the formation of multiple coral terraces.

3.3.3. Paleotsunamis on the Sakishima Islands

Because of historical records that describe the movement of huge boulders during the 1771 CE tsunami and the presence of numerous boulders deposited on the reef and on land, identification of historically described 1771 CE tsunami boulders has proved an attractive topic. With this objective, geological and historical studies of the 1771 CE tsunami and boulders began earlier than tsunami studies in other regions, with the earliest work by Makino (1968). Makino (1968, 1981) interpreted that all boulders deposited on land were transported by the 1771 CE tsunami. However, following studies cautioned the need for careful identification because many of the boulders are composed of the Pleistocene Ryukyu Limestone that forms the island itself and may have been sourced from higher elevations and moved by gravity (Kato and Kimura, 1983; Kawana and Nakata, 1994). Nevertheless, there are numerous coral boulders with fresh Holocene coral skeletons that must have been transported from the sea possibly by large tsunamis during Holocene (Kato and Kimura, 1983; Kawana and Nakata, 1994).

Radiocarbon dating of coral boulders has been performed since early 1980's and boulders with ages around 1771 CE have been discovered, suggesting their origin from this tsunami (e.g., Kato and Kimura, 1983; Kawana and Nakata, 1994). However, many studies have revealed that there are boulders with much older ages than 1771 CE, suggesting that these boulders are likely to have been deposited by repeated tsunamis during the few thousand years before 1771 CE (Kawana and Nakata, 1994; Omoto, 2012; Araoka et al., 2013).

Problems associated with pre-2010 CE studies on these boulders include 1) discrimination of tsunami boulders from storm boulders, and 2) dating accuracy and reliability of tsunami age estimation. Regarding the first issue, Goto et al. (2010b, 2013) investigated clast size and the spatial distribution of boulders on the Sakishima Islands and suggested that coral and reef boulders can be classified into two groups. One group has a characteristic distribution of storm wave boulders such as a narrow range in spatial distribution from the reef edge and an exponentially fining landward trend in clast-size distribution. The boulders of this group can be considered as of storm wave origin (Goto et al., 2010b, 2013). On the other hand, the other group can be characterized by boulders that are deposited along the coast and on land. They are deposited far beyond the possible landward limit of storm wave boulders (i.e. ~300 m from the reef edge, Goto et al., 2009). Also, there is no clear clast size distribution indicative of storm deposition. Based on these features, Goto et al. (2010b, 2013) identified the boulders in this group as of tsunami origin. Later, Watanabe et al. (2016) numerically confirmed the tsunami origin of the boulders in the latter group, identifying that extremely long wavelengths are required to explain the deposition of boulders along the coast.

Regarding the second issue, Omoto (2012, 2019) (Fig. 2, site 3) and Araoka et al. (2013) (Fig. 2, site 2) carefully selected samples for dating. They limited measurements to the boulders that meet tsunami boulder identification criteria proposed by Goto et al. (2010b, 2013) and that were composed of single massive corals (i.e. *Porites* sp., Fig. 3B). Based on this approach, Araoka et al. (2013) estimated a tsunami recurrence interval of about 150-400 years on the Sakishima Islands. It should be noted, however, that they measured ages of various sizes of boulders so it is uncertain whether the paleotsunamis that cast the boulders ashore were large or small.

There are some studies that focus on sandy tsunami deposits from the Sakishima Islands. Possible sandy tsunami deposits, especially relating to 1771 CE, were previously reported by archaeologists (e.g., Yamamoto, 2008; Kugai, 2011; Nakaza, 2017) and coastal engineers (Nakaza et al., 2013) following the work done by Kawana and Nakata (1994) but their detail history had remained uncertain. Recently, Ando et al. (2018) performed a 120 m-long trench survey and found that paleotsunamis may have occurred at approximately 600-year intervals over the last 2000 years (Fig. 2, site 1). This observation is consistent with numerical modeling of the movement of a very large boulder at southeastern Ishigaki Island by Hisamatsu et al. (2014) who suggested at least 2 large tsunamis equivalent to or even larger than the 1771 tsunami occurred after 2000 years ago but before 1771 CE. Interestingly, the recurrence interval of paleotsunamis estimated from sandy tsunami deposits (Ando et al., 2018; Kitamura et al., 2018a, b) is longer than that estimated from boulders (Araoka et al., 2013). This discrepancy may be explained by coastal boulders being more sensitive in recording tsunamis because even small tsunamis can transport small boulders to the coast. Therefore, evidence of small to medium tsunamis are included in the estimation of tsunami recurrence intervals from boulders. Meanwhile, sandy tsunami deposits that extended far inland may only be suitable to recognize large events. Alternatively, this discrepancy might also reflect the small number of studies carried out on sandy deposits in this region so further research is required. Although some researchers have suggested the absence of a predecessor to the 1771 CE tsunami (Nakaza et al., 2013, 2015), geological evidence of both boulders and sandy deposits provides crucial evidence that either small or large tsunamis have repeatedly occurred over the last few thousand years around the Sakishima Islands.

4. When and where: the history of great earthquakes

This section summarizes the spatial and temporal distribution of paleotsunami evidence along the Nankai and Ryukyu subduction zones. Table 1 summarizes when the historical earthquakes occurred in these regions.

4.1. Nankai subduction zone

Great earthquakes occurred in the Tōkai region nine times in total, including the controversial 1614 CE event, from the end of the 7th century until 1944 CE. On the other hand, great earthquakes occurred in the Nankai region eight times from 684 to 1946 CE. The 1099 CE earthquake was recorded with an incorrect date, and may have actually occurred in 1096 CE (Ishibashi, 2016). The Tōkai and Nankai earthquakes always occur in pairs within a few days to several years, except for the 1498 CE Meiō earthquake. A Nankai earthquake paired with the 1498 CE Meiō Tōkai earthquake is likely to have occurred, but we have not been able to definitively ascribe any of the multiple candidates from the historical record. The 1707 CE Hōei and 887 CE Nin-na earthquakes were full-length ruptures of both the Tōkai and Nankai fault segments.

Some questions remain about the recurrence pattern of Nankai Trough earthquakes. First, the recurrence intervals of earthquakes in the Tōkai and Nankai regions appears to be different before and after the 1361 CE earthquake (Table 1). Even if the 1614 CE earthquake did exist, it seems unsuitable to include it within a series of great Nankai Trough earthquakes because the seismic intensity and tsunami during the 1614 CE event were much smaller than other earthquakes. If this earthquake is included, the recurrence interval from the 1361 CE earthquake to the 1854 CE earthquake were 137 years (1361 - 1498 CE), 116 years (1498 - 1614 CE), 93 years (1614 - 1707 CE), and 147 years (1707 - 1854 CE). On the other hand, before 1361CE, recurrence intervals were much longer; 203 years (684 - 887CE), 209 years (887 - 1096 CE), and 265 years (1361 - 1096 CE). It is unlikely that the process of plate boundary earthquakes suddenly changed at some point and, while it is possible that the apparent change is an artifact of the small sample number, other explanations may also be considered.

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

4. 2. Ryukyu subduction zone

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

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

5. Variation in size and recurrence time

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

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

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

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

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

Progress in methods for quantifying the tsunami size from deposits are needed to elucidate the variations in earthquake size. Most lowlands along the Nankai Trough coast are small, with mountains and hills close to the coast. In such a situation, tsunamis reach the landward margin of the lowland irrespective of their size, and it is not possible to compare the differences in inundation distances. The tsunami height could, however, be estimated using sediment transport modeling, using parameters such as grain size, the thickness of tsunami deposits, and the height of coastal barrier (cf. Baranes et al., 2016). Tsunami boulders may be a useful tool for estimating local tsunami size especially in the Ryukyu Islands (e.g., Watanabe et al., 2016). However, the future improvement of models and development of methodologies to collect field data about the boulder source and shape are required for further high accuracy modeling.

6. How big?

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

6.1. Nankai subduction zone

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

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

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

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

6.2. Hyūga-nada and the Ryukyu subduction zone

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

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

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

Arai et al. (2016) and Nakamura (2017) suggested there is a strongly coupled zone at shallower plate boundary depths in the southern Ryukyu Trench. Therefore, localized tsunamigenic earthquakes at the southern Ryukyu Trench inferred from tsunami boulders (Goto et al., 2013) may be explained by the heterogeneity of the coupling along the Ryukyu Trench. Nevertheless, the source model of the 1771 CE tsunami is still controversial with proposed models including an intra-plate earthquake plus a submarine landslide (Imamura et al., 2008; Miyazawa et al., 2012), a tsunami earthquake (Nakamura, 2009), splay faulting (Hsu et al., 2013), and a large submarine landslide near the trench axis (Okamura et al., 2018). By assuming a large subduction zone earthquake along the Ryukyu Trench, Ando et al. (2018) estimated a seismic coupling coefficient of 20%. Alternatively, if a submarine landslide played an important role in the generation of the 1771 CE tsunami and predecessors, then a special tectonic state capable of generating large earthquakes only at the southern Ryukyu Trench may not need to be considered. Relationships among VLFEs, short-term SSEs, and large earthquakes (Mw ~ 8) are still uncertain, so further seismological research with reference to paleotsunami results is required along the Ryukyu Trench.

7. Knowledge gaps and future work

7.1. Earthquake recurrence model

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

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

Seno (2012) proposed a new earthquake recurrence model for Nankai Trough earthquakes that challenged the traditional idea of the characteristic earthquake model, which partitions the Nankai Trough seismogenic zone into ruptures A-E and assign the epicenters of historical earthquakes to them. He assumed two complementary earthquakes that have different non-overlapping asperities mainly based on the spatial distribution of seismic intensities, while also considering tsunami heights and coastal deformation estimated from documented records and liquefaction features in archaeological sites (Fig. 5B). According to his idea, historical great Nankai Trough earthquakes are classified into two types; “Hōei-type” earthquakes (887, 1361, 1707 and 1944+1946 CE) and the “Ansei-type” earthquakes (684, 1096+1099, 1489, 1854+1854 CE), which alternately occur with recurrence intervals of around 350 and 400 years, respectively. Seno’s (2012) model seems to explain the variations in the recurrence intervals and extent of source area of each earthquake in Nankai Trough; however, the reliability of model is reduced by the decrease in the number of reliable documentary records with increasing age, especially before the 1096 CE earthquake.

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

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

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

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

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

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

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

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

Crustal movements suggesting outsized earthquakes along the Nankai Trough

In discussing the possibility of a giant earthquake, it is necessary to assess a wide range of crustal deformation patterns occurring at different temporal and spatial scales. In the case of eastern Hokkaido, where the recurrence of giant tsunami along the Kuril Trench is shown by tsunami deposits, there is a contradiction between the coastal deformation pattern shown by geological and geodetic data. In this area, 60 to 80 km above the top of the subducting plate, marine terraces showing 20 to 50 m of net uplift in the past 125 ka (Okumura, 1996) coexist with tide gauge and bench mark-inferred subsidence of up to 1 m in the past 100 years (Shimazaki, 1974a; Ozawa et al., 1997). Little, if any, uplift is known from the M8-class earthquakes in this region's written history during the past 200 years (Shimazaki, 1974b; Kasahara, 1975; Kasahara and Kato, 1980; Yamanaka and Kikuchi, 2003).

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

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

Major differences between this area and eastern Hokkaido include the inter-seismic uplift after the 1854 CE Ansei-Tōkai earthquake and the short-term uplift associated with SSEs. Inter-seismic uplift, averaging ~6 mm/y between 1901 and 2008 CE (Geographical Survey Institute, 2008), may compensate for the coseismic subsidence during "ordinal" Tōkai earthquakes; however, other factors are needed to explain the formation of high and tilted terraces. This average uplift rate includes the effect of SSEs (see chapter 2.2). SSEs seem to occur with decadal recurrence intervals and are high frequency uplift events on a geological time scale, but associated uplift seems too small to form the terraces. For example, a Tōkai slow slip event (Ozawa et al., 2002; Miyazaki et al., 2006) occurred under Lake Hamana from January 2000 to July 2005, with a cumulative moment magnitude of ~7.1, and generated ~5 cm uplift in this area (Suito and Ozawa, 2009). At present the importance of SSEs for terrace formation remains unclear and their contribution may need to considered further.

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

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

Possibility of full-length rupturing event of the Nankai and Ryukyu subduction zone

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

8. Conclusions

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

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

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

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*English translation from the original written in Japanese.

Figure captions

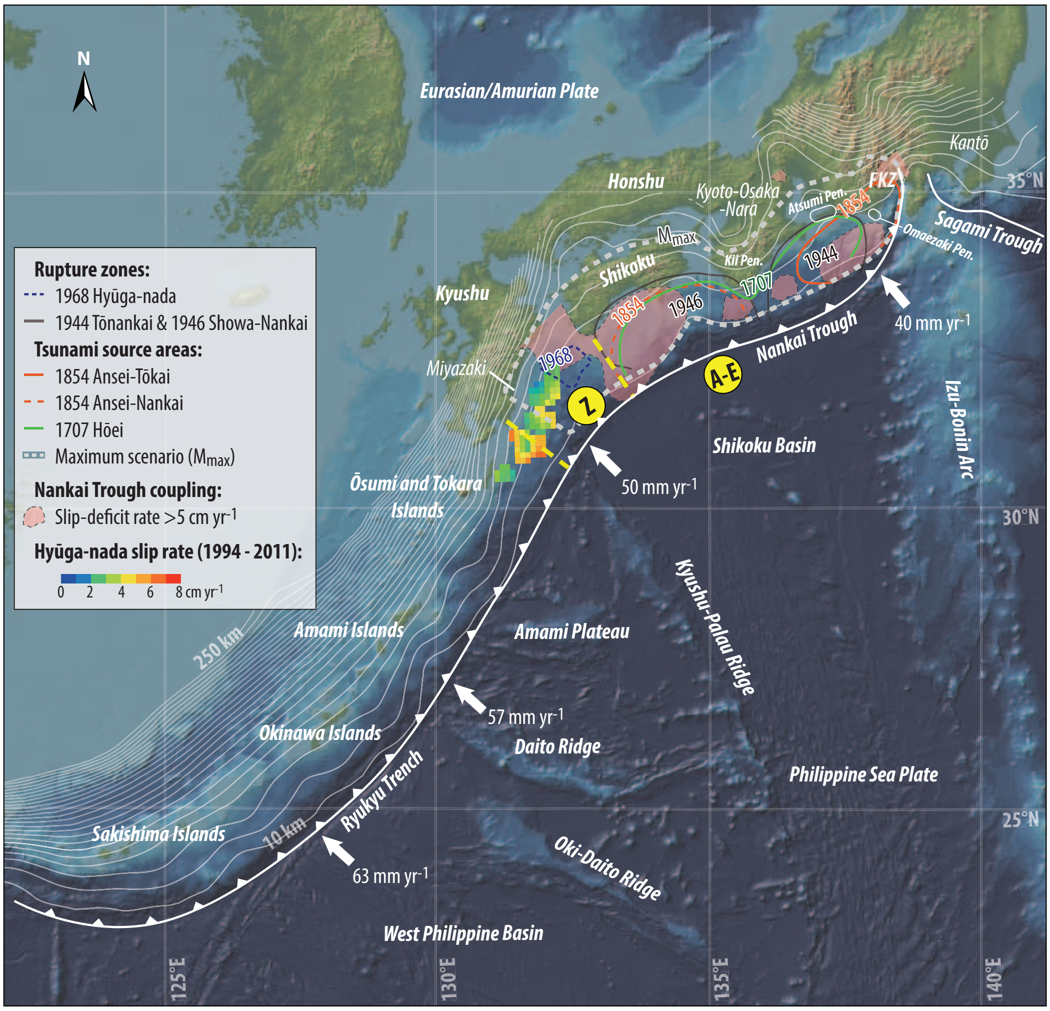


Fig. 1. Map showing the outline of the seismotectonic setting around the Nankai and Ryukyu subduction zones.

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

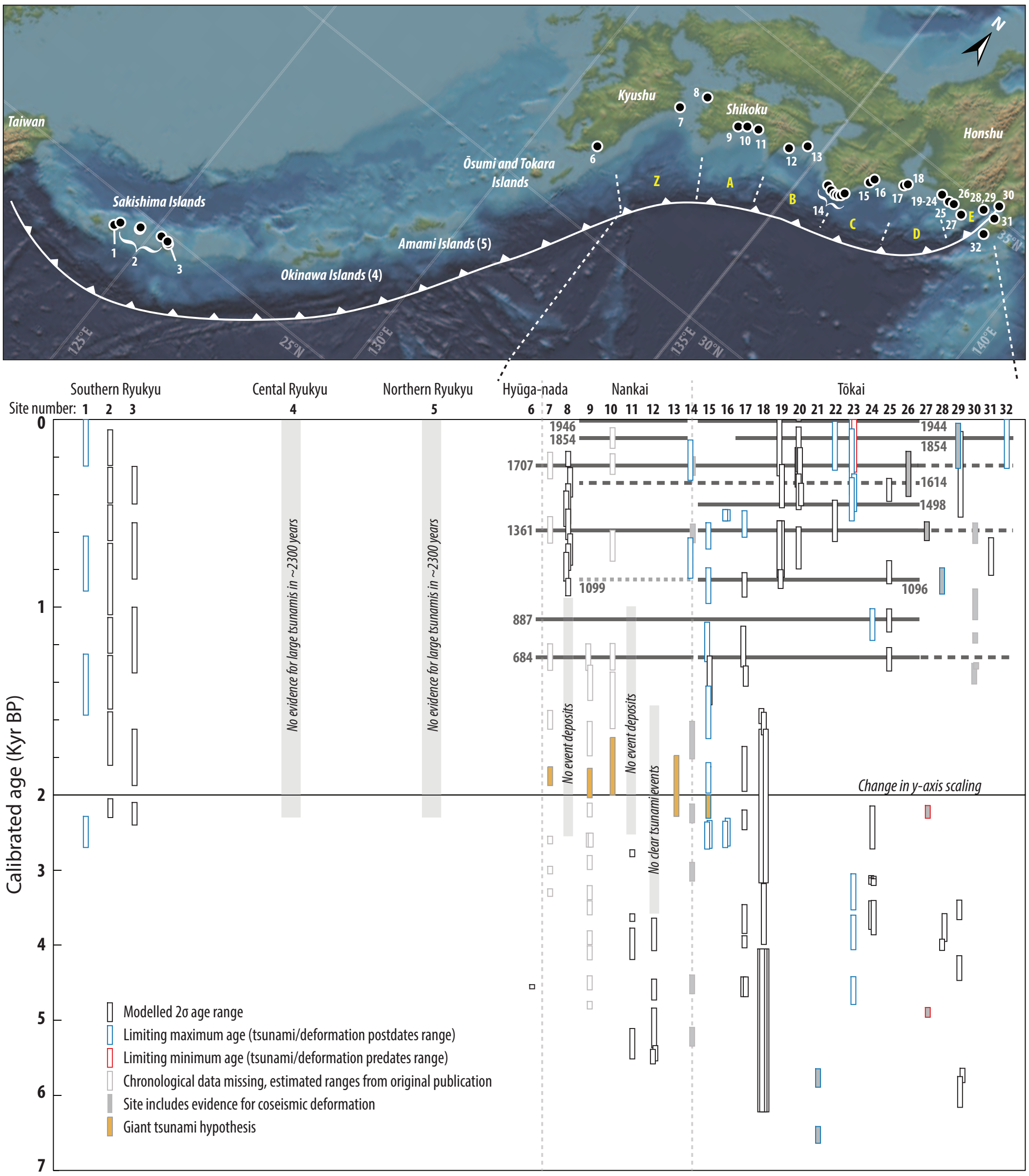


Fig. 2. Spatiotemporal compilation of representative paleotsunami evidence along the Ryukyu and Nankai subduction zones

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.

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

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

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

Main references cited in Garrett et al. (2016) but not discussed further in this paper are as follows.

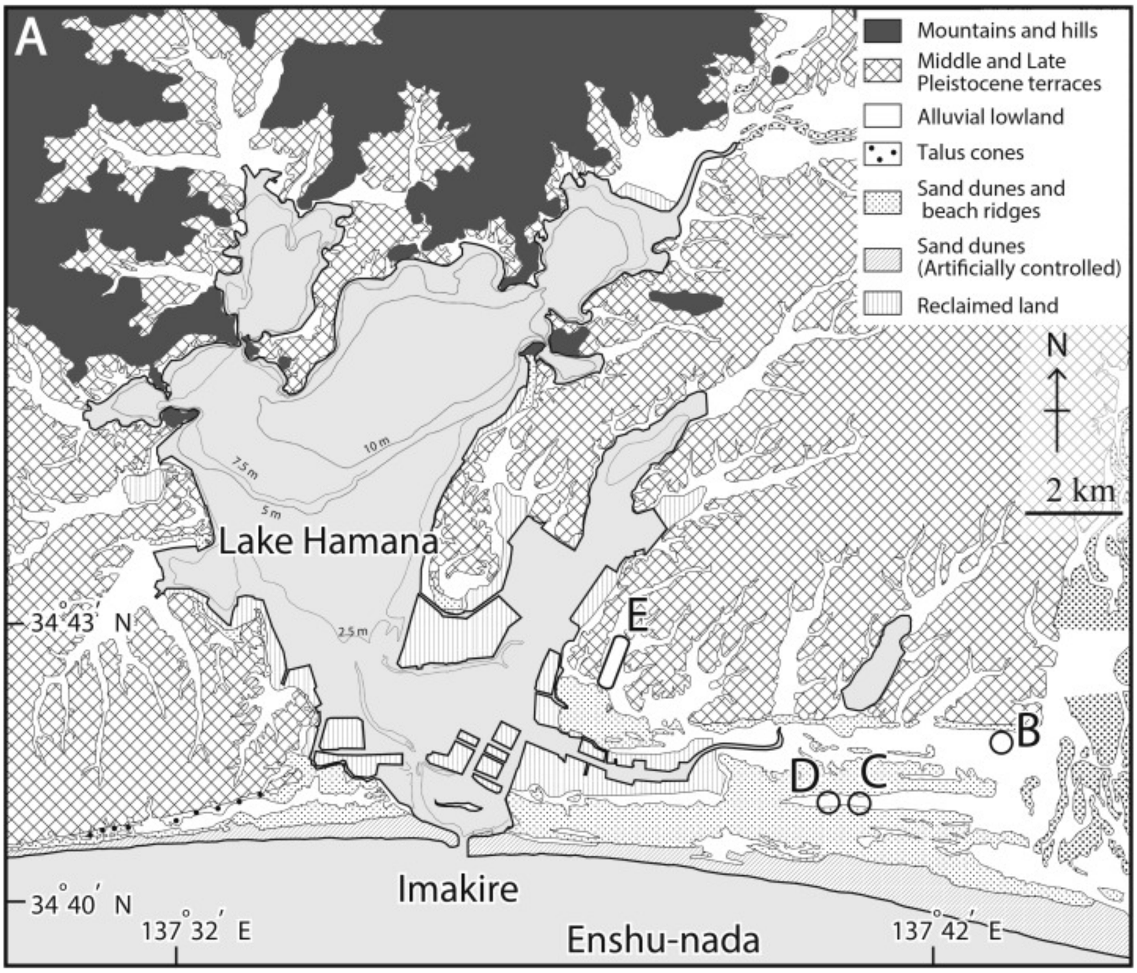
Site 14: Shishikura et al. (2008a), Site 15, 16: Tsuji et al., (2002), Site 19: Takada et al., 2002, Site 22: Fujiwara et al. (2013b), Site 26: Fujiwara et al. 2007, Site 28: Kitamura et al. 2013), Site 32: Kitamura et al. (2014).

Fig. 3. Photographs showing examples of tsunami deposits

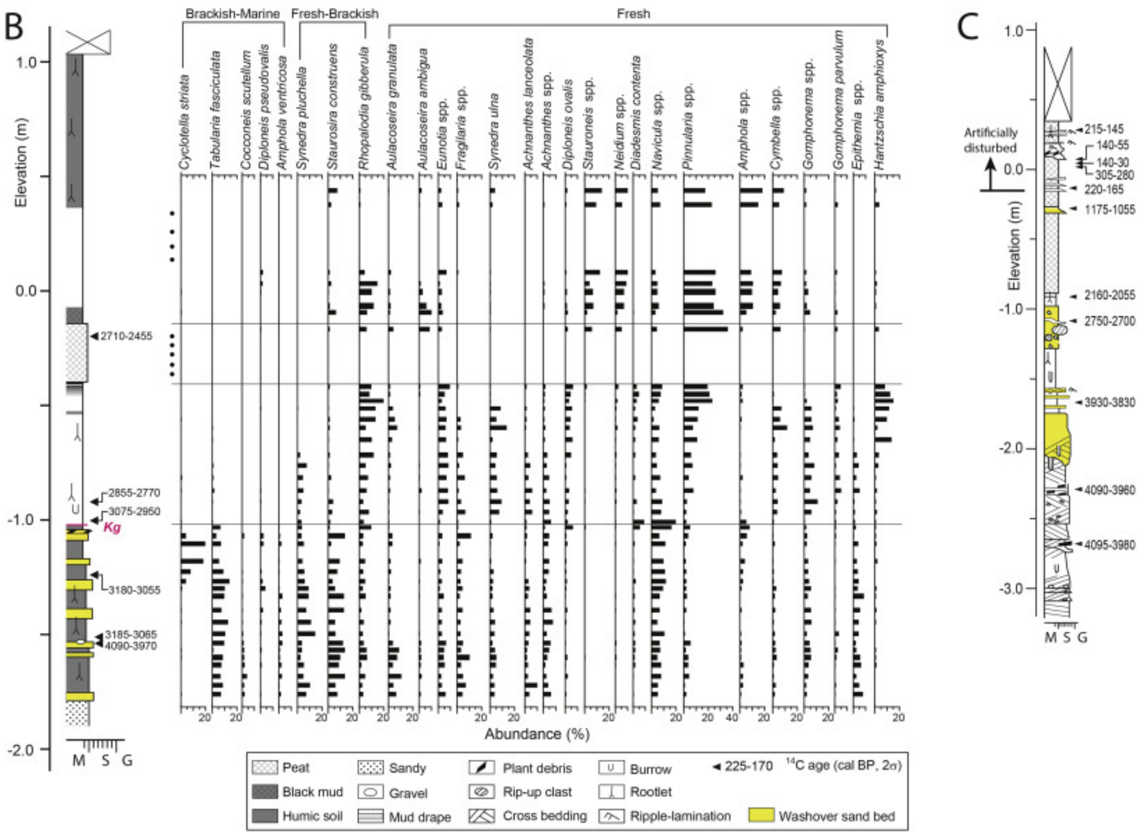
A) Overview of four historical tsunami deposits at the Ōtagawa Lowlands, facing the Tōkai segment of the Nankai Trough. Stratified sandy tsunami deposits are distinguished as horizontal color bands, which serve as an aquifer, in back-marsh mud.

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

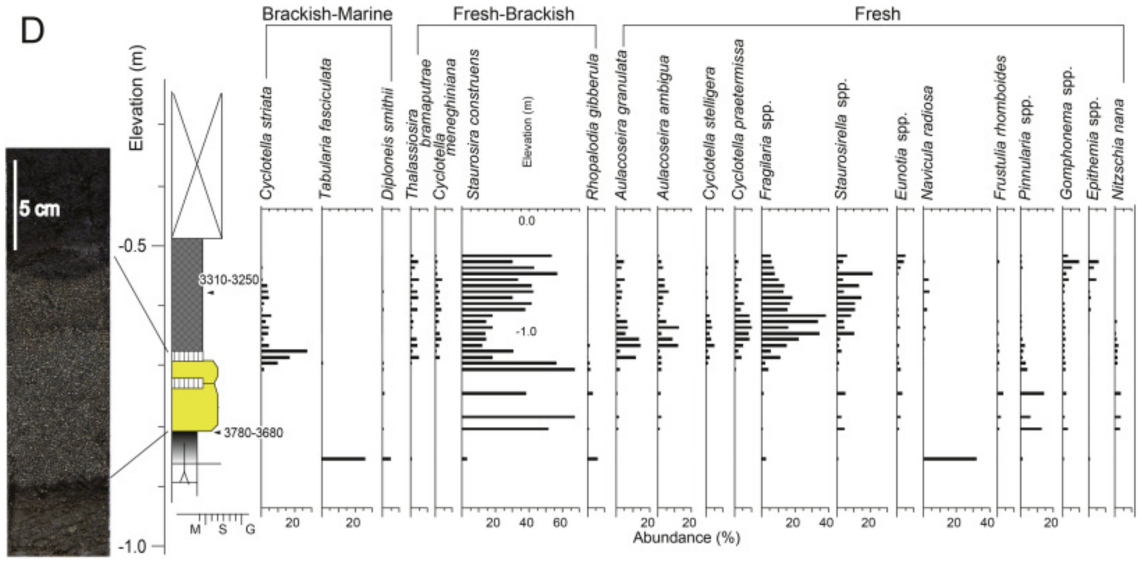
2001


A
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
E
B
D
C
34°43′ N
34°40′ N
137°32′ E
137°42′ E
Imakire
Enshu-nada

2002


B
Brackish-Marine
Fresh-Brackish
Fresh
Elevation (m)
Cyclotella striata
Tabularia fasciculata
Cocconeis scutellum
Diploneis pseudovalis
Amphora ventricosa
Synedra pluchella
Staurosira construens
Rhopalodia gibberula
Aulacoseira granulata
Aulacoseira ambigua
Eunotia spp.
Fragilaria spp.
Synedra ulna
Achnanthes lanceolata
Achnanthes spp.
Diploneis ovalis
Stauroneis spp.
Neidium spp.
Diadesmis contenta
Navicula spp.
Pinnularia spp.
Amphora spp.
Cymbella spp.
Gomphonema spp.
Gomphonema parvulum
Epithemia spp.
Hantzschia amphioxys
2710-2455
2855-2770
3075-2950
Kg
3180-3055
3185-3065
4090-3970
M S G
Abundance (%)
C
Artificially disturbed
215-145
140-55
140-30
305-280
220-165
1175-1055
2160-2055
2750-2700
3930-3830
4090-3960
4095-3980
Peat
Sandy
Plant debris
Burrow
225-170
^{14}C age (cal BP, 2σ)
Black mud
Gravel
Rip-up clast
Rootlet
Humic soil
Mud drape
Cross bedding
Ripple-lamination
Washover sand bed

2003


D
5 cm
Elevation (m)
3310-3250
3780-3680
M S G
Brackish-Marine
Fresh-Brackish
Fresh
Cyclotella striata
Tabularia fasciculata
Diploneis smithii
Thalassiosira bramaputrae
Cyclotella meneghiniana
Staurosira construens
Elevation (m)
Rhopalodia gibberula
Aulacoseira granulata
Aulacoseira ambigua
Cyclotella stelligera
Cyclotella praetermissa
Fragilaria spp.
Staurosirella spp.
Eunotia spp.
Navicula radiosa
Frustulia rhomboides
Pinnularia spp.
Gomphonema spp.
Epithemia spp.
Nitzschia nana
Abundance (%)

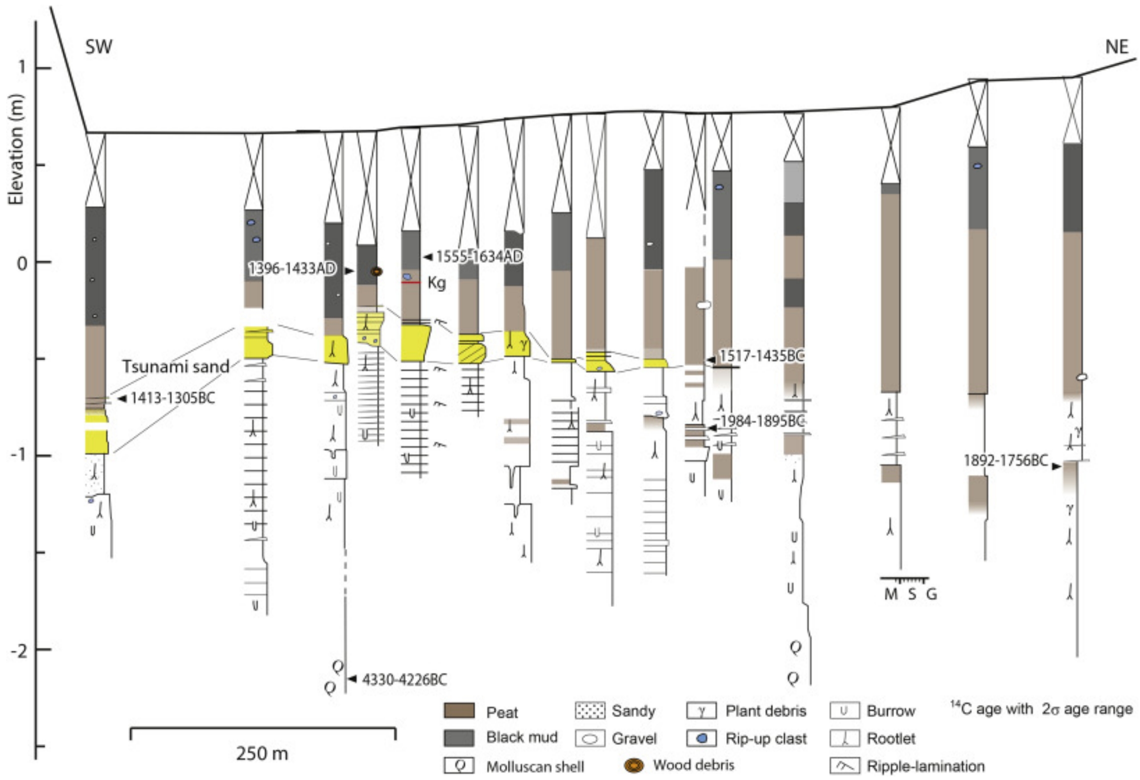


Fig 4. Paleotsunami deposits in the western Hamamatsu Plain

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.

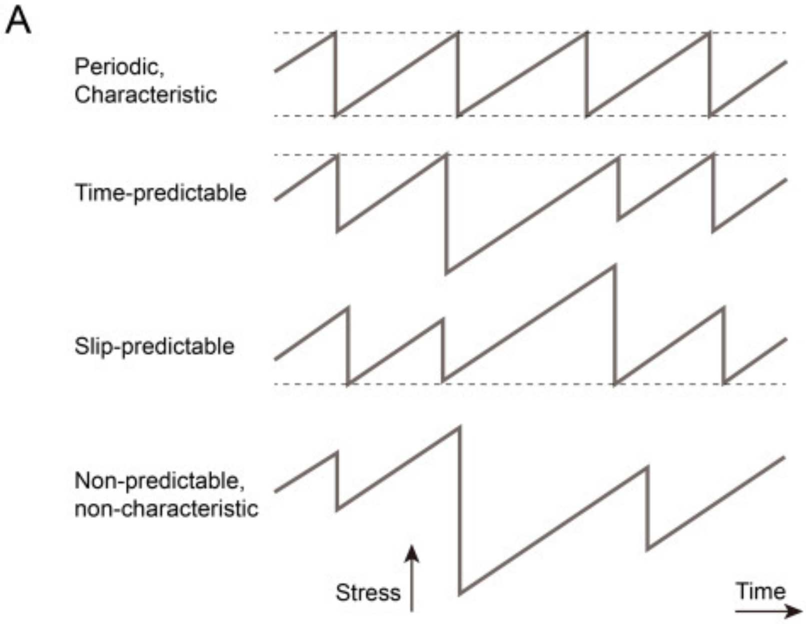
C) Geological columnar section from Site C. The occurrence of possible washover sand beds is limited to the period older than 1175-1055 cal BP (775-895 CE).

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.

Figs. 4B to 4D and Fig. 4E are modified from Sato et al. (2016a) and Fujiwara et al. (2013a), respectively.

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			?							

Fig. 5 Schematic illustration of proposed recurrence models for subduction zone earthquakes A) modified from Satake and Atwater (2007), B) simplified from Seno (2012).

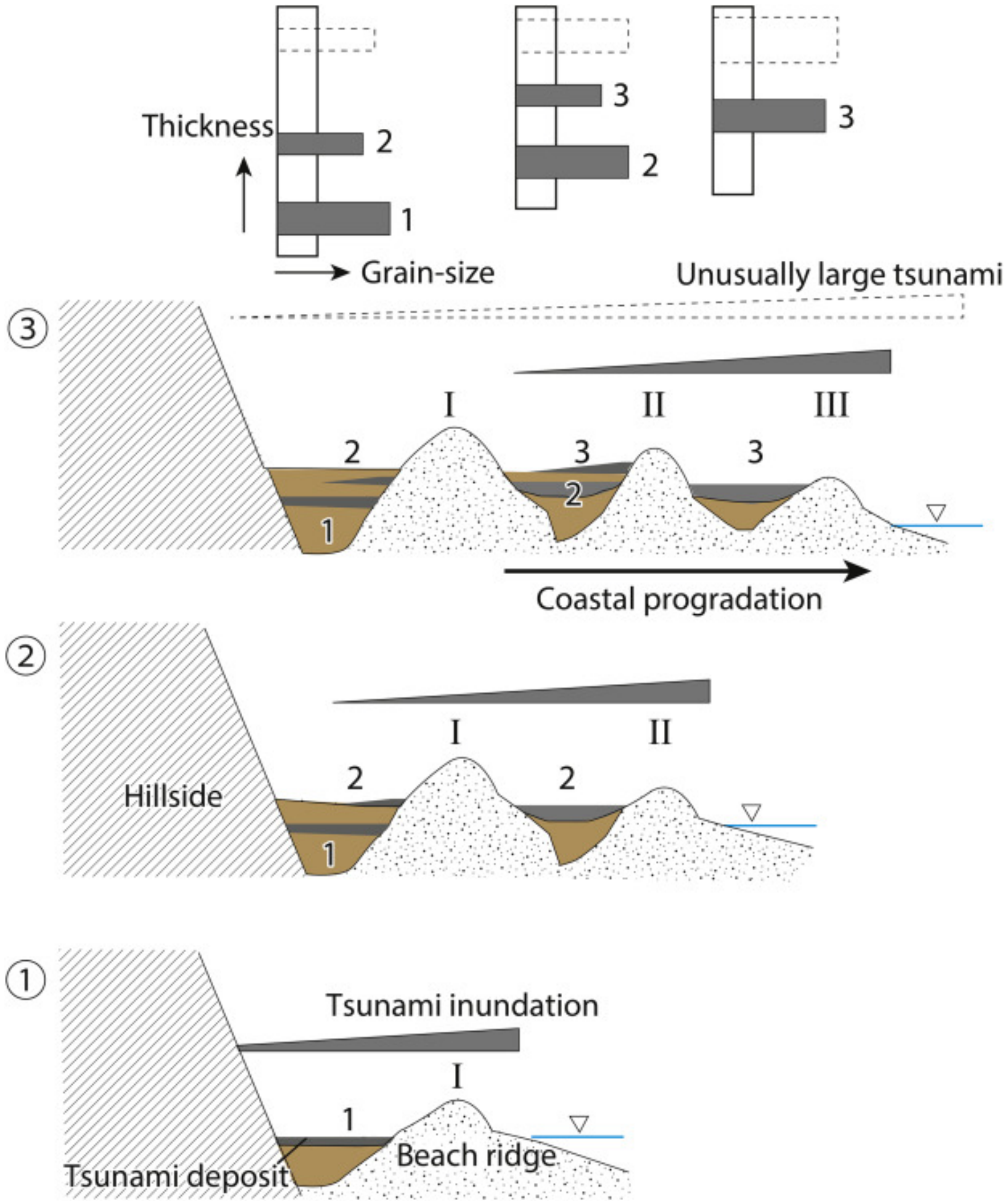


Fig. 6. Conceptual model illustrating the reconstruction of paleotsunami size from the spatial distribution of tsunami deposits considering the geomorphic evolution of a strand plain

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

		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										
	1944 Showa										
	1911 Kikai										
	1854 Ansei										
	1854 Ansei										
	1791 event										
	1771 Meiwa										
	1707 Hoei										
	1614 Keicho										
	1498 Meio						△	△			
	1361 Shohei										
	1361 Shohei										
	1099 Kowa										
	1096 Eicho										
	887 Ninna										
	684 Hakuho										
	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
	…										

	Shaking+tsunami+coastal deformation	△ Liquefaction	
	Shaking+tsunami		tsunami
	Shaking+coastal deformation		

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