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Is the Ryukyu subduction zone in Japan coupled or decoupled? —The necessity of seafloor crustal deformation observation

Masataka Ando[1], Mamoru Nakamura[2], Takeshi Matsumoto[2], Masahide Furukawa[2], Keiichi Tadokoro[3], and Muneyoshi Furumoto[3]

[1]*Institute of Earth Sciences, Academia Sinica, Taiwan*
[2]*Department of Earth and Material Sciences, University of Ryukyus, Japan*
[3]*Graduate School of Environmental Sciences, Nagoya University, Japan*

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The 2004 Sumatra-Andaman earthquake of M_w 9.3 occurred in a region where a giant earthquake seemed unlikely from the point of view of tectonics. This clearly implies that our current understanding of strain accumulation processes of large earthquakes at subduction zones needs to be reexamined. The Ryukyu subduction zone is one such zone since no large earthquake has been anticipated there for reasons similar those pertaining to the Sumatra-Andaman arc. Based on our analysis of historical earthquakes, plate motion, back-arc spreading, and GPS observation along the Ryukyu trench, we highly recommend monitoring seafloor crustal deformation along this trench to clarify whether a large earthquake ($M_w > 8$) could potentially occur there in the future.
Key words: Subduction zones, Ryukyu trench, decoupled/couple subduction, back-arc spreading, seafloor crustal deformation.

1. Introduction

Japan is surrounded by the major subduction plate boundaries with the Pacific plate, Philippine Sea plate, and Amur plate, as shown in Fig. 1 (Wei and Seno, 1998; Heki *et al.*, 1999; Taira, 2001). All earthquakes above M_w 8.0 have occurred along the subduction boundaries between these plates. In 2002 and 2004, the Diet of Japan passed legislation approving measures for disaster preparedness related to large earthquakes ($M_w > 8.0$) along the Nankai trough and Kuril trench. However, no similar action has been undertaken for the Ryukyu trench since it is widely believed among earth scientists that the subduction along this trench is possibly aseismic and not associated with large earthquakes. In fact, no large earthquake has been reported along the Ryukyu subduction for the last 300 years, with the exception of two events, which we discuss in a later section of this article.

We therefore highly recommend a reexamination of the subduction along the Ryukyu trench to ascertain if the subduction along this arc is truly aseismic. We likewise propose carrying out a seafloor crustal deformation observation as an effective approach to resolve this issue.

2. Sumatra-Andaman Arc

The main factor that motivated this study was the occurrence of the 2004 M_w 9.3 Sumatra-Andaman earthquake. Since this $M_w > 9.0$ earthquake had never been anticipated for the Sumatra-Andaman subduction prior to its occurrence on 26 December 2004, it is worthwhile to reexamine the possibility of such earthquakes along the Ryukyu subduction. There are four main reasons why this giant earthquake was not expected for the Sumatra-Andaman arc.

2.1 Historical earthquakes

Bilham *et al.* (2005) reported that there have been three large earthquakes in the north Sumatra-Andaman arc during the past 160 years: M 7.5–7.9 on October 31, 1847; M 7.9±0.1 on December 31, 1881; M 7.7±0.1 on June 26, 1941. All other events were smaller than these events. Ortiz and Bilham (2003) presumed a dislocation model of the 1881 earthquake based on the tsunami records and estimated its magnitude at M 7.5–7.8. Consequently, there was no anticipation that events $M \geq 8.0$ would ever occur in the region.

In northern Sumatra, the 17th century marked Aceh's Golden Age under Sultan Iskandar Muda Meukuta Alam. During his reign, Iskandar Muda controlled the Straits of Malacca by consolidating the northern pepper ports of Sumatra to centralize all trade at Banda Aceh (Dexter, 2004). However, no record or legend describing a giant tsunami similar to the 2004 event has been found. This implies that during the last 400 years, no giant tsunami has struck Banda Ache.

2.2 Convergence rate and age of oceanic plate

Based on the historical occurrences of earthquakes, the subduction rate is considered to be one of the main factors indicating whether or not a giant earthquake could occur along subduction zones. Examples of major earthquake events ($> M_w$ 8.5) along subduction zones with a high convergence rate (Fig. 2) are the 1960 M_w 9.5 Chilean earthquake region (11 cm/yr), the 1964 M_w 9.2 Alaska earthquake region (6 cm/yr), and the 1963 M_w 9.0 Kamchatka earthquake region (9 cm/yr). Since the Indo-Australia plate

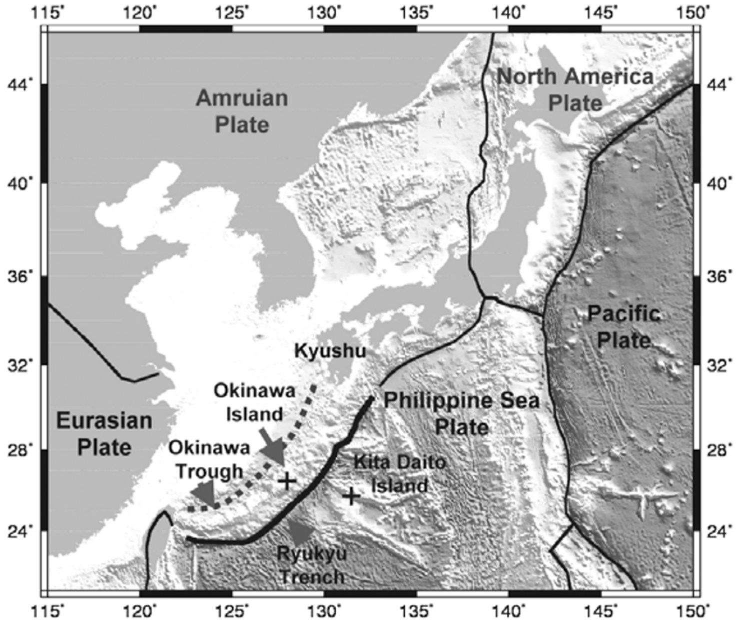

Fig. 1. Plate boundaries around the Japan islands. The Amurian plate boundary is according to Heki *et al*. (1999). Locations of Okinawa Island in the Eurasian plate and Kita Daito Island in the Philippine Sea plate are indicated with an "+". The Okinawa trough and the Ryukyu trench are shown with dashed and solid lines, respectively.

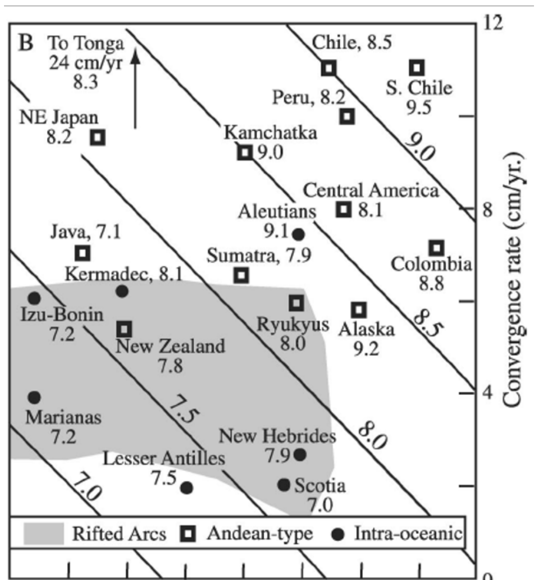

Fig. 2. Velocity versus age of the oceanic plate, showing the nature of subduction of oceanic plate. The Sumatra-Adaman arc falls near the rifted arc region in this diagram. The convergence rate of the Ryukyu arc in this figure is slightly underestimated compared with about 80 mm/yr as discussed in the text. The number at each subduction zone is the associated maximum M_w (seismic moment magnitude), and the contours of constant M_w are defined by regression analysis. Shaded area outlines subduction zones associated with back-arc spreading or interarc rifting. Note that most of the largest earthquakes occur at Andean-type margins. (From Stern, R. J., Subduction zones, *Rev. Geophys*., **40**(4), 1012, doi:10.1029/2001RG000108, 2002. Copyright 2002 American Geophysical Union. Reproduced by permission of American Geophysical Union.)

subducts the northern Sunda-Sumatra plate almost parallel to the trench axis, the subduction rate normal to the trench axis in the northern area is only 1.4 cm/yr (Bilham *et al*., 2005). Thus, with such a low subduction rate, a giant earthquake in this region was not anticipated.

Furthermore, the age of the oceanic plate in the north Sumatra-Andaman area is about 60–80 Ma (Müller *et al*., 2006). Figure 2 shows the subduction velocity versus the age of the oceanic plate (taken from Stern, 2002). The plot of convergence rate versus lithosphere age shows the strong influence of this relationship on seismicity, as modified according to Ruff and Kanamori (1983). The subduction zones are classified in this study as Andean-type or Mid-oceanic type, with the former subducting a continental edge and the latter subducting a mid-oceanic plate. They are almost equivalent to the Chilean and Mariana types described by Uyeda and Kanamori (1979). Thus, the Sumatra-Andaman subduction can be classified as a mid-oceanic type (or Mariana type), where compressional stress is low, and no giant thrusting earthquake is anticipated.

2.3 Back-arc spreading

Back-arc basin spreading is progressing along the Andaman arc (Curry *et al*., 1979; Eguchi *et al*., 1979). However, it is not the typical back-arc spreading, such as that observed in the Mariana, Fiji-Tonga, and Ryukyus arcs, where the spreading direction is almost normal to the trench axis. Rather, the direction of the back-arc spreading in the Andaman arc is almost parallel to the trench axis. Eguchi *et al*. (1979) suggest that the spreading is caused by crustal stretching due to the collision of the Eurasia plate against the Indian plate. Nevertheless, it is certain that the arc is characterized by a tensional regime rather than by a compressional one, as found in the well-coupled region of the Chile, Alaska, and Japan arcs.

2.4 Remarks

The abovementioned factors provide some support to the idea that a giant earthquake could not be anticipated for the northern Sumatra-Andaman arc. Nevertheless, it did happen with unimaginable tragic consequences in 2004. This clearly suggests that a 400-year history and tectonic environmental data alone do not provide earth scientists with sufficient information for estimating a maximum earthquake size for a subduction zone. This is the most important lesson learned from the 2004 Sumatra-Andaman earthquake.

3. Ryukyu Trench

The tectonics of the northeastern Asian margin is characterized by two major factors: (1) the subduction of oceanic plates, such as the Pacific plate (PA) and the Philippine Sea plate (PH); (2) the regional tectonics and seismicity of the Amur plate (AM), which covers northeast China, the Korean peninsula, the Japan Sea, and southeastern Russia (Heki *et al.*, 1999). The Ryukyu arc is located southeast of the Eurasian plate and north of the Philippine Sea plate (Fig. 1). It extends 1400 km from southwestern Kyushu Island in the northeast to east of Taiwan in the southwest. Northwest of the Ryukyu Arc is the Okinawa Trough, which is now in an extensional stage within the continental lithosphere of the Eurasia plate (Kimura, 1985; Sibuet *et al.*, 1987). The regional stress field caused by the motions of these plates produces intense seismic activity and active crustal movement in the region.

3.1 Historical earthquakes

Figure 3 shows a map of destructive earthquakes that have occurred along the Ryukyu trench since 1700, taken from the catalogue of Utsu (1989). This catalogue was mainly based on historical records of damage caused by strong ground motions and tsunamis. Two remarkable events are apparent on this map, both of which likely occurred near the subduction boundary: the 1911 earthquake near Amami Island (Event 1 in Fig. 3) and the 1771 large tsunami off Ishigaki Island (Event 2 in Fig. 3).

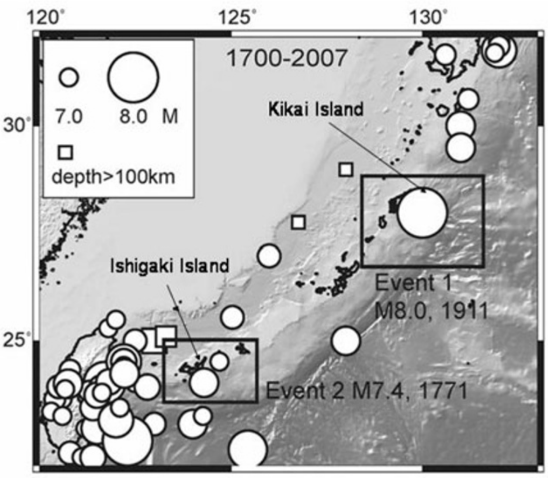

Fig. 3. Map of destructive earthquakes since 1700 along the Ryukyu trench based on seismic data compiled by Utsu (1989). See the text for Event 1 and Event 2.

The 1771 event occurred off Ishigaki Island accompanied by a large tsunami (Utsu, 1989; Usami, 1996) that was initially believed to be as high as 80 m (Makino, 1968), but which was later corrected to 30 m (Kato, 1986). The distribution map of inundation heights revealed a localized feature that high waves appeared only in the east and southeast shores of the island. It should be noted that this pattern can not be explained by a tsunami caused by a large subduction earthquake ($M_w \geq 8.0$). Nakamura (2006) proposed that the tsunami was most probably caused by a shallow normal fault running east of Ishigaki Island with a fault length of 30 km and $M_w = 7.0$.

The 1911 of M_w 8 occurred off Kikai Island on June 15 of that year. During this earthquake, 422 houses on the island were completely destroyed and there were 12 casualties (Usami, 1996). The earthquake was felt even up to central Honshu and over an area with an epicentral distance of 1,500 km. However, the tsunami generated by this event was small and caused little damage. Konishi and Sudo (1972) proposed that this earthquake was due to a normal fault, possibly an intraplate event of the Philippine Sea slab. Taking these data into account, Usami (1996) inferred its depth as 100 km.

Given the explanations for these two earthquakes and the absence of any record describing an extensive strong ground motion and tsunami related to a great thrusting earthquake along the Ryukyu trench, it is believed that the subduction in this area is aseismic and/or stable sliding.

3.2 Convergence rate and age of oceanic plate

Although a number of models of plate-motion have been presented, only a few include the Philippine Sea plate as a constitutive plate in their model. Among these latter models are NUVEL-1A (DeMets *et al.*, 1994), REVEL (Sella *et al.*, 2002), and GSRM (Kreemer *et al.*, 2003). The NUVEL-1A and GSRM models are constructed on the basis of geological data, such as geomagnetic anomalies and the configuration of transform faults and trenches. REVEL, on the other hand, is based on GPS velocity data. We have calculated the relative plate motion between the Eurasian and Philippine Sea plate, near Okinawa Island, at around 26°N, 129°E, for the three models. The relative slip rate obtained is 82 mm/yr for the NUVEL-1A model, 80 mm/yr for the GSRM model, and 74 mm/yr for the REVEL model (Table 1). Clearly, the results from these three models do not differ to any great extent.

Ruff and Kanamori (1983) determined that the coupling condition is correlated with convergence rate and lithosphere age and that back-arc spreading is associated with subduction zones where there is a low level of coupling between the upper and lower plates. Convergence rate and lithosphere age therefore influence the seismic coupling in the subduction zone. The age of an oceanic plate falls within the range 60 and 80 Ma (Müller *et al.*, 2006) and the convergence velocity is 74–82 mm/yr. Therefore, based on our model calculations, the Ryukyu subduction falls in the higher range of convergence rates in a region of the Andean-type (Stern, 2002).

3.3 Back-arc spreading

Uyeda and Kanamori (1979) classified subduction zones into two types depending on whether or not they were asso-

Table 1. Relative velocity and azimuth for relative plate motion, GPS baseline length change, and back-arc rifting. The plate motion is for the Philippine Sea plate (PH) relative to the Eurasia plate (EU) in and around Okinawa Island.

	V_{plate}			$V^{\mathrm{d}}_{\mathrm{GPS}}$	$V^{\mathrm{e}}_{\mathrm{rift}}$
	NUVEL-1[a]	GSRM[b]	REVELA[c]		
Direction (deg)	331	329	330	310	180
Velocity (mm/yr)	82	80	74	87	10
Velocity, arc-normal (mm/yr)	80	79	73	86	7

[a]DeMets *et al.* (1994), [b]Sella *et al.* (2002), [c]Kreemer *et al.* (2003), [d]Velocity of GPS baseline length changes between islands of Okinawa and Kita Daito island, [e]Velocity of back-arc basin rifting (Nakamura, 2006).

ciated with actively opening back-arc basins. One extreme case is the Chile subduction, where the plate motion is seismic. The other extreme case is the Marianas subduction, where the plate motion is aseismic. The stress state in the back-arc area is compressional in the former and tensional in the latter. The nature of the contact zone between the upper and lower plates changes from tight coupling (Chile) to decoupling (the Marianas). The decoupling results in an oceanward retreat of the trench and back-arc opening.

Geological and geophysical data reveal that the Okinawa Trough shows incipient continental rifting with crustal separation that started from about 2 Ma (Kimura, 1985). Linear magnetic anomalies and high-velocity crustal rocks associated with the central trough region are interpreted as being due to magmatic intrusion associated with crustal separation that occurred between the Pliocene and recent times (Lee *et al.*, 1980).

3.4 Remarks

The current geological state at the Ryukyu trench subduction should be taken into account. Although the convergence rate is high, there is no clear evidence that suggests any indication of a large earthquake occurring.

Scholz and Campos (1995) propose a model of force balance at a subduction zone that can predict three regimes: (1) seismically coupled compressional arcs with advancing upper plates; (2) seismically decoupled extensional arcs with retreating upper plates; (3) strongly extensional arcs, which also have back-arc spreading. They have applied this model globally and found that it can successfully predict the state of seismic coupling and back-arc spreading in more than 80% of the world's subduction zones. However, the model clearly fails to interpret some regions. The Ryukyu arc is one of these exceptions. It seems to be clearly decoupled and has active back-arc spreading in the Okinawa trough, neither of which is predicted by the model, although the model does indicate that the Ryukyu arc is moderately compressive. These authors attributed this anomaly to the problem that the model treats subduction zones as isolated mechanical systems, whereas the Ryukyu arc is strongly coupled to the Izu Bonin-Mariana system on the other side of the Philippine Sea plate. This implies that in terms of its potential for a large earthquake, the Ryukyu arc is still puzzling to earth scientists given its coupling behavior at the subduction.

Conversely, this situation is similar to the Sumatra-Andaman arc prior to the occurrence of the 2004 Sumatra-Andaman earthquake. Thus, we strongly believe that the Ryukyu trench and its tectonic situations warrant serious discussion given the possible risk of giant earthquakes that would impact on not only the southern Japanese islands but also on several adjacent and distant countries fronting the Pacific.

4. Crustal Deformation Due to a Slab Subduction

In the case of a decoupled (unlocked) subduction, the lower plate is subducting freely without dragging and pushing the upper plate. Consequently, when monitoring a distance between two arbitrary points across the trench along a decoupled subduction, the plate-motion-parallel component should equal the plate convergence rate since no internal deformation exists in the upper plate (Fig. 4(a)). In contrast, in the case of a coupled (locked) subduction, the plate-motion-parallel component between two arbitrary points across the trench becomes shorter than the plate convergence rate (Fig. 4(b)). The plate-motion-parallel component is a function of distance from the trench axis because the wedge in the hanging-wall side is dragged and pushed by the subducting slab, resulting in backward displacements in the hanging-wall side away from the trench axis.

The GPS velocity vectors of GEONET (Miyazaki *et al.*, 1998) on the Ryukyu Islands for the 10-year period between 1997 and 2007 are shown in Fig. 5: the maximum shortening axis is 310° and its rate is 87 mm/yr. These vectors are referred to Kita Daito located on the Philippine Sea plate. The velocity vector increases along the Ryukyu Islands from the northeast to southwest, which is mainly

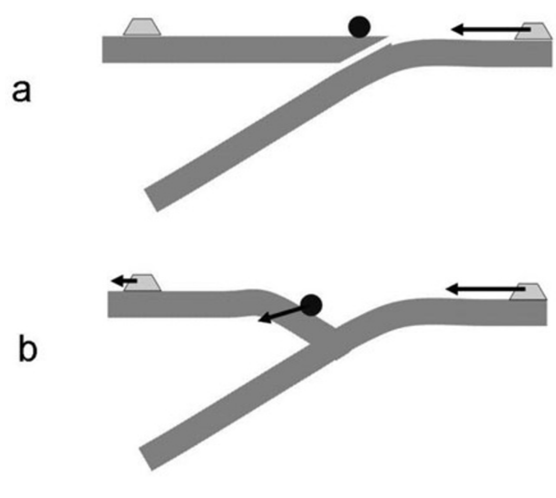

Fig. 4. (a) Schematic view of the coupled subduction where the lower plate subducts freely without dragging the upper plate. Hence, the solid black circle above the interface stays at the same place relative to the island on the left on the upper plate. (b) In the coupled case, the solid circle above the interface is pushed backward and also dragged downward.

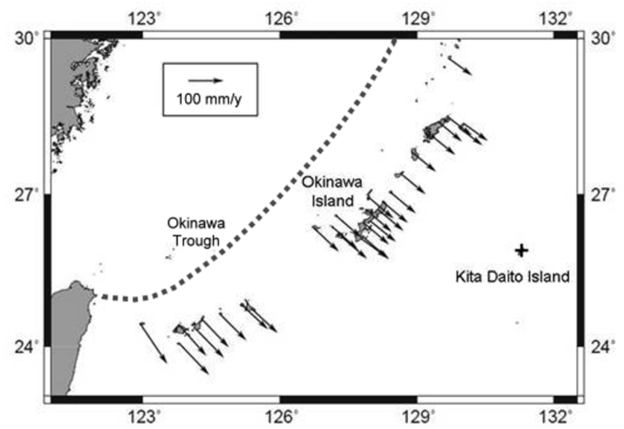

Fig. 5. GPS velocity vectors on the Okinawa islands between 1997 and 2007. We used data between 26°N and 29°N for comparing the plate motions. The reference point Kita Daito Island is shown with an "+".

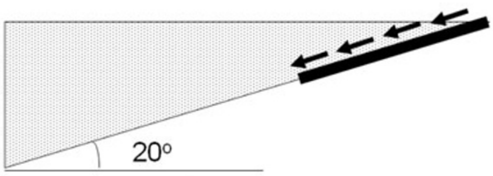

Fig. 6. Outline of the finite element calculation for crustal deformation caused due to drag by slab subduction. The calculation area is 400 by 136.8 km, with a dip angle of 20°. Along the plate interface, a unit slip 100 is given parallel to the coupled interface (thick line with arrows), and the remaining part of the interface below the coupled interface remains fixed without motion.

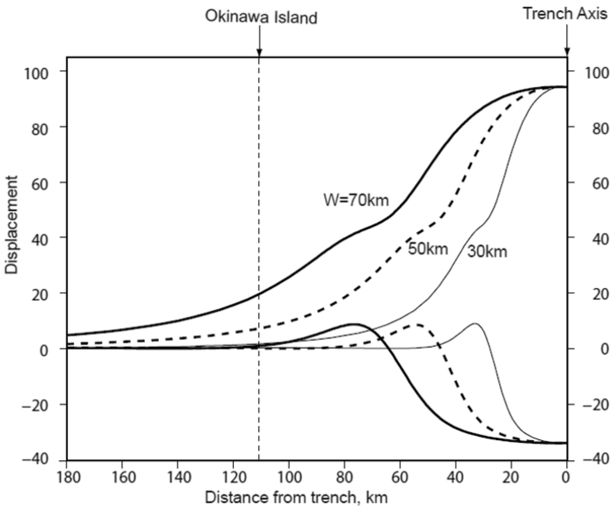

Fig. 7. Horizontal and vertical displacements at the free surface of the hanging-wall side due to a unit slip of 100. Horizontal displacements are normal to the trench axis in the hanging wall by locked zones with widths of 30, 50, and 70 km on the plate subduction interface. The positive direction is the direction away from the trench axis.

caused by the back-arc spreading of the Okinawa Trough (Nakamura, 2004). For this reason, we select GPS data on the islands from 26° to 29°N, where the rifting rate is likely to be uniform.

To estimate coupling effects on the internal deformation of the hanging-wall side, we calculate the horizontal and vertical displacements due to a slab drag by using a two-dimensional finite-element method. Figure 6 shows the outline of the calculation field. The triangle is the calculation field with a horizontal side of 400 km, a vertical side of 136.8 km, and a dip angle of 20° (according to Nakamura and Kaneshiro (2000), who determined the dip angle of the slab from the deep seismic zone).

The left boundary is assumed to be fixed without deformation. The slab interface is divided into slip and non-slip interfaces. The former is dragged parallel to the interface without a normal component, and the latter is fixed along the interface where the displacement is confined. This configuration is basically the same as the back-slip model by Savage (1983), although the back-slip model is associated with a slip of its foot-wall side to some extent.

Figure 7 shows the horizontal and vertical displacements at the free surface in the hanging-wall side for three cases—coupled interfaces of 30, 50, and 70 km, respectively. The coupled interface is assumed to start from the trench axis, which has little effect on the displacement pattern. In the calculation, the positive horizontal direction is taken to be away from the trench axis. As shown in Fig. 7 the vertical displacement is approximately one third of the horizontal component, and in this study, we mainly use the horizontal component for the evaluation of the coupling condition of the plate interface.

Okinawa Island is located 110 km from the trench axis. Given a unit slip of 100 at the coupled interface, Okinawa is displaced at this step by 1.0 for the case of the 30-km decoupled width, by 6.3 with the 50-km decoupled width, and by 18.8 with the 70-km decoupled width. If such landward horizontal motions are detectable with high accuracy, we can elucidate the interface condition in terms of whether it is coupled or not and, in addition, estimate the width of the coupled interface that could lead to a future large earthquake. However, for the evaluation of subduction mode, decoupled or coupled, we additionally need two parameters: (1) a back-arc spreading component; (2) a relative plate motion across the trench axis. Consequently, in the next section we discuss those effects that act on Okinawa Island.

5. Decoupled or Coupled Subduction?

Referring to the previous discussions, we try to clarify whether the Ryukyu subduction is coupled (locked) or decoupled (unlocked). In this part of the study, we consider the slab traction force to be normal to the arc axis and thus focus on the arc-normal component of all deformation velocities. If the two plates are decoupled, the rate in distance between a point on the hanging-wall side and another point in the oceanic plate oceanward of the trench axis, its arc-normal component (V_{length}), equals the arc-normal relative plate motion (V_{plate}), as shown in Fig. 8.

On the contrary, if the two plates are coupled, an arbitrary point on the hanging-wall side is pushed backward from the slab (V_{back}). If we measure the rate in distance between a point on the hanging-wall side and another point in the oceanic plate oceanward of the trench axis, its arc-normal component (V_{length}) is smaller than the arc-normal relative

Table 2. Backward velocities at Points A and B for three plate motion models.

Plate model	NUVEL-1[a]			GSRM[b]			REVELA[c]		
Coupled width[d]	30 km	50 km	70 km	30 km	50 km	70 km	30 km	50 km	70 km
V_a^e (mm/yr)	87	82	72	86	81	71	80	75	66
Δd^f (mm/yr)	−1	4	14	0	5	15	6	11	20
Point A[g] (mm/yr)	14	35	—	14	35	—	13	32	—
Point B[h] (mm/yr)	57	70	—	56	69	—	52	64	—

[a]DeMets *et al.* (1994), [b]Sella *et al.* (2002), [c]Kreemer *et al.* (2003), [d]See Fig. 8, [e]V_a obtained from Eq. (1), [f]Δd obtained from Eq. (2), [g]Velocity at Point A in Fig. 10, [h]Velocity at Point B in Fig. 10. The case of width of 70 km is excluded from the estimation of displacements at Points A and B because of its large misfit (Δd) as described in the text.

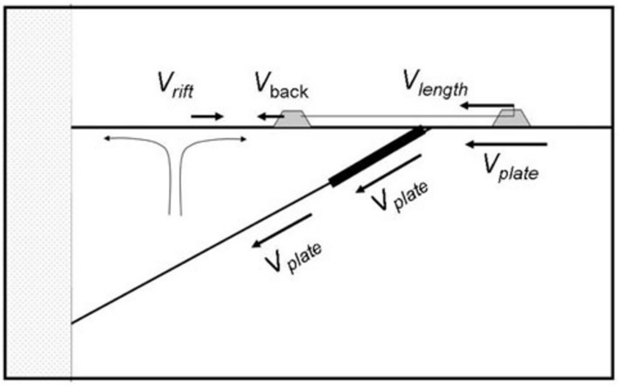

Fig. 8. The deformation due to a coupled subduction. $V_{\rm back}$ is the backward velocity caused on an island by the slab subduction. $V_{\rm length}$ is the relative GPS-observed velocity of the upper plate to the lower plate, $V_{\rm plate}$ is the relative plate velocity, and $V_{\rm rift}$ is the back-arc spreading rate.

plate motion ($V_{\rm plate}$). In addition, it should be taken into account that the Okinawa Trough is rifting (Nakamura, 2004) and that the islands are being pushed oceanward, away from the rifting axis of the Okinawa Trough ($V_{\rm rift}$). Thus, the rate balance is shown as in Fig. 8 and expressed as:

$$V_{\rm a} = V_{\rm plate} + V_{\rm rift} - V_{\rm back} \tag{1}$$

$$\Delta d = V_{\rm length} - V_{\rm a}, \tag{2}$$

where Δd is the misfit.

The directions obtained from the plate motions, GPS velocities on the islands, and back-arc spreading, respectively, are somewhat different. However, these differences are very small and are roughly perpendicular to the trench axis of 48°. Thus, we compare the velocities perpendicular to this angle and parallel to 138°. In the following, we describe all velocities ($V_{\rm rift}$, $V_{\rm a}$, $V_{\rm plate}$, $V_{\rm rift}$, $V_{\rm back}$, and $V_{\rm length}$) in terms of mm/yr instead of the normalized value since a relative plate motion ($V_{\rm plate}$) is different among the models, and we need to discuss velocity balance in mm/yr.

The three models of plate motion mentioned above yield $V_{\rm plate} = 74 \sim 82$ mm/yr. Since the rifting rate ($V_{\rm rift}$) is 10 mm/yr near Okinawa Island (Nakamura, 2004) along 180°, the result would be 7 mm/yr along the reference direction of 138° (Table 1). Therefore, $V_{\rm plate} + V_{\rm rift}$ (74∼82 mm/yr + 7 mm/yr) is close to $V_{\rm length}$ (86 mm/yr), suggesting the absence of a coupling interface or a narrow one. As summarized in Table 2, in cases of a coupled width of 30 and 50 km, respectively, at the interface, $V_{\rm back} = 1$ mm/yr and 5 mm/yr are obtained at Okinawa Island, respectively. The misfit Δd is −1∼6 and 4∼11 mm/yr, respectively. These values can be acceptable considering the ambiguities involved in the estimation of the back-arc spreading and the plate motions. However, in the case of a coupled width of 70 km, a backward motion of 14∼15 mm/yr is obtained, resulting in a discrepancy 14∼20 mm/yr. Even if we consider the ambiguities in the estimation, the misfit of 14∼20 mm/yr is too large to explain, and this hypothesis of a coupled width of 70 km can be rejected.

If the coupled region exists in the Ryukyu trench of the plate boundary and the slip deficit has accumulated in the interface for the past 400 years, the resultant accumulated slip would amount to 30–33 m. If an earthquake occurs here to release all of the strain at once, this would become an extremely large event. The possibility of a huge tsunami being generated by an earthquake due to a slow slip is high. In the case where only the upper 50 km is considered, this would be a $M_{\rm w}$ 8.5 event along the Ryukyu trench. However, based on currently available GPS data and the known plate motions, the width of the coupled interface in Ryukyu arc is unknown—if it exists at all.

6. Discussion

6.1 Necessity for a new observation

Despite our analysis of currently available GPS observation data and the plate motion models, we still have a very unclear picture of the coupling along the subduction zone fronting the Ryukyu trench. This is mainly due to the scarcity of GPS observation points in this region. Moreover, since the Okinawa islands are located more than 100 km from the trench axis, the crustal deformation due to a plate coupling would have significantly diminished by the time it could be measured on the islands. Thus, the present questions regarding the Ryukyu trench coupling cannot be resolved based on currently available data. It is necessary to obtain crustal deformation data closer to the trench axis where the coupling effect is large enough to distinguish between the two cases—decoupled or coupled subduction (Fig. 7).

In order to overcome the above difficulty, we propose to undertake a seafloor crustal deformation observation near the trench axis. The outline of the measurement system is shown in Fig. 9 and is similar to that used in previous observations (e.g. Spiess *et al.*, 1998; Tadokoro *et al.*, 2006; Ikuta *et al.*, 2008). Initially, a seafloor unit is freely released from a vessel to settle on the seafloor. Using tracking equipment

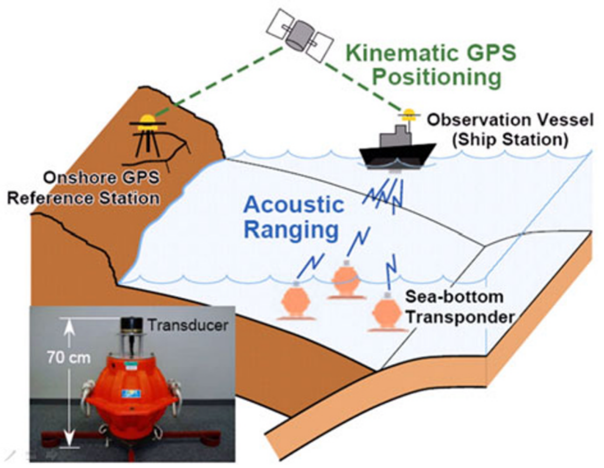

Fig. 9. The outline of the seafloor geodetic observation system according to Tadokoro *et al.* (2006). Insert represents the seafloor unit. See the text for detailed discussion.

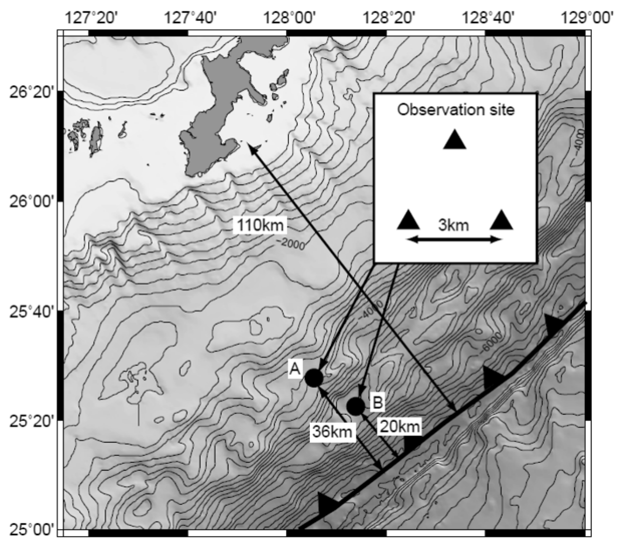

Fig. 10. Planned sites A and B for seafloor geodetic observation based on the results of this study. Each site consists of three seafloor units located with a separation interval approximately the same as the seafloor depth.

on the vessel, the position of the seafloor unit is determined relative to an onland fixed observation point. The position and attitude of the vessel are determined accurately by a kinematic GPS technique. Once the vessel is located accurately, the seafloor unit is located by measuring two-way acoustic travel-times between the vessel and the seafloor unit. The technique in determining the seafloor unit is quite similar to the hypocenter determination in seismology. All of the technical details on the observation system and the analytical methods involved are summarized in Ikuta *et al.* (2008).

Assuming that we deploy seafloor units off Okinawa Island at Point A (36 km from the trench axis) or Point B (20 km from the axis), then the horizontal displacement would be 32∼35 and 64∼70 mm/yr at Point A and B, respectively, for the coupled width of 50 km. These velocity values are derived from the GPS data and plate motion determinations (Table 1) previously discussed. Since the error in determining the horizontal location is 5 cm (Ikuta *et al.*, 2008), the displacement can be measurable within 1∼2 years by the current method of seafloor crustal deformation. However, in the case of a coupled width of 30 km, the displacement at Point A—13–14 mm/yr—would be too small to detect within 1∼2 years. It would therefore take about 5 years to obtain a meaningful result in this kind of deformation. For this reason, it is necessary to do the observations on the trench axis within the proximity of Point B to obtain meaningful results.

In this study, we utilized the horizontal component only. As already mentioned, the positioning accuracy is 5 cm in the horizontal component and worse, possibly twofold higher, in the vertical component (Ikuta *et al.*, 2008). In addition, as shown in Fig. 7, the vertical displacements are approximately threefold smaller than the horizontal displacements. In terms of time, it takes about sixfold longer to detect the vertical component to obtain any significant results. Thus, it is hard to detect vertical displacements of the interseismic stage from the seafloor observation. Consequently, we should initially focus on monitoring the horizontal component and only then study the vertical component as a supplemental dataset.

6.2 Up-dip limit of coupled interface

In this study, the coupled portion on the slab interface is assumed to start at the trench axis. One concern regarding the coupling along the slab interface is the presence of some small amount of friction up to a depth of about 10 km due to high pore pressure in the sedimentary layers. Thus, the results of some studies suggest that the up-dip limit of seismogenic coupling is about 10 km and that most of this is probably due to dehydration of the stable sliding smektite to unstable sliding illite or chlorite (Hyndman and Wang, 1993). This could lead to a likelihood of slow faulting on the shallow interface, which could then affect the seafloor geodetic observation proposed above. This concern was carefully investigated. Figure 11 shows a comparison of

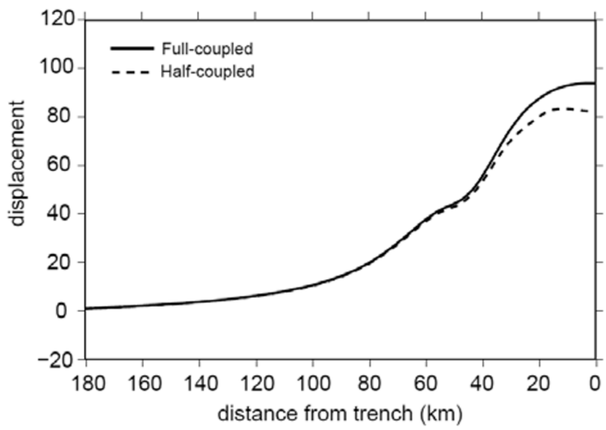

Fig. 11. Horizontal displacement calculated by FEM used in Fig. 4 at a given slip of 100 on the interface. In the full-coupled case, the 50-km interface is coupled; in the half-coupled case, the upper half is allowed to slide freely along the interface and the deeper half is coupled. Note that the resultant crustal deformations are almost identical in both cases for the distance beyond 60 km from the trench.

crustal deformation for two cases: (1) Case 1—the whole interface is coupled; (2) Case 2—only the lower half of the interface is coupled.

These two cases of coupling are calculated by FEM, as shown in Fig. 4. In the half-coupled case (2), the upper half is allowed to slide freely along the interface. In this scenario, the resultant crustal deformations are almost identical in both the vertical and horizontal component with only slight discrepancies. Furthermore, the hanging-wall side of the upper-half interface is pulled by the lower interface and slides down in a fashion similar to that seen in Case 1, where the whole interface is coupled. A number of models (e.g. Gagnon *et al.*, 2005) assume that the deeper interface is coupled, with no relative motion associated in the shallower layer above the interface. Based upon this assumption, an attempt is made to estimate an up-dip limit of the coupled interface. However, this is very unrealistic because the shallow layer where the sediments are dominant can sustain its strength without breaking.

Even when a significant seafloor deformation occurs near the trench axis, the half-coupled and fully-coupled deformations are not distinguishable from the observed seafloor deformation. On the contrary, if the interface is totally decoupled, a backward motion (V_{back}) becomes zero and is distinguishable from cases of coupled interface from the seafloor geodetic observation. Consequently, the two cases, coupled or decoupled, can be still distinguishable from the seafloor observation. If the coupled portion exists, its width would be narrower than the 50-km area discussed earlier.

6.3 Tsunami earthquake

If the coupled portion is found to be very shallow in the Ryukyu trench, attention should be drawn to the fact that this situation is very similar to that of the region along the offshore of Java, which is believed to be an aseismic subduction. Typical examples of earthquake events occurring in this latter region are the M_{w} 7.4 earthquake of June 2, 1994 (Polet and Kanamori, 2000) and the M_{w} 7.3 event of July 17, 2006 offshore western Java (Ammon *et al.*, 2006). These magnitudes, which are based on the tsunami inundation heights, M_{t} (Abe, 1981), are certain to exceed M_{w} 8.0. Both earthquakes occurred very near the trench axis, indication that strain had accumulated even near the trench axis and was released by slow earthquakes.

The occurrence of a large abrupt strain release in the shallow portions of megathrust subduction zone raises another long-standing problem in nature regarding the frictional contact in this depth range. A number of parameters can affect the shallow megathrust frictional environment, including sediment thickness, composition, fluid content and hydrologic properties, and bathymetric irregularities (Ammon *et al.*, 2006). A reasonable assumption is that the existence of sediments along the megathrust is related to the relatively slow rupture propagation and large slip of larger and moderate-size shallow megathrust earthquakes (Kanamori and Kikuchi, 1993).

Thus, aside from the uncertainty of possibly being an aseismic region, the Ryukyu trench may also have the physical properties of the Java trench. If this latter case is true, it is highly probable that a slow slip could be produced in the shallow portions of the Ryukyu trench in the future.

7. Conclusions

The occurrence of the Sumatra-Andaman earthquake highlights the fact that the current limited body of information provided by the records on historical earthquakes and a knowledge of tectonic processes alone is not sufficient to estimate a maximum earthquake size for a subduction zone. Given these uncertainties, we have investigated the Ryukyu trench and its tectonics in order to estimate the probability and risk of future giant earthquakes. Analyses of historical earthquakes, plate motion, back-arc spreading, and GPS observation data along the Ryukyu trench were undertaken. However, currently available information cannot provide any clear answer to the question of whether the interface between the upper and lower plates at Ryukyu trench is coupled or not. Moreover, the data provided by the current GPS network do not allow us to reject or accept the hypothesis that the upper 50 km of the interface is coupled. If the coupled interface exists, it should be very shallow, and it is anticipated that the slow slip at the interface would generate tsunami earthquakes. For these reasons, it is necessary to carry out an observation of seafloor crustal deformation offshore of Okinawa Island using the kinematic GPS and acoustic ranging method. This is the only method that can resolve the present issue of whether the subduction is decoupled or coupled in the Ryukyu subduction zone. From the standpoint of earthquake disaster mitigation, clarification of the subduction process along the Ryukyu trench is very important in order to gain an understanding of the coupling condition that may potentially produce large earthquakes in the future.

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M. Ando (e-mail: ando@earth.sinica.edu.tw), M. Nakamura, T. Matsumoto, M. Furukawa, K. Tadokoro, and M. Furumoto