Geodynamic Context of the Taiwan Orogen

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Four independent arguments suggest that the Ryukyu subduction zone extended from Japan to southwest Taiwan (118°E) from the late Cretaceous to early Miocene (17–18 Ma): i) An analysis of the structure and timing of rifting in the basins of the East Asia continental shelf and west of Taiwan shows that they are located within four belts parallel to the mainland Chinese shoreline, which becomes younger oceanward since early Tertiary. Ridges with volcanic products are present between these belts. We interpret these basins and associated ridges as relict backarc basins and arcs of the Ryukyu subduction system. ii) Subsidence curves across west Taiwan Basins show that rifting ceased 17–18 Ma. iii) A new shear wave velocity model suggests that the Ryukyu slab extended in the past southwest of Taiwan, beneath the northern China Sea margin. iv) A deep seismic line shot across the northeastern South China Sea margin also suggests that this margin was active in the past. We conclude that about 15-20 Ma, the southwestern extremity of the Ryukyu subduction zone jumped from 118°E (southwest of the Tainan Basin) to 126°E (where the present-day trend of the Ryukyu subduction zone changes direction). Since that time, the southwestern extremity of the Ryukyu subduction zone continuously moved westwards to its present-day location at 122°E. Since the beginning of formation of proto-Taiwan during late Miocene (9 Ma), the subducting PH Sea plate moved continuously through time in a N307° direction at 5.6 cm/yr with respect to EU, tearing the EU plate.

1. INTRODUCTION

The collision of the Luzon volcanic arc with the South China Sea (SCS) margin is probably one of the best examples of modern arc-continent collision associated with the Taiwan uplift. The geology of the island as well as the mechanisms of deformation are quite well understood but in general, most of the geodynamic context in which this uplift occurs is largely unknown or at least oversimplified. For example, the continental margin before collision is widely considered as a typical passive margin and shelf basins are considered to be tensional basins formed during the rifting of the adjacent continental margin. In this paper, we define the geological and kinematic contexts of the formation of Taiwan not only during the collision of the Luzon arc with the Eurasian plate, which results in the formation of the proto-and present-day Taiwan, but also before the Taiwan uplift. This work is partly based on a series of papers published during recent years [*Hsu*

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and Sibuet, 1995; *Sibuet et al.*, 1998; *Sibuet and Hsu*, 1997; *Sibuet and Hsu*, 2004; *Sibuet et al.*, 2002; *Sibuet et al.*, 1995]. We demonstrate that the Tertiary shelf basins, located between Japan and southwest Taiwan, were formed as backarc basins linked to the Ryukyu subduction zone. Backarc basins older than 15-20 Ma are linked to the subduction of the Philippine and Taiwan Seas, which are bounded to the south by the Luzon-Ryukyu transform plate boundary (LRTPB) now located in the northeastern part of the SCS. South of the LRTPB, the SCS was opening between 37 and 15 Ma. The Pearl River shelf basins were formed by extension during the rifting of the SCS continental margins [*Clift and Lin*, 2001]. A major issue in understanding the tectonics of Taiwan is the local plate kinematic configuration, which is partly hidden by extensive deformation. We here establish that the kinematic pattern controls the formation of proto- and present-day Taiwan, in particular where and how the newly formed Luzon Arc started to collide with the Eurasian margin. We also demonstrate that the upper portion of the Luzon Arc is accreted along the Eurasian margin while the lower crust is subducted with the Philippine (PH) Sea plate beneath Eurasia (EU). We define the location of the western boundary of the PH Sea plate (Point A) that bounds the western termination of the Ryukyu Benioff zone and triggers the occurrence and development of a tear fault within the EU plate. To the south, the PH Sea plate boundary extends along the Longitudinal Valley of Taiwan and then runs east of the Manila accretionary prism. The PH Sea plate has been moving westward through time with respect to the EU plate in the northwestern direction (*e.g.* Seno and Maruyama [1984]). We attempt to explain the process of Taiwan uplift within a simple plate tectonic context, which includes a comprehensive history of the westward migration of the collision. For that purpose, we clarify a series of points concerning the geometry of plate boundaries, as well as the nature of the deformation and the associated stress patterns.

2. PRE-COLLISIONAL GEOLOGY: RYUKYU SUBDUCTION ZONE AND ASSOCIATED BACKARC BASINS

The Tertiary basins of the Chinese continental shelf around Taiwan are roughly oriented NE-SW [*Letouzey et al.*, 1988], *i.e.* parallel to the East China shoreline and margin (Plate 1). The tectonic history of these basins shows that rifting generally occurred between early Paleocene and middle Miocene (65-15 Ma), except for the active Okinawa Trough (OT), which is mostly in the rifting stage [*Sibuet et al.*, 1987]. From the geological evolution of a continental shelf basin, it is impossible to decipher if its tensional environment is related to the rifting of the nearby continental margin or if it is a backarc basin linked to the presence of a subduction zone. This is true for the continental shelf basins of the East and South China Seas, which are traditionally considered as having formed during the rifting of the adjacent continental margins. For example, Teng and Lin [2004, in press] consider that the Ryukyu subduction zone, which extended at least from Japan to Indochina, ceased in late Cretaceous and resumed during early Miocene in order to explain the formation of the OT backarc basin. In this scenario the Ryukyu subduction zone was not active during most of the Tertiary, which is in opposition with all published kinematic works. For example, Teng and Lin [2004, in press] also suggest that the Paleogene rifting of all the other continental shelf basins of the East and South China Seas continental shelf occurred during the simultaneous Paleogene extensive crustal attenuation of the adjacent continental margin (rifting episode for us). However, such a hypothesis does not take into account any of the recent data acquired in the deep ocean. Here we describe these shelf basins, distinguish their rifting from post-rifting periods and identify what basins are backarc basins or rifted shelf basins on the basis of the geodynamic context established from the interpretation of marine data.

2.1. Taiwan and Surrounding Basins

2.1.1. Pearl River Basins. The Pearl River Basins are a continuous succession of connected basins (Zhujiangkou Basins), which extend from the island of Hainan to west of the Tainan Basin (west of the LRTPB, Plate 1a). Well data indicate that basement rocks comprise Paleozoic metasedimentary rocks and Jurassic and upper Cretaceous formations consisting mainly of non-marine sequences, granitic intrusions and volcanic rocks [*Sun*, 1981; *Sun*, 1985]. The basins are mostly Paleogene NE-SW trending half-grabens or grabens bounded by normal faults with some tilted fault blocks. Sediments are composed of non-marine sediments, mostly shallow lake mudstones overlain by synrift Eocene deep lake mudstones with some fluvial sandstones interbedded with shales [*Ru*, 1988]. Deposition of these sediments is controlled by the throw of normal faults, mostly striking NE-SW [*Ru and Pigott*, 1986]. A regional late Eocene-early Oligocene unconformity separates synrift from flat and undeformed postrift sediments, which display the late Oligocene-early Miocene thin layers of limestone indicative of the first marine transgressions [*Zhao*, 1988]. Continuous marine deposition was interrupted in the middle Miocene by a significant tensional event characterized by E-W trending normal faults, a major unconformity and widespread basaltic extrusions [*Yu*, 1994]. The principal rifting event seems to have occurred mostly during Eocene and early Oligocene (Plate 1b) perhaps during the rifting of the northeastern SCS margin, before the onset of seafloor spreading (*e.g.* [*Briais et al.*, 1993; *Hayes et al.*, 1995]) (Plate 1a).

2.1.2. West Taiwan Basins. The Tainan Basin is located at the northeastern end of the SCS continental margin and extends onshore into southwestern Taiwan (Plate 1a). Numerous seismic and well data obtained by the Chinese Petroleum Company show that, although the basin is ENE-WSW trending, roughly parallel to the continental slope, the main boundary of the northern side of the basin is composed of EW segments which either lie together in an en-échelon pattern or are connected by orthogonal transfer faults with strike-slip motion [*Lee et al.*, 1993; *Tzeng et al.*, 1996; *Yang et al.*, 1991]. These transfer faults do not seem to be related to pre-existing features. In the basin itself, normal faults are also discontinuous and EW-oriented, though the direction of opening of the basin is NNW-SSE [*Yang et al.*, 1991]. This structural pattern is similar to the one of the OT backarc basin where en-échelon normal faults do not trend in the direction of the trough itself [*Sibuet et al.*, 1995]. The reconstruction of the different evolutionary stages and seismic sequences developed in the Tainan Basin show that, although the basin was relatively symmetrical during its formation, the bounding horst limiting the basin to the south and whose southern flank corresponds to the continental slope, has continuously subsided since the early Pliocene [*Lee et al.*, 1993]. This explains the present-day asymmetrical shape of the basin and the relative depth of the horst (3 km, Figure 3), which limits the Tainan Basin on the shelf edge side. Rifting occurred from early Oligocene to early middle Miocene (Plate 1b). Fault-controlled subsidence is the dominant tectonic mechanism for the deposition of sedimentary sequences [*Lee et al.*, 1993]. The rifting phase was followed by a thermal subsidence phase followed by a rapid flexural subsidence phase linked to the uplift of Taiwan since the Pliocene (6.5 Ma) [*Lin et al.*, 2003].

The Taishi Basin located northwest of Taiwan (Plate 1a) seems to differ from the NE-SW trend of most of the other basins. A series of horsts and grabens are trending N080° [*Huang et al.*, 1993]. Two rifting phases seem to occur in the Taishi Basin (Figure 1). The first one started during late Paleocene (58 Ma) and seems to terminate during middle Eocene (37 Ma). The second one started in early Oligocene (30 Ma) and seems to terminate during late early Miocene (17–18 Ma) (Plate 1b and Figure 1). Since this period, several compressional or transpressional regimes have caused the tectonic inversion of the basin.

The Nanjihtao Basin is separated from the Taishi Basin by the Nanjihtao Ridge (Plate 1a), which is a NE-SW trending half-graben limited in the southeast by a major boundary fault with more than 3-km normal displacement. Stratigraphic correlations indicate that rifting occurred between the early Paleocene and late Eocene (Plate 1b), after which time a major unconformity characterized by erosional truncation represents the late Eocene to late Oligocene [*Chow et al.*, 1991].

The Penghu Basin, located between the Nanjihtao and Pearl River Basins (Plate 1a), is also NE-SW trending, with bounding faults approximately oriented in the same direction [*Letouzey et al.*, 1988]. Though poorly described in the published literature, the main rifting period in the Penghu Basin seems to have occurred from Paleocene to middle Eocene times and possibly during late Eocene [*Hsiao et al.*, 1991] (Plate 1b).

2.1.3. North and northeast Taiwan Basins. The Tungyintao and Tahchentao Basins, located close to Mainland China, are half-grabens (Plate 1a). From seismic and well data, the basement is known to be late Cretaceous, with rifting mostly dated as Paleocene (3.5 km of synrift sediments) (Plate 1b), after which subsidence slowed considerably during the Eocene and early Oligocene [*Huang et al.*, 1992].

The north and south Pengchiahsu Basins, located southeast of the basins already mentioned, are characterized by NE-SW trending normal faults (Plate 1a). Prerift sequences are Paleocene or slightly older, and the synrift sequences are Eocene or slightly younger [*Huang et al.*, 1992] (Plate 1b). The Eocene synrift sequence is more than 3-km thick, with a lower section showing chaotic reflections suggestive of deposition in a semi-enclosed environment overlain by an upper section deposited in a more marine environment [*Huang et al.*, 1992].

The NE-SW trending Taipei Basin is located between the Pengchiahsu Basins and the Taiwan-Sinzi Ridge (Plate 1a). The prerift sequences are dated as Eocene, and the synrift sequences as Oligocene [*Huang et al.*, 1992] (Plate 1b), although middle Miocene and late Miocene to Pliocene/Pleistocene tensional phases, post-dating the main rifting phase, are also observed [*Huang et al.*, 1992]. The Oligocene synrift sequence is more than 4-km thick and composed of a prograding lower Oligocene deltaic sequence overlain by strongly stratified sandstone and shale alternations, and an extensive late Oligocene deep marine mudstone sequence [*Huang et al.*, 1992].

The OT backarc basin extends from Taiwan to Kyushu Island and is separated from the Taipei Basin by the Taiwan-Sinzi Ridge (Plate 1a). Rifting of the OT started during late middle Miocene times [*Letouzey and Kimura*, 1986; *Sun*, 1981] and is still active today [*Sibuet et al.*, 1987] (Plate 1b). However, the beginning of extension could be much more recent (6 Ma) as suggested by Kimura [1996]. Three main phases of opening have been identified, and poles and angles of rotation have been determined [*Sibuet et al.*, 1998; *Sibuet et al.*, 1995]. The first phase of rifting is the most important one with 50 to 75 km of extension from the southern to the northern OT [*Sibuet et al.*, 1995]. This extension started in the middle Miocene (12 Ma) or late Miocene (6 Ma). The two most recent phases of rifting are dated as late Pliocene-Pleistocene and late Pleistocene to Recent, with only a total of a few tens of kilometers of extension [*Sibuet et al.*, 1998].

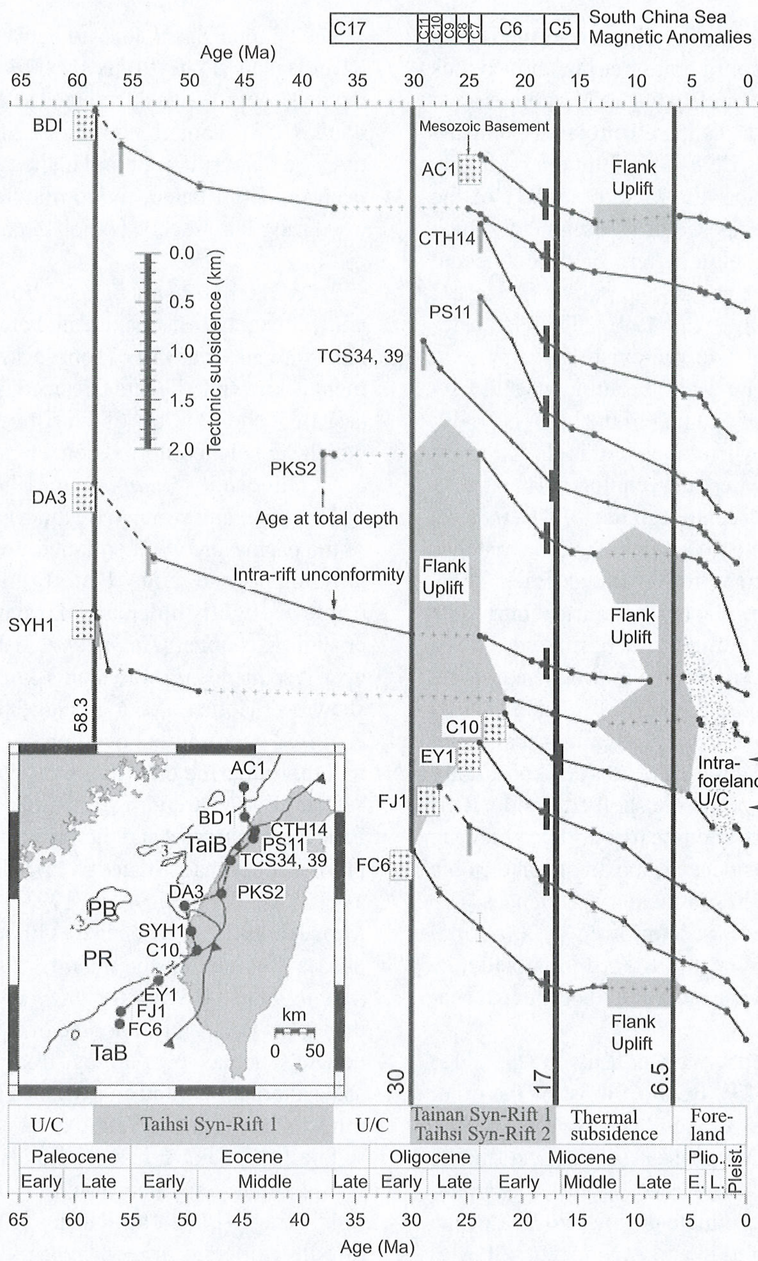


Figure 1. Tectonic subsidence pattern (lines) and tectonic development of the west Taiwan basins shown at the bottom of the figure (modified from Lin et al. [2003]). Also shown at the top of the figure are the seafloor magnetic lineations of the South China Sea [*Briais et al.*, 1993; *Hsu*, 2004, in press] according to the age model of Cande and Kent [1995]. The vertical gray bars indicate paleobathymetric uncertainties. The vertical black bars at 17–18 Ma indicate the end of the rifting episode in the Tainan Basin, as well as the end of a second rifting episode in the Taishi Basin. 17–18 Ma also corresponds to the cessation of the Ryukyu subduction zone in northeastern South China Sea (see text). Boxes with a cross pattern represent the Mesozoic basement. The insert figure shows the location of wells. Crosses, non-depositional hiatuses; U/C, unconformity. PB, Penghu Basin; PR, Penghu Ridge; TaB, Tainan Basin; TaiB, Taishi Basin.

2.1.4. Taiwan onshore Basins (Hsüehshan Trough). Several authors have already suggested the existence of a large basin in the area of Taiwan before its uplift [*Ho*, 1984; *Teng*, 1992]. Located on the outer continental shelf before the collision, it was named Hsüehshan Trough (Plate 1a) by Teng et al. [1991]. This basin corresponds to the present-day Western Foothills and the Slate Belt (Hsüehshan and Backbone Ranges). In the Hsüehshan Range, Eocene and Oligocene shallow marine sediments are estimated to be up to 15-km thick and early to middle Miocene sediments are about 1-km thick. Syn-sedimentary growth faults trend nearly in an E-W direction [*Lu et al.*, 1991], as in the offshore Tainan and Taishi Basins. However, no such tensional features were found in the Backbone Range. There, Oligocene sediments seem to be absent, suggesting to Teng [1992] that the Mesozoic basement had always been closer to the surface there. Before collision, the Hsüehshan Trough, which was floored with a Mesozoic basement, was at least 250-km long, as shown by the north-south extension of the Hsüehshan and Backbone Ranges, and about 200-km wide, taking into account compressional features in the Western Foothills and the Hsüehshan and Backbone Ranges [*Suppe*, 1980; *Teng et al.*, 1991]. The main rifting episode occurred during middle Eocene to early Miocene [*Lu and Hsü*, 1992; *Teng and Lin*, 2004, in press] (Plate 1a).

We propose that the Hsüehshan Trough corresponds to two parallel basins: a northwestern one, which was very deep, filled with more than 15-km of sediment and corresponding to the Hsüehshan Range and the eastern part of the Western Foothills; and a southeastern one, a few kilometers deep, corresponding to the western Backbone Range. These basins were probably separated by a linear topographic high, which forced Oligocene turbidites derived from the Chinese platform first to fill the western depression. Sediments in the southeastern depression either bypassed the western basin after the late Oligocene and/or were derived from the adjacent Ryukyu Arc. Igneous rocks such as pillow lavas, diorite and gabbros are common in the Hsüehshan Trough. Subordinate andesites and dolerites are present within the Backbone Range sedimentary rocks, but are not found in the Hsüehshan Range [*Chen*, 1991]. The Lishan Fault, which limits the Hsüehshan and Backbone Ranges, was originally a normal, listric fault located between the northwestern and southeastern depressions and bounded the linear topographic feature to the west. It was reactivated as a reverse fault during later arc-continent collision.

2.2. Structure, Nature and Timing of the Rifting of Shelf Basins

2.2.1. Structure and timing of rifting of shelf basins. Plate 1a shows that basins of the continental shelf around Taiwan are NE-SW oriented and are parallel to the mainland Chinese shoreline. However, the trend of normal faults is significantly different from the mean direction of the basins and is generally EW oriented. This is true for the Tainan and Taishi Basins, the OT, and also for the Pearl River Basins, where two families of normal faults are present. One set of faults was active during almost the entire rifting phase and runs parallel to the northern border of the basins, while the other is E-W-oriented and dates from the end of the rifting phase [*Ru*, 1988; *Ru and Pigott*, 1986; *Sun*, 1981; *Sun*, 1985].

Plate 1b shows the timing of rifting in continental basins around Taiwan. Some uncertainty exists about the exact beginning and end of their main rifting phase, mostly because the shutdown tensional rifting activity could be interpreted as part of the postrift thermal subsidence phase, or subsequent less intense extensional tectonic phases that could complicate the interpretation. However, we identify four belts of basins corresponding to four main rifting phases: (1) the oldest belt (Belt 1) formed during the late Cretaceous-Paleogene and consists of the Penghu, Nanjihtao, Tungyintao and Tahchentao Basins; (2) Belt 2 formed during the Eocene. Extension decreased and ceased in the older basins but started further southeast in the south and north Pengchiahsu and Taishi Basins (Plate 1a); (3) Belt 3 formed during Oligocene-early middle Miocene consists of the Tainan and Taishi Basins, the Hsüehshan Trough and the Taipei Basin, *i.e.* in basins located southeast of those that were previously active; (4) rifting ceased in all basins but started in the OT (Belt 4) during the late middle Miocene (12 Ma) or later during late Miocene (6 Ma).

To summarize, the age of rifting episodes and the orientation of all continental shelf basins of the East and South China Seas located northeast of the Pearl River Basins, including the Hsüehshan Trough, are consistent with the idea that these shelf basins were formed in the same backarc basin tectonic setting. In opposition with Teng and Lin [2004, in press], we suggest that the Ryukyu subduction process occurring northeast of the LRTPB was continuous from the early Tertiary to middle Miocene during the formation of the three oldest belts of shelf basins.

2.2.2. Tectonic subsidence pattern of the west Taiwan Basins. Figure 1, adapted from Lin et al. [2003], shows the tectonic subsidence pattern of a series of wells drilled along a NE-SW profile in the Tainan and Taishi Basins. Rifting occurred during Paleogene/early Eocene (58-37 Ma) in the Taishi Basin and during Oligocene/early Miocene (30-17/18 Ma) in the Tainan Basin. A second Oligocene/early Miocene phase of rifting occurred in the Taishi Basin. 17–18 Ma, a major change in slope occurred in the subsidence curves of all wells (Figure 1) that we interpret as the end of active extension and the beginning of thermal subsidence in the Tainan and Taishi Basins. Since 6.5 Ma, the concave-down subsidence pattern is interpreted as the result of flexure driven by Taiwan mountain loading [*Lin and Watts*, 2002]. As described in previous sections, the first rifting episode in the Taishi Basin started in late Cretaceous/Paleocene and its duration is poorly defined. In the Tainan Basin, the duration of the rifting episode is well defined, lasting from early Oligocene to late early Miocene (30-

17/18 M.y.). Interestingly, the second rifting phase in the Taishi Basin is synchronous with the first rifting phase of the Tainan Basin. Lin et al. [2003] link the early Oligocene uplift followed by the rapid Oligocene/early Miocene subsidence (that we identify as a rifting phase in the Tainan Basin) to the appearance and disappearance of induced mantle convection under the newly created SCS. In fact, the western Taiwan basins are facing an old portion of oceanic domain (proto-SCS, early Cretaceous?) [*Sibuet et al.*, 2002] and the Pearl River Basins are facing the SCS oceanic domain where magnetic lineations C15 to C17 (37 Ma) (Plate 2 and Figure 2) have been identified [*Hsu*, 2004, in press]. Rifting of the passive continental margins adjacent to the west Taiwan Basins and Pearl River Basins occurred during the early Cretaceous and Paleogene respectively, that is before the rifting of adjacent shelf basins. Consequently, it is impossible to link the rifting

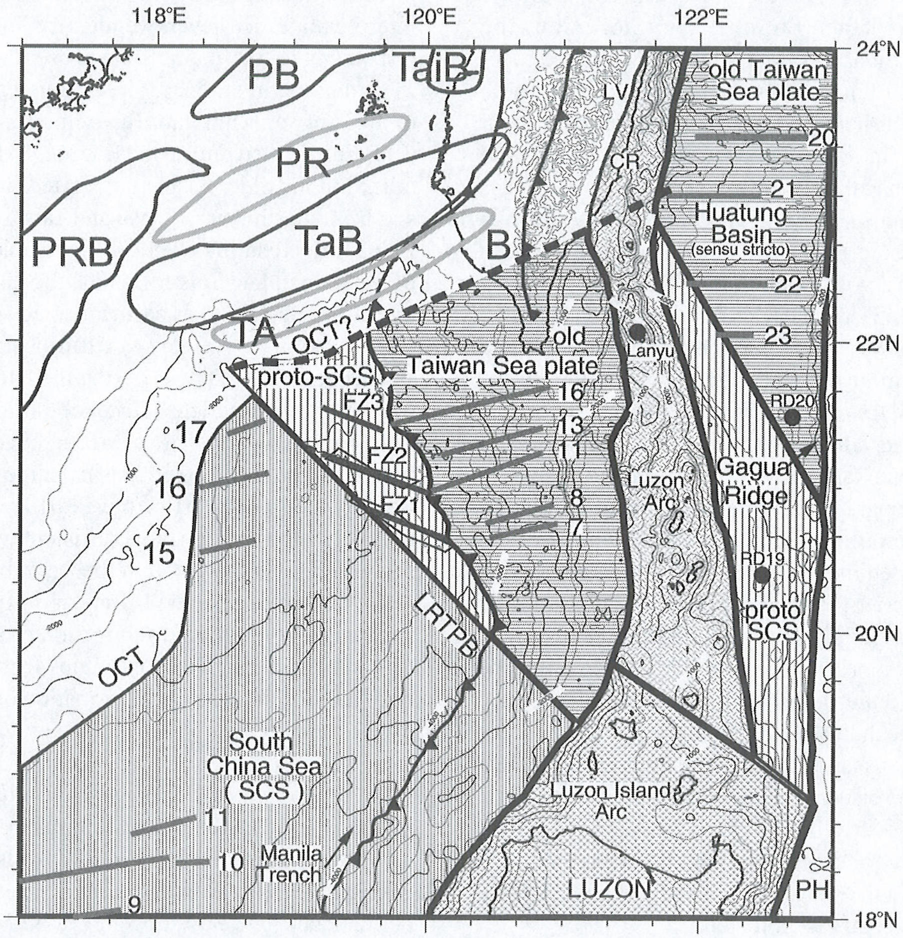


Figure 2. Geodynamic interpretation of the South China Sea (SCS), Huatung Basin and surrounding areas in terms of pieces of ocean belonging to different plates (adapted from Sibuet et al. [2002]). The SCS ocean-continent transition (OCT) has been approximately drawn from bathymetric, gravity and magnetic data. Magnetic lineations, fracture zones (FZ1, FZ2 and FZ3) and plate boundaries are identified in the northeastern SCS (proto-SCS), beneath the Manila accretionary prism (old Taiwan Sea plate) and in the Huatung Basin *sensu stricto* as well as in the SCS [*Briais et al.*, 1993; *Hsu*, 2004, in press]. The southern part of the Huatung Basin might belong to the proto-SCS. The three black circles correspond to locations where early Cretaceous rocks have been discovered [*Deschamps et al.*, 2000]. Point B is the surface projection of the Eurasian ocean-continent transition (OCT), which subducts eastward beneath the Philippine Sea plate; CR, Coastal Range; LRTPB, Luzon-Ryukyu transform plate boundary; LV, Longitudinal Valley; PB, Penghu Basin; PR, Penghu Ridge; PRB, Pearl River Basins; TA, Tainan Arc; TaB, Tainan Basin; TaiB, Taishi Basin. Note that the west Taiwan basins are facing an old portion of oceanic domain (proto-SCS, early Cretaceous) and that the Pearl River Basins are facing the SCS oceanic domain where magnetic lineations C15 to C17 (37 Ma) have been identified [*Hsu*, 2004, in press]. Rifting of the passive continental margins adjacent to the west Taiwan Basins and Pearl River Basins occurred during early Cretaceous and Paleogene times respectively.

of the shelf basins to the rifting of the adjacent continental margins. We propose here that the rifting phases in the west Taiwan Basins are linked to backarc extension relative to the Ryukyu subduction zone, which extended at that time southwestward to the LRTPB (Figure 2) and that the early Oligocene uplift is correlated to the collision of ridges or asperities belonging to the subducting oceanic domain with the EU margin. We suggest that the Ryukyu subduction zone and the associated backarc basins ceased to be active 17/18 Ma.

2.2.3. *Shear wave tomography.* We present in Plate 3 tomographic results for the studied region from the global surface wave inversion of Debayle et al. [2004]. This global shear wave velocity model is constrained by the waveform inversion of 100,466 fundamental and higher mode Rayleigh wave seismograms analyzed in the 50-200 s period range. The waveform inversion is performed using the approach described in Cara and Lévêque [1987] and Debayle [1999]. Debayle et al. [2004] extended the data set compiled by Debayle and Sambridge [2004] to a global data set with a majority of seismograms corresponding to epicenter-station paths shorter than 6000 km. The use of these relatively short paths in a global surface wave inversion allows to minimize some spurious effects not taken into account in the theory and to improve the horizontal resolution to a few hundred of kilometers in the upper mantle [*Debayle and Sambridge*, 2004]. This is especially true in well-sampled regions of the model. The area surrounding southeast China and the northern Philippine Sea corresponds to one of the highest ray density region in the global model of Debayle et al. [2004]. Over almost the entire region covered by the present study the number of rays per 400 by 400 km boxes exceeds 2000 (Plate 3) and the azimuthal coverage is excellent. Possible decrease in horizontal resolution may occur in the dark blue regions where the number of rays is smaller than 1200. However, excepted in the dark blue regions on the border of the maps, it is likely that the horizontal resolution is rather homogeneous and the high velocity signature of the Ryukyu slab on the maps at 150 and 200 km depths suggests that heterogeneities with horizontal wavelengths of few hundred of kilometers are detected by the tomographic inversion. Teleseismic body wave tomography allows us to sample the upper mantle with vertical rays beneath the stations, locally providing a better horizontal resolution in the upper mantle but failing to maintain this resolution in oceanic regions where the number of stations decreases. Spike tests per Lévêque formed in the body wave model of Bijwaard et al. [1998] suggest that resolution strongly decreases southwest of Taiwan [*Lallemand et al.*, 2001] although this should not be the case in the surface wave model.

The map at 50 km depth emphasizes the contrast between the oceanic lithosphere of the PH Sea and of the younger SCS, and the EU continental lithosphere. On the map at 150 km depth, the Ryukyu slab is imaged beneath the EU and PH Sea lithospheres. It extends continuously from Japan to west of Taiwan, in a region where the Ryukyu slab is now inactive. On the map at 200 km depth, the slab extends even more westwards. Cross-sections AA′ to DD′ (Plate 3) aligned on the PH Sea and SCS margins clearly show that the high velocity 'slabs' in blue are also aligned, even if this correspondence is questionable for section DD'. In opposition to the body wave tomography [*Lallemand et al.*, 2001], which provides poor resolution in the upper mantle southwest of Taiwan, the shear wave tomography enlightens for the first time the extent of the Ryukyu extinct slab beneath the northeastern SCS margin, even if the resolution of data cannot allow us to properly locate its western extremity.

2.2.4. *Subduction of the northeastern South China Sea beneath Eurasia.* Seismic profiles EW9509 45 (located in Plate 1a) lies perpendicular to the northern margin of the northeastern SCS. A large igneous body associated with a magnetic anomaly at 21°15′N; 119°45′E (Figure 3) was intruded (during Miocene?) through undeformed sediments overlying the oceanic crust along narrow conduits not imaged on seismic data. Northwest of this igneous body (CDP 15,000 on profile 45; Figure 3), the layering of the undeformed sedimentary layer completely disappears (CDP 11,500) and a deformed sedimentary layer is seen to thicken toward the margin. The layer of deformed sediments is interpreted as an old accretionary prism linked to the subduction of the proto-SCS beneath EU [*Sibuet et al.*, 2002]. Above the accretionary prism, about 4 s two-way travel time (TWT) of undeformed turbiditic sediments have been deposited. NW of CDP 10,000 (Figure 3), the shape of the seismic reflections on top of oceanic crust and the velocity structure defined there from an OBS refraction survey [*Chen and Jaw*, 1996] indicate that the underlying crust is oceanic and dips gently northwestward, suggestive of former subduction. Beneath the continental shelf, the Tainan Basin, whose main rifting period occurs from 30 to 17–18 Ma (Figure 1), shows a complex tectonic evolution. We interpret the Tainan Basin as a backarc basin linked to the subduction of the northeastern SCS beneath EU [*Sibuet and Hsu*, 1997]. Beneath the upper part of the margin, a deep body devoid of seismic reflections may represent the former arc and forearc crust, now buried beneath sediments eroded from the Taiwan mountains. Because most of the seismic profiles previously shot by the Chinese Petroleum Company did not extend over the continental slope [*Tzeng et al.*, 1996], the relationship between the Tainan backarc basin and the arc/forearc body, the sedimentary prism and the subducting oceanic crust was not recognized. Thus, we interpret profile 45 as a seismic section

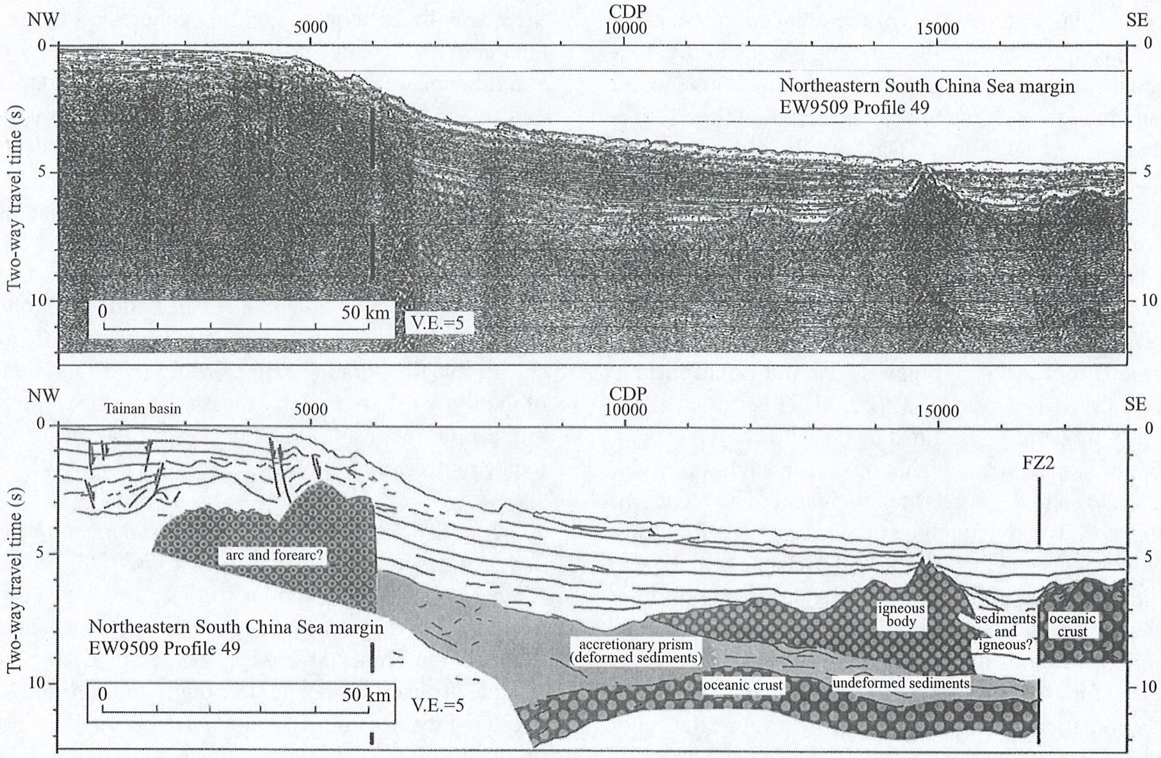


Figure 3. Migrated seismic profile EW9509 45 [*Sibuet et al.*, 2002] located in Plate 1a (as P45) and interpretation. This profile is perpendicular to the northern margin of the northeastern South China Sea (proto-SCS). The igneous body is associated with a large magnetic anomaly and was probably intruded through undeformed sediments overlying the oceanic crust. Northwest of the igneous body, the layering of this undeformed sedimentary layer completely disappears, giving way to a strongly deformed sedimentary layer, interpreted as an old accretionary wedge, that thickens towards the margin. The underlying oceanic crust dips northeastward, even if the sedimentary load is removed. Beneath the outer continental shelf and upper slope, the basement body is interpreted as arc and forearc features.

across a portion of an extinct northwest-dipping subduction zone located beneath the Tainan Basin, and which probably extended beneath the Hsüehshan Trough [*Sibuet and Hsu*, 1997]. Both the Hsüehshan Trough and the Tainan Basin ceased active extension during early middle Miocene. Other seismic profiles (EW9509 34 and 35) located east of the LRTPB are parallel and perpendicular to profile EW9509 45 and show that the sedimentary layers corresponding to the deformed sediments (accretionary prism) of profile 45 are also folded [*Sibuet et al.*, 2002]. West of the LRTPB, deep seismic profiles shot perpendicular to the margin show that the northeastern SCS is a rifted margin [*Hayes et al.*, 1995; *Nissen et al.*, 1995]. The LRTPB was consequently a plate boundary between a passive and an active margin. In the deep ocean, the LRTPB separates two oceanic domains of different ages. Recent magnetic surveys carried out west of the LRTPB show that the oldest identified magnetic lineations in the northeastern SCS is C17 (37 Ma) [*Hsu*, 2004, in press] (Plate 2) instead of C11 [*Briais et al.*, 1993]. After correction for the sedimentary load, the oceanic basement located east of the LRTPB is 0.8 km deeper than the rest of the northeastern SCS. This piece of oceanic crust is consequently inferred to be older than C17, and could possibly belong to the early Cretaceous proto-SCS [*Sibuet et al.*, 2002].

In summary, the interpretation of deep seismic profile 45 suggests the existence of a former subduction zone, inferred from the apparent frozen subduction of oceanic crust, the existence of a thick accretionary prism located in the lateral continuity of undeformed oceanic sediments, an arc and forearc feature, and a backarc basin (Tainan Basin). However, the interpretation of seismic data alone is not sufficient to prove the existence of a former subduction zone, as dipping oceanic crusts in the direction of passive continental margins are not so uncommon in the world (*e.g.* beneath the eastern US margin). Between the LRTPB and the deformation front, the amount of subduction beneath EU is unknown but kinematic considerations suggest that this amount of subduction cannot exceed 150 km [*Sibuet et al.*, 2002].

2.2.5. Ryukyu subduction zone and backarc basins formation. As the four belts of continental shelf basins become younger in direction of the ocean, such a chronology is almost impossible to explain as being the result of rifting of the adjacent passive continental margins. We consequently suggest that all these basins belong to four belts sequentially formed as backarc basins associated with the Ryukyu subduction zone. In this hypothesis, from late Cretaceous/early Paleocene to early middle Miocene, the Ryukyu subduction zone extended from Japan to the western end of the Tainan Basin where the continental margin changes direction (118°E longitude, Plate 1a). During the early middle Miocene, the portion of subduction zone corresponding to the Tainan Basin and the Hsüehshan Trough became inactive and the southwestern limit of the subduction zone located at 118°E longitude jumped north of Taiwan to 126°E longitude. Such an interpretation has been already proposed at a reduced regional scale for the northern Taiwan Basins by Huang et al. [1992] and Chi [1996].

2.2.6. Relict arcs on the East China continental margin. The Taiwan-Sinzi Ridge (Plate 1a), located between the Taipei Basin and the OT, is associated with a large positive free-air anomaly (Plate 2). It is composed of paleo-highs made of deformed metamorphic rocks of Cretaceous age or older, intruded by Miocene volcanic rocks which crop out in the Sinkoku, Danjyo and northwest Kyushu islands. Strong positive magnetic anomalies [*Geological Survey of Japan and Committee for Co-ordination of Joint Prospecting for Mineral Resources in Asian Offshore Areas (GSJ and CCOP)*, 1994] over the Taiwan-Sinzi Ridge confirm the existence and extension of a continuous remnant volcanic arc. The Penghu Ridge is also underlain by strong positive gravity anomalies larger than 40 mGals and a belt of positive magnetic anomalies, which lies just north of the Tainan Basin. By comparison with the Taiwan-Sinzi Ridge, we interpret the Penghu Ridge as a remnant arc. A few radiometric ages have been obtained in the basaltic flows of Penghu Islands [*Chen et al.*, 1996; *Juang*, 1988]. Dates from these alkali-tholeiitic basalts span between 8.0 and 17.8 Ma.

Other ridges have been identified between basins and may also be interpreted as remnants of volcanic or non-volcanic island arcs (Plate 1a). For example, between the Tungyintao and Pengchiahsu Basins, arc volcanic rocks of probable Paleogene age exist on the Tungyintao Ridge and are associated with significant magnetic anomalies on the Magnetic Anomaly Map of East Asia [1994]. The inversion of magnetic data in the East China Sea also suggests the existence and location of several ridges in that area (Plate 1a) [*Hsu et al.*, 2001] as the Nanjihtao Ridge located between the Nanjihtao and Taishi Basins, though no relative positive gravity anomaly is associated with it.

We interpret these ridges, generally associated with positive or relative positive gravity anomalies and often-positive magnetic anomalies as remnant volcanic arcs linked to the Ryukyu subduction zone. As for the shelf basins, four belts of remnant arcs could be identified on the continental shelf, though less well defined that the four belts of basins: The oldest one (Belt A) formed mostly during the Paleogene consists of the Yandang Ridge, the Tungyintao Ridge and the Nanjihtao Ridge. Belt B formed later during the Paleogene and includes the Yushan Ridge and possibly the Penghu Ridge, although the oldest volcanism dated there is early Miocene. Belt C formed mostly during the Oligocene and early Miocene consists of the Tainan Arc, the Tananao Complex and the Taiwan-Sinzi Ridge. Belt D comprises the Ryukyu Arc, which has been active since the middle Miocene [*Sibuet et al.*, 1998; *Sibuet et al.*, 1995] (Plate 1a). Thus, we suggest that the following belts of backarc basins and volcanic (or non-volcanic) arcs were formed within the Ryukyu subduction system at different periods: Belts 1 and A during the Paleocene, Belts 2 and B during the Eocene, Belts 3 and C during the Oligocene and early Miocene, Belts 4 and D since the Miocene.

The structure and stratigraphy of the Pearl River Basins are related to the Paleogene rifting of the northern SCS margin [*Holloway*, 1982] followed by the SCS oceanic opening starting during late Eocene (C17, 37 Ma) [*Hsu*, 2004, in press] (Figure 1). In addition, southwest of the Tainan Basin, west of 118°E longitude, no major positive free-air anomalies exist on the continental shelf toward Indochina (Figure 1). Large negative anomalies correspond to the Pearl River Basins. Therefore, the very large contrast between free-air anomaly values northeast and southwest of the southwestern termination of the Tainan Basin (LRTPB) also suggests that, at least during the Paleogene, the former Ryukyu subduction zone did not extend southwest of the Tainan Basin and that the Pearl River Basins are rifted basins of the continental shelf linked to the rifting of the northern SCS margin. In addition, there is no arc volcanism on the northern SCS margin, southwest of the LRTPB.

2.2.7. Plate kinematics of the Taiwan region since about 15 Ma. We have established a set of five plate kinematic reconstructions of East Asia [*Sibuet et al.*, 2002] (Figure 4) with a major kinematic change occurring about 15 Ma or even earlier (17/18 Ma). We assume that:

- Spreading in the Taiwan Sea was 2 cm/yr from chron 23 to 20 (51 to 43 Ma) and 1 cm/yr from chron 20 (43 Ma) to 5b (15 Ma),
- Spreading ceased at the same time (about 15 Ma) in the Taiwan Sea, SCS, northeastern SCS, Sulu Sea, Japan Sea, and Shikoku and Parece Vela Basins.
- Since about 15 Ma, the mean PH/EU velocity has been 5.6 cm/yr in the N307° direction, matching magnetic lineations in

Kinematic reconstructions of East Asia since about 15 Ma

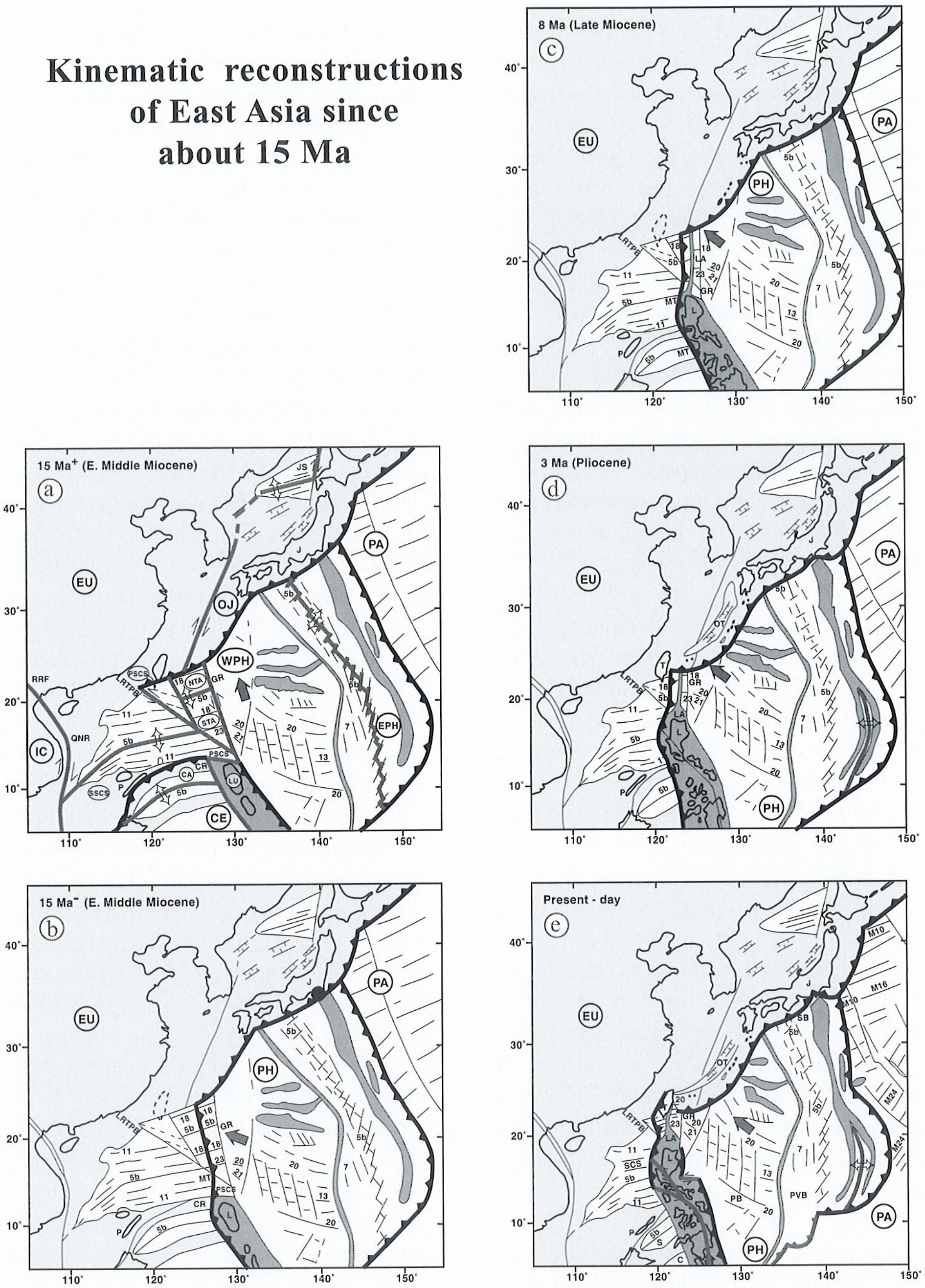


Figure 4. Plate kinematic reconstructions of East Asia (adapted from Sibuet et al. [2002]): a) Early middle Miocene (15 Ma+); b) Early middle Miocene (15 Ma-); c) Late Miocene (8 Ma); d) Pliocene (3 Ma); e) Present-day. Light gray, continental domain; white, oceanic domain; dark gray, intra-oceanic arcs; thick lines, active plate boundaries; large arrows, WPH/EU or PH/EU motions; large black dot, pole of rotation. Fracture zones and main magnetic lineations with their identifications are shown. Spreading ceased at the same time (about 15 Ma) in the Taiwan Sea, South China Sea, Sulu Sea, Japan Sea, and Shikoku and Parece Vela Basins. CR, Cagayan Ridge; GR, Gagua Ridge ; JS, Japan Sea; MT, Manila Trench; L, Luzon Island; LRTPB, Luzon-Ryukyu transform plate boundary; LA, Luzon Arc; OT, Okinawa Trough; PB, Philippine Sea Basin; QNR, Qui Nhon Ridge; RRF, Red River Fault; SB, Shikoku Basin; SCS, South China Sea; T, Taiwan; CA, Cagayan plate; CE, Celebes plate; EPH, east Philippine Sea plate; EU, Eurasia plate; IC, Indochina plate; LU, Luzon plate; NTA, north Taiwan Sea plate; OJ, Okinawa-Japan plate; PA, Pacific plate; PH, Philippine Sea plate; PSCS, proto-south China Sea plate; SSCS, southern south China Sea plate; STA, south Taiwan Sea plate; WPH, west Philippine Sea plate.

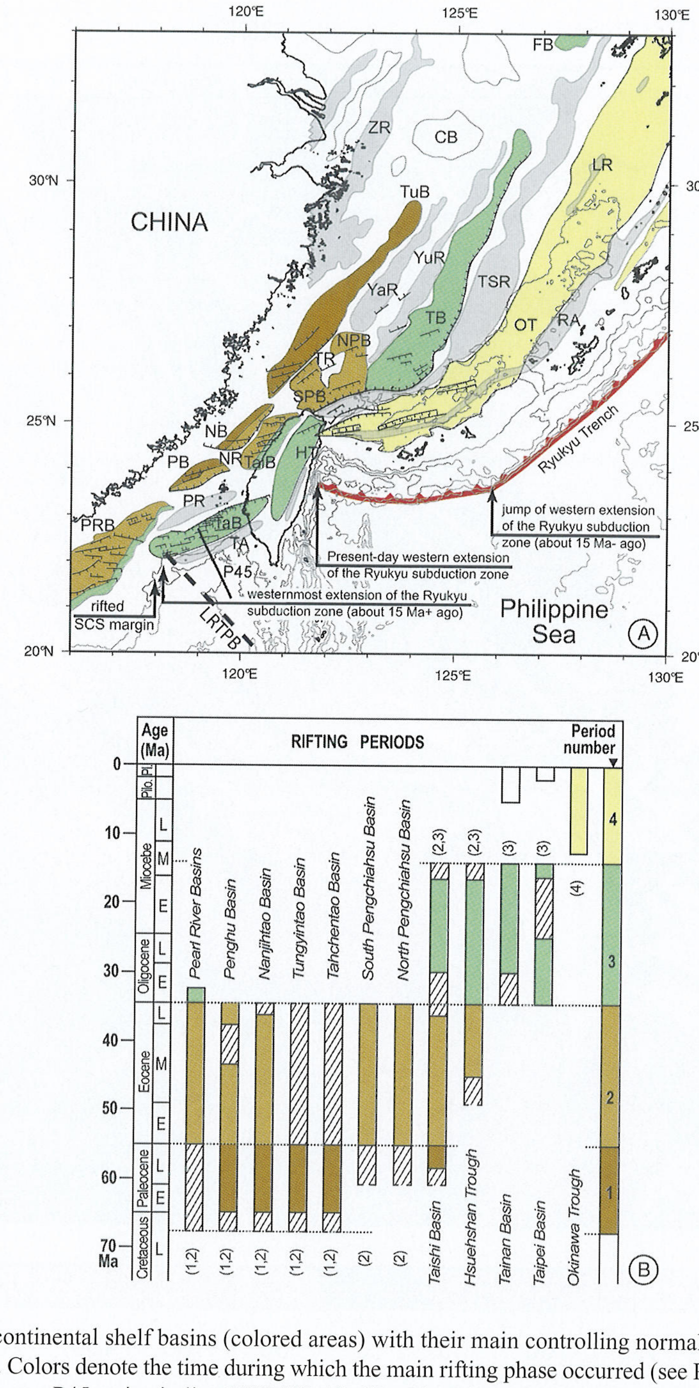


Plate 1. (A) Chinese continental shelf basins (colored areas) with their main controlling normal faults [*Hsu et al.*, 2001; *Sibuet and Hsu*, 1997]. Colors denote the time during which the main rifting phase occurred (see Plate 1b). Ridges between basins are marked in gray. P45, seismic line EW9509 45; CB, Changjiang Basin; FB, Fujiang Basin; HT, Hsüehshan Trough; LR, Longwan Ridge; NB, Nanjihtao Basin; NPB, North Pengchiahsu Basin; NR, Nanjihtao Ridge; OT, Okinawa Trough; PB, Penghu Basin; PR, Penghu Ridge; PRB, Pearl River Basins; RA, Ryukyu Arc; SPB, South Pengchiahsu Basin; TA, Tainan Arc; TB, Taipei Basin; TaiB, Taishi Basin; TaB, Tainan Basin; TR, Tungyintao Ridge; TSR, Taiwan-Sinzi Ridge; TuB, Tungyintao and Tahchentao Basins; TW, Taiwan; YaR, Yandang Ridge; YuR, Yushan Ridge; ZR, Zhemin Ridge. (B) Diagram showing the main rifting phases of each continental basin around Taiwan including the Hsüehshan Trough, the basin lying before the uplift of Taiwan (adapted from Sibuet and Hsu [1997]). Detailed explanations and references in the text. In color, main rifting phases; hachured areas, possible extension of main rifting phases; white areas, posterior tensional activity. Four main rifting periods have been identified. Periods during which rifting occurs appear for each basin.

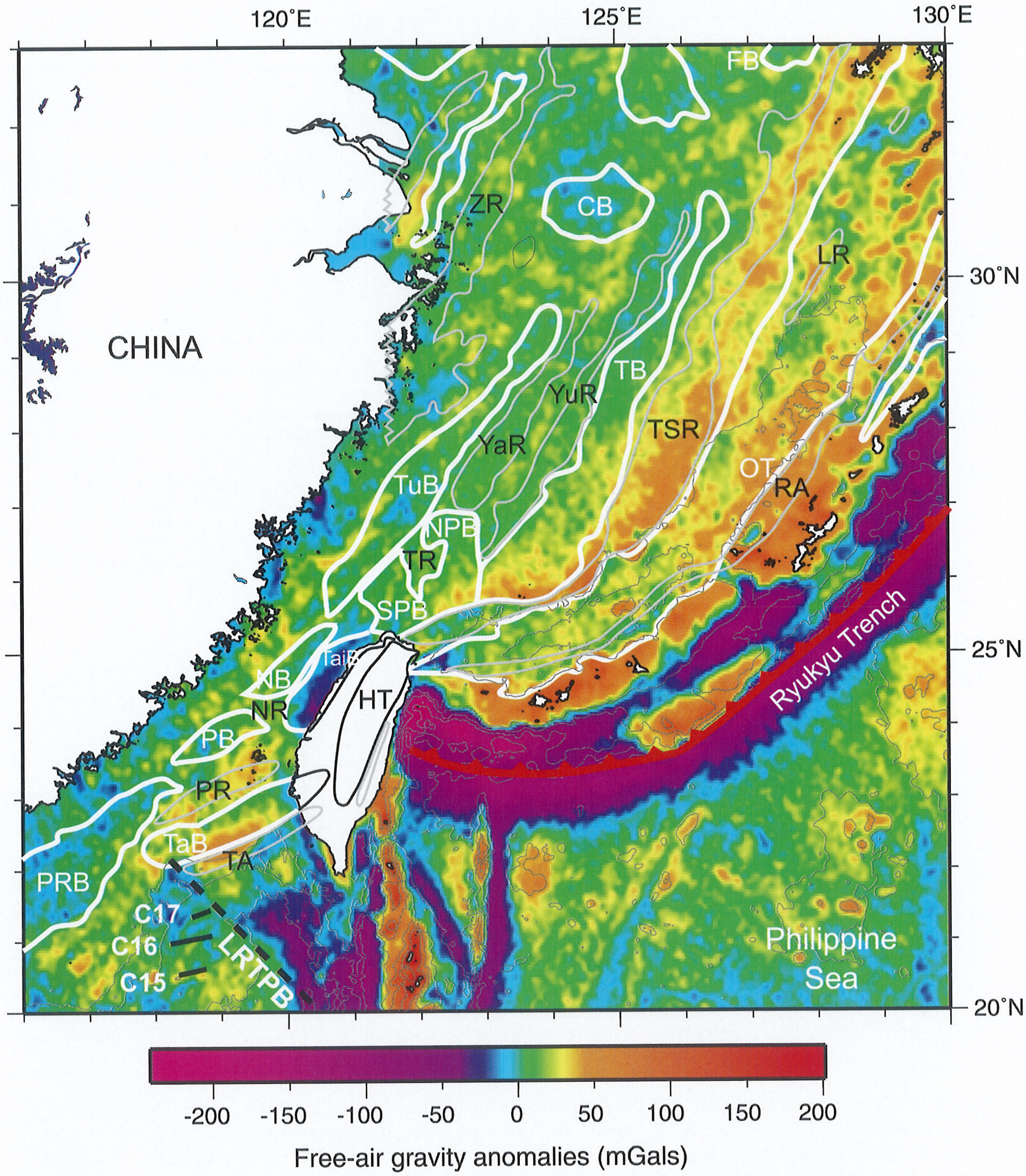


Plate 2. Satellite-derived marine free-air gravity anomalies around Taiwan [*Sandwell and Smith*, 1994]. Contours of continental shelf basins in white and ridges in light gray. Symbols as in Plate 1. LRTPB, Luzon-Ryukyu transform plate boundary; C15, C16 and C17 are the locations of the oldest magnetic lineations recently identified in the northeastern South China Sea [*Hsu*, 2004, in press]. The oceanic domain located northeast of the LRTPB is about 1 km deeper than southwest of the LRTPB, suggesting that the age of this portion of oceanic domain is much older that the rest of the South China Sea (early Cretaceous?). It is consequently excluded to link the formation of the adjacent west Taiwan basins (Tainan, Penghu, Nanjihtao and Taihsi Basins) to the rifting of the continental margin.

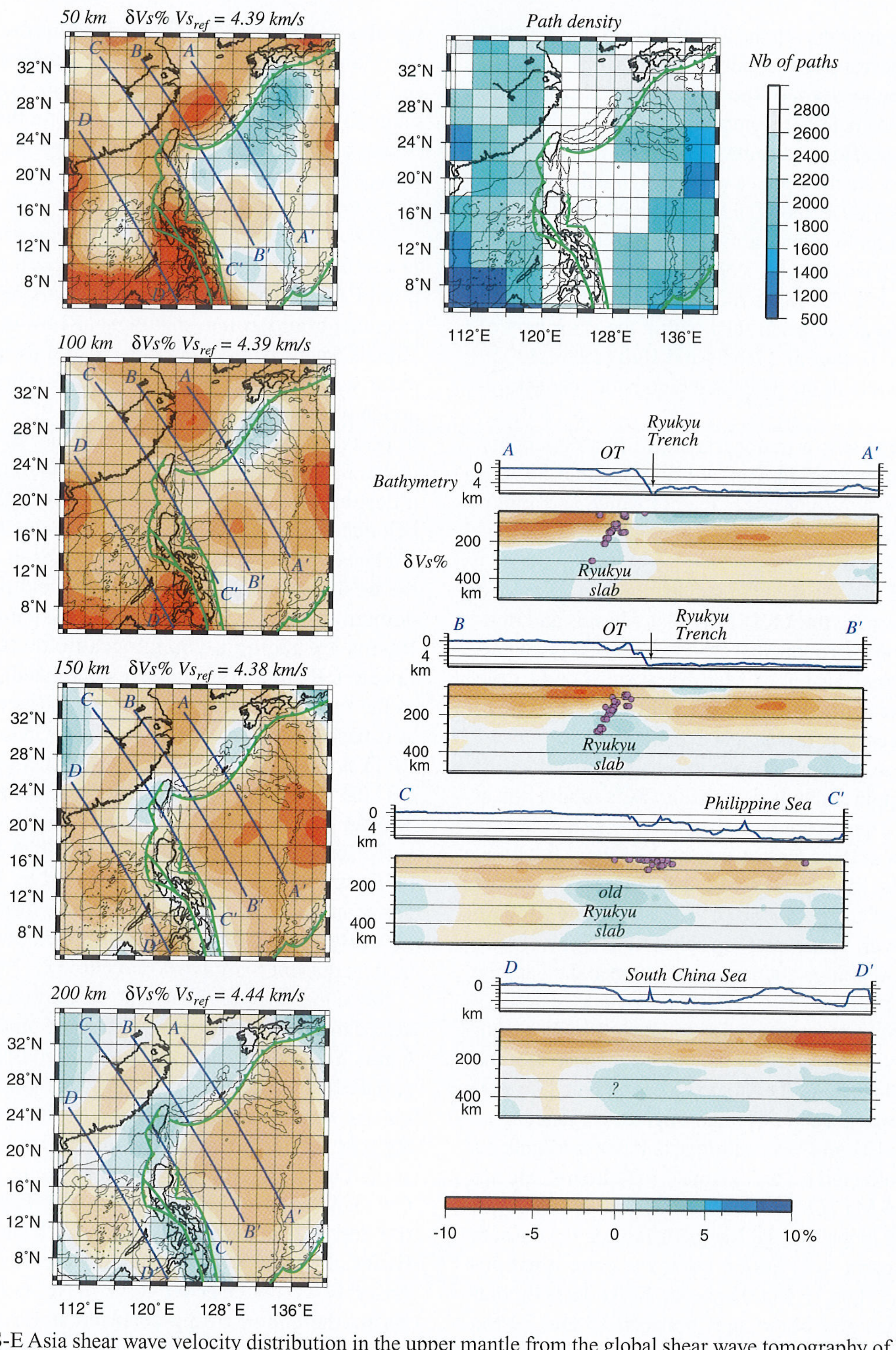


Plate 3. S-E Asia shear wave velocity distribution in the upper mantle from the global shear wave tomography of Debayle et al. [2004]. Velocity perturbations from the reference velocity are given at each chosen depth (50, 100, 150 and 200 km) in percent. The 1 and 4-km isobaths (thin dark lines) underline oceanic basins. Present-day plate boundaries are in green. The blue-white color scale on the path density map indicates the number of paths in each 400x400 km cell. Vertical cross-sections AA' to DD' located on the four maps are aligned on the Philippine and South China Sea margins. Purple circles corresponding to seismicity in a 10-km bandwidth on each side of cross-sections underline the location of the Ryukyu slab.

the Huatung Basin and beneath the Manila accretionary prism. However, large uncertainties remain in these reconstructions. Amongst them, the location of the Luzon plate is crucial, as is the age of emplacement of the northern Luzon Arc, assumed here to follow the seafloor spreading extinction in the Taiwan Sea (about 15 Ma). As the PH Sea plate was mostly bounded by subduction zones since about 15 Ma, a special attention was given to paleomagnetic constraints concerning the PH/EU rotation parameters [*Hall et al.*, 1995]. Since 15 Ma, we have assumed that the PH Sea plate rotated 15° clockwise with respect to EU. This assumption falls within the error bars of rotations given by Hall et al. [1995] and Hall [1996; 2002].

The key constraints to the proposed kinematic reconstructions are:

- The Gagua Ridge was a major transform plate boundary during the opening of the Taiwan and South China Seas. It was an about 1000-km long linear feature which accommodated the northwestward shear motion of the west PH Sea plate with respect to EU.
- Before about 15 Ma, the Ryukyu subduction zone extended southward of Taiwan to the LRTPB. Backarc basins and remnant arcs formed on the continental shelf as a succession of parallel belts of basins. About 15 Ma, the western end of the Ryukyu subduction zone moved from southwest of the Tainan Basin (118°E) to east of Taiwan (126°E).
- The construction of the intra-oceanic Luzon Arc, located north of the Philippine islands, is linked to SCS subduction at the Manila Trench and is younger than about 15 Ma. There is a large debate about the sense of vergence of the subduction zone before 15 Ma. Maleterre [1989] and Wolfe [1981] linked the construction of Luzon Island to an east-verging subduction zone. However, recent tomographic inversions clearly show a deep west-dipping subduction zone named the Sangihe subduction zone [*Rangin et al.*, 1999], which ceased activity near Luzon Island, suggesting prior to about 15 Ma a convergent motion evolving northwards into a transform motion. The internal consistency of the kinematic reconstructions presented below supports such a change of plate vergence.

North of Luzon Island, 15 radiometric dates available on volcanic rocks collected on the islands of the Luzon Arc are younger than 9.4 Ma [*Lo et al.*, 1994; *Maury et al.*, 1998; *Yang et al.*, 1996]. In the Coastal Range of Taiwan, the onshore portion of the Luzon Arc, fission track age determinations give ages younger than 16 Ma. However, K-Ar dates on outcropping andesites give some ages between 15 and 29 Ma [*Richard et al.*, 1986] but the effects of various alteration reactions as delineated by Lo et al. [1994] have not been taken into account. Samples analyzed by Richard et al. [1986] are characterized by very low K2O content, which could artificially increase radiometric ages by loss of potassium. We suggest that the change of plate vergence could have occurred about 15 Ma, simultaneously with the onset of the northward extension of the Manila Trench connecting the Ryukyu Trench through the Taiwan Sea. However, because the younger identified magnetic lineation is chron 7 (27 Ma) in the Taiwan Sea, we cannot exclude an earlier change of vergence in the Philippine islands from east to west, sometime between 15 and 27 Ma.

2.2.8. Summary of the geodynamic environment of Taiwan before its uplift. Before about 15 Ma, well before the formation of the Taiwan mountains, the Ryukyu subduction zone extended from Japan to southwest of Taiwan (LRTPB, west of Tainan Basin). This is particularly well imaged on the global shear wave velocity model of Plate 3. The oldest magnetic anomaly identified in the SCS is C17 (37 Ma) [Hsu, 2004, in press] (Figure 2). The portion of oceanic crust presently located between the LRTPB and the Manila Trench is much older than the rest of the SCS (Figure 2), and might have belonged to the proto-SCS, possibly being as old as the early Cretaceous (115-125 Ma) [Deschamps et al., 2000]. Thus, the west Taiwan Basins are facing an old portion of oceanic domain (proto-SCS, early Cretaceous) and the Pearl River Basins are facing a young oceanic domain (37 Ma, late Eocene). Rifting of the passive continental margins adjacent to the west Taiwan Basins and Pearl River Basins occurred at different periods: during the early Cretaceous east of the LRTPB and during early Tertiary (Paleogene?) west of the LRTPB. Thus, it is unlikely that shelf basins located west of Taiwan can be considered as tensional basins linked to an early Tertiary rifting episode of the adjacent portion of the SCS continental margin, as proposed by Lin et al. [2003]. However, southwest of the LRTPB, the formation of the Pearl River Basins could be reasonably linked to the rifting episode of the adjacent margin prior to chron C17 (37 Ma).

Since the early Tertiary, four belts of tensional basins separated by volcanic ridges were formed on the East and South China Seas continental shelves, the youngest being located at the oceanward edge. We suggest that these continental shelf basins were formed as backarc basins in liaison with the Ryukyu subduction zone. Rifting in the oldest backarc basin started in late Cretaceous-Paleocene time, close to mainland China. As soon as these backarc basins became too wide and the depth of the slab too big beneath the axes of backarc basins, the extensional backarc rift systems jumped eastward, probably within the active Ryukyu Arc system. In the Tainan Basin, the end of rifting occurred at 17–18 Ma (Figure 1), approximately at the time of the major plate kinematic reorganization, when subduction ceased southwest of Taiwan. Thus in our model, the geological settings of the west Taiwan and Pearl River Basins are completely different: the west Taiwan Basins are backarc basins linked to the Ryukyu subduction zone and the Pearl River Basins are tensional rifted basins linked to the early Tertiary rifting of the southeastern SCS continental margin.

3. THE UPLIFT OF TAIWAN

The Taiwan mountain belt is one of the youngest orogens in the world. Whatever are the proposed models to interpret the formation of Taiwan, all authors agree that the uplift of Taiwan results from the collision of the Luzon Arc with the EU margin. The Luzon Arc is an intra-oceanic volcanic arc that belongs to the PH Sea plate. Exposures of the Luzon Arc occur in the Coastal Range of Taiwan and the contact between the arc and the EU margin is marked by the Longitudinal Valley (Figure 5). Debate continues regarding the nature of the collided margin: whether a sedimentary monocline belonged to the EU plate and was deformed by the collision of the Luzon Arc [*Suppe*, 1981; *Suppe*, 1984], by lithospheric collision between the Luzon Arc and EU lithosphere [*Wu et al.*, 1997], by the accretion of an exotic block against the EU margin [*Lu and Hsü*, 1992] or was part of a previously formed Ryukyu Arc/backarc basin system [*Hsu and Sibuet*, 1995; *Sibuet and Hsu*, 1997] subsequently deformed by Luzon Arc collision. Davis et al. [1983] and Suppe [1984] demonstrated that the collision was propagating southward slightly faster than that of the convergence rate, which means that moving south 90 km along the collision is analogous to viewing development of the collision 1 m.y. earlier. Thus, most authors (*e.g.* Teng [1990], Malavieille et al. [2002] and Clift et al. [2003]) suggest that the evolution through time of Taiwan mountain building is a continuum represented by a series of cross-sections from the present-day Manila subduction system to the south (before collision), through middle Taiwan (collision) and northeast of Taiwan, across the southern Okinawa Through and Ryukyu subduction system (post-collision).

Starting from East Asia plate kinematic reconstructions (Figure 4), we define the location of the western boundary of the PH Sea plate (point A) that bounds the termination of the Ryukyu Benioff zone and triggers the occurrence and development of a tear fault within the EU plate. We try to explain the process of Taiwan mountain building within a simple plate tectonic context, which includes a comprehensive history of westward migration of the collision. For that purpose, we first clarify a series of points concerning the geometry of plate boundaries, as well as the nature of deformation and associated stress patterns.

3.1. Location of Western Philippine Sea Plate Boundary (Point A)

At about 15 Ma, the vergence and location of the Luzon subduction zone shifted from east to west of the Philippine archipelago [*Maleterre*, 1989; *Wolfe*, 1981], giving rise to the Manila subduction zone. Later on, the Luzon Arc started to form as an intra-oceanic arc. At first the newly created Luzon Arc was probably a modest topographic feature and its northern part subducted beneath EU as part of the PH Sea plate. It was only at 9-6 Ma that the young Luzon Arc acquired a significant topographic expression, resisting subduction and started to collide with EU, giving rise to the proto-Taiwan chain as part of the deformed EU border [*Sibuet et al.*, 2002]. In Taiwan, the Coastal Range corresponds to the outcropping portion of the Luzon Arc, and the Longitudinal Valley is the plate boundary between the PH Sea and EU plates (Figure 5). The proto-Taiwan chain was located at the position of the present-day southwestern OT and adjacent northern continental shelf [*Hsiao et al.*, 1999; *Sibuet et al.*, 1998] (Figure 6). Its eastern termination has been carefully mapped [*Hsiao et al.*, 1999]. In this scheme, the eastern limit of thrust faults (at 124°E) is located in the N307° direction with respect to the location where the Ryukyu Trench changes direction from NE-SW to E-W (at 126°E, Figure 6). From the PH/EU kinematic parameters, the mean westward motion of the Luzon Arc is 4.5 cm/yr for the last 8 M.y. and the colliding point of the Luzon Arc with EU continuously migrated westwards for a total amount of 400 km (Figures 4 and 5).

The Ryukyu Trench terminates in its westward direction at 122°E, but the subducted slab is recognized further west, beneath northern Taiwan [*Kao et al.*, 2000] (Figure 6). The boundary of the PH Sea plate follows the subducted plate edge. If Point A is the surface projection of the western limit of the Ryukyu slab, one can delineate it from either the distribution of intermediate-depth and/or deep focus seismicity, or a high velocity anomaly on tomographic images. Global tomographic models (*e.g.* Lallemand et al. [2001] and Rangin et al. [1999]) are too coarse to shed light on such a small region, while local tomographic models do not extend deep enough into the lithosphere (*e.g.* [*Cheng et al.*, 2002; *Hsu*, 2001]). Thus, Point A is not well constrained in northern Taiwan, but corresponds to the western boundary of the deep seismicity [*Font et al.*, 1999] (Plate 4) and extends approximately from Hualien to somewhere between Chungli and the Taitun volcano (Figure 5).

Northeast of Point A, focal mechanisms show that earthquakes belonging to the Ryukyu seismogenic zone are associated with low-angle thrust faults over depths of 10 to 35 km. Above the Benioff Zone, within the EU plate, shallow crustal extensional earthquakes occur in the southwestern OT and under the Ilan Plain. On land, several active faults with a normal component have been identified (Figure 5). The NE-SW oriented Shanjiao Fault is a transtensional fault [*Lee and Wang*, 1987], conjugate to the transpressional NW-SE oriented Nankan Fault [*Hsu and Chang*, 1979; *Tsai*, 1986]. Further north, the

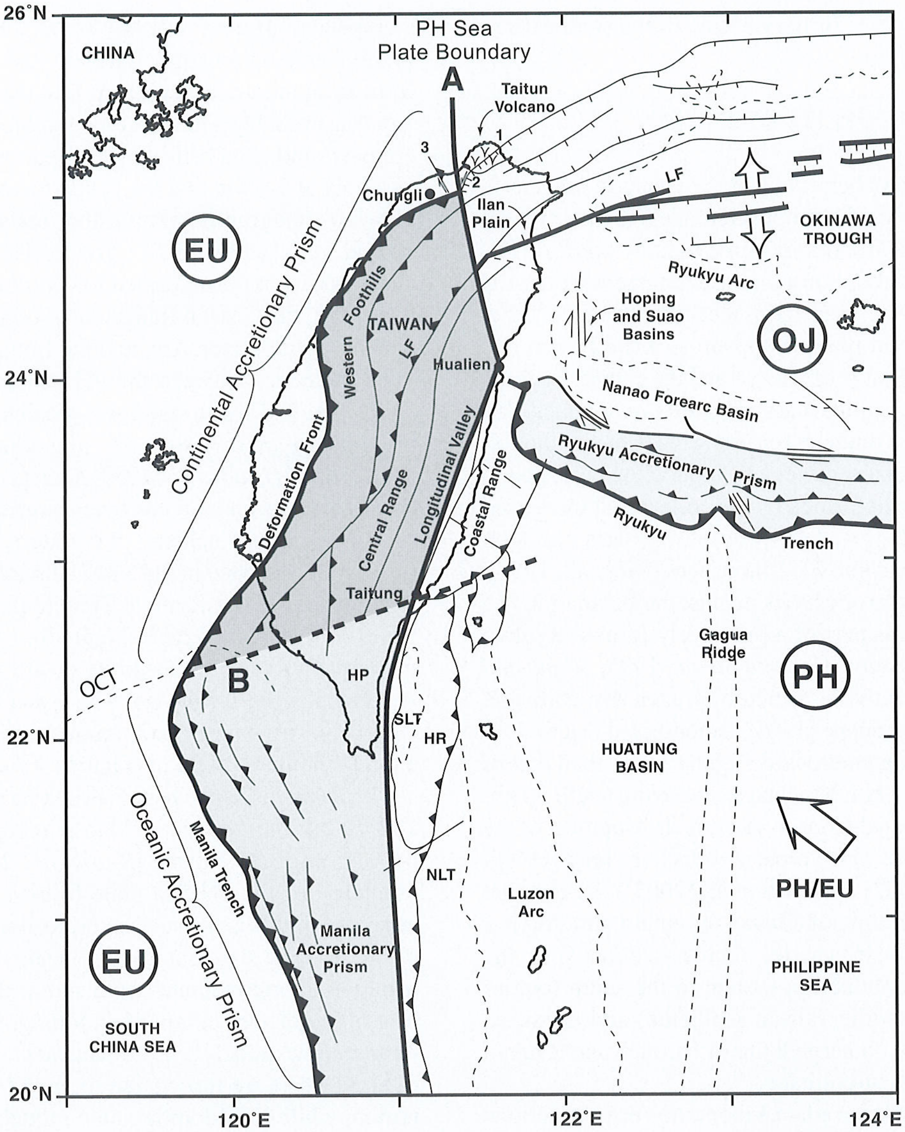


Figure 5. Geodynamic context of the Taiwan area established from Deffontaines et al. [2001], Teng and Lee [1996] for northern Taiwan, Huang et al. [1992] for the prolongation of these features on the continental shelf north of Okinawa Trough, Sibuet et al. [1998] for the Okinawa Trough, Lallemand et al. [1999] for the Ryukyu accretionary wedge and forearc basins, Malavieille et al. [2002] for the Coastal Range, Luzon Arc, southern Longitudinal Trough (SLT) and North Luzon Trough (NLT) forearc basins, Liu et al. [1997] for the Manila accretionary prism. Point A is the western boundary of the PH Sea plate and Point B is the ocean-continent transition (OCT) zone within the EU subducted slab (in the prolongation of the South China Sea OCT). East of Point A, in northern Taiwan and on the northeastern continental shelf, normal faults are reactivated thrust faults [*Hsiao et al.*, 1999]. Between Points A and B and south of Point B, features are compressional. In light gray, Manila oceanic accretionary prism above the EU oceanic slab; in dark gray, Taiwan continental accretionary prism above the EU continental slab. HP, Hengchun peninsula; HR, Huatung Ridge; LF, Lishan Fault and its extension in the Okinawa Trough. 1, Chinshan Fault; 2, Shanjiao Fault; 3, Nankan Fault [*Teng and Lee*, 1996].

Taitun volcano is cut by a set of Quaternary normal and strike-slip faults [*Lee and Wang*, 1988] and the modern Chinshan Fault is in fact a series of normal faults with a strike-slip component [*Chen and Yeh*, 1991; *Lee and Wang*, 1988].

West of Point A, earthquakes show a consistent orientation of P-axes in the direction of collision, with crustal thrust faulting mechanisms mostly oriented in the N290° direction. From the Longitudinal Valley to the western coast of Taiwan, earthquakes correspond to shallow thrust events (<20 km) generally of lower magnitude than east of Taiwan, suggesting that the Central Range is relatively aseismic [*Kao et al.*, 1998]. According to the above observations, compressive earthquakes mechanisms and reverse faults testify to contractional motions within the upper plate, west of Point A.

Consequently, in northern Taiwan, the surface projection of the PH Sea plate boundary (Point A) is associated with a zone of strain change in the EU crust (extensional in the OT and northern Taiwan and compressional in the rest of Taiwan).

South of Hualien, the PH Sea plate boundary follows at the surface the Longitudinal Valley (Figure 5) (*e.g.* [*Lu and Hsü*, 1992; *Malavieille et al.*, 2002]). By convention, we have taken the surface expression of this limit as the PH Sea plate boundary. South of Taitung, the location of Point A is poorly defined. Arbitrarily, this boundary has been located west of the Luzon Arc and adjacent forearc basins. In this scheme, the Manila accretionary prism corresponds to deformed sediments belonging to the EU plate (South China Sea), with the rest of the EU plate subducting beneath the PH Sea plate. The Manila Trench continues northwards by the deformation front considered as the western boundary of the Taiwan deformed zone. Similarly, the Central Range and Foothills are considered as a continental accretionary prism, which includes sediments and upper crust material belonging to the EU plate, the rest of the EU plate subducting beneath the PH Sea plate. Thus, a major accretionary prism has been identified west of the boundary of the rigid PH Sea plate and consists of deformed sediments previously belonging to the EU plate, whose nature is continental in the north and oceanic in the south.

3.2. Ocean-Continent Transition Zone Within the Eurasian Slab (B)

Recent studies of the structure of the lithosphere and mantle in East Asia based on seismic tomographic images [*Bijwaard et al.*, 1998; *Kuo et al.*, 2003; *Lallemand et al.*, 2001; *Rangin et al.*, 1999; *Tajima et al.*, 1997] show that the Manila slab is a continuous feature extending northwards up to the latitude of Hualien (northern end of the Coastal Range) where it can be followed down to 500-600 km [*Rangin et al.*, 1999]. However, the deep seismicity associated with the Manila slab essentially occurs south of Point B (Plate 4) and is almost absent north of B. Point B extends from the Manila Trench to south of the Longitudinal Valley and terminates at 121.8°E. On tomographic images, a slight change in the orientation of the slab occurs at its intersection with Point B, as well as an increase of the slab steepness north of B. South of Point B, the distribution of seismicity at depths >80 km represents the geometry of the Manila SCS subducted slab (Plate 4). South of Point B, there is a good corre-

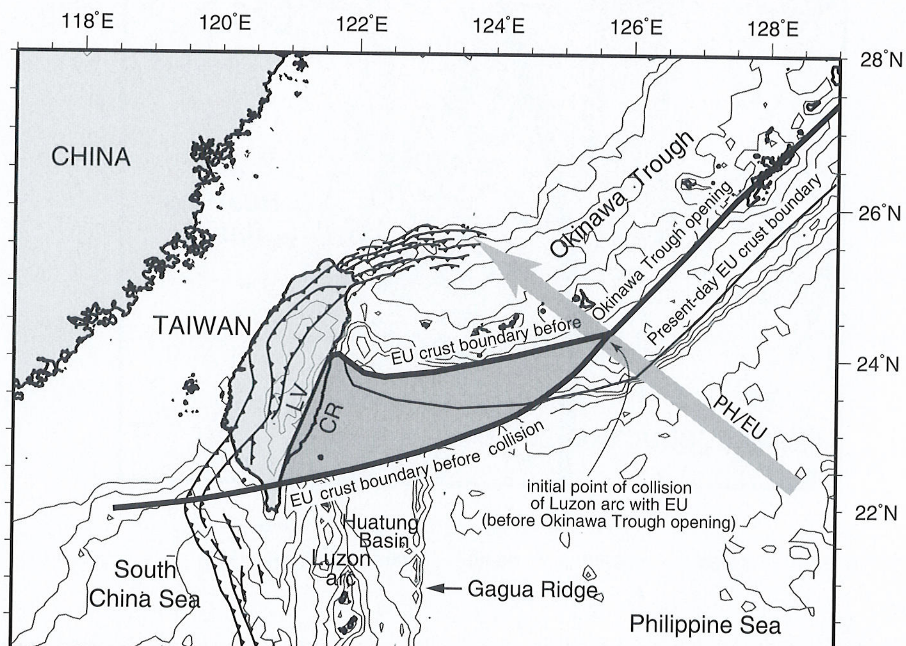


Figure 6. Simplified bathymetry (isobath spacing, 1 km) near Taiwan with the portion of Eurasian continental crust (in dark gray) subducted beneath the Luzon Arc, during its collision with Eurasia [*Sibuet and Hsu*, 2004]. Proto-Taiwan thrust faults northeast of Taiwan from Hsiao et al. [1999]. The position of Eurasian crust boundary before the Okinawa Trough opening is with respect to Eurasia supposed to be fixed (75 km of extension, except west of 124°E longitude where extension decreases to zero [*Sibuet et al.*, 1995]). CR, Coastal Range; LV, Longitudinal Valley.

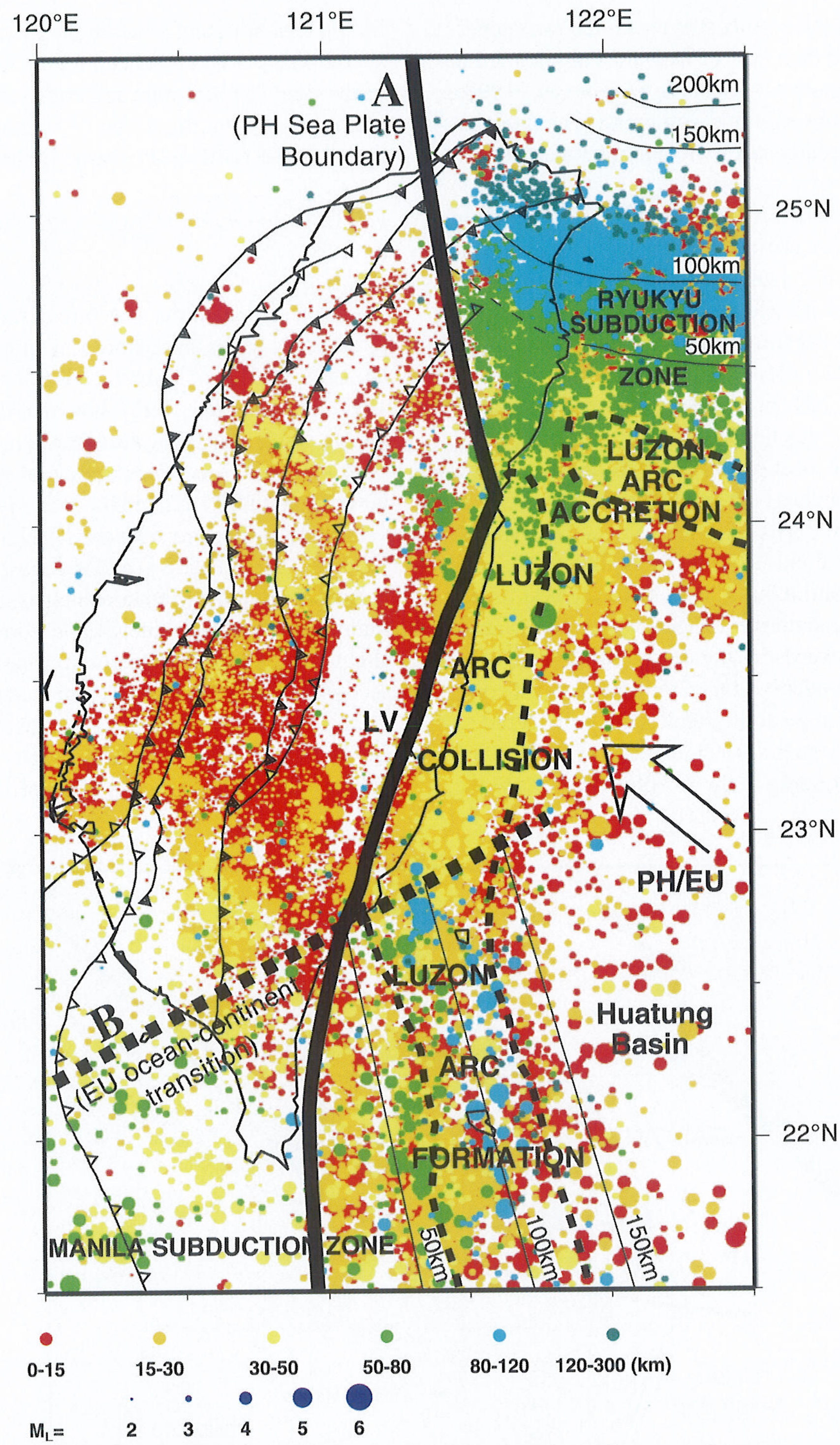


Plate 4. Seismicity map of Taiwan with earthquakes recorded during the 1991-1997 period by the Central Weather Bureau in Taiwan [*Sibuet and Hsu*, 2004]. Deeper earthquakes are plotted above shallower earthquakes in order to underline the Ryukyu and Manila slabs. Point A is the western boundary of the PH Sea plate and Point B the ocean-continent transition zone within the EU subducted slab. Isobaths of subducted slabs are every 50 km. South of Point B, the Luzon Arc is forming; between Points A and B, the Luzon Arc is colliding with the Eurasia margin. There are only a few deep slab events east of the Longitudinal Valley (LV); East of Point A, the upper portion of the Luzon Arc is accreted against the Ryukyu forearc.

spondence between the geometries of the slab deduced from both tomographic images and the location of deep earthquakes. However, north of B, the paucity of deep earthquakes (Plate 4) suggests that the portion of the slab imaged on tomographic data might be detached [*Lallemand et al.*, 2001]. Point B is approximately located in the northeastward prolongation of the northern continental margin of the SCS, suggesting that the slab might be of continental and oceanic nature north and south of Point B, respectively. In other words, within the EU slab, Point B would follow the ocean-continent transition and might underline the southern limit of the possible northward-detached portion of the SCS slab. The considerable intraplate deformation occurring north of Point B would be consistent with the tectonic setting of the regional collision of the Luzon Arc with EU, and a continental subduction, whereas the seismicity pattern in the south would correspond to the subduction of the oceanic part of the EU lithosphere (South China Sea) and the formation of the intra-oceanic Luzon Arc.

According to Malavieille et al. [2002], an important tectonic boundary corresponding to the location of Point B exists offshore, south of the Coastal Range (Figure 6). Although the Longitudinal Valley is bounded by the Coastal Range west-vergent thrusts, the Southern Longitudinal Trough is controlled by the east-verging antiformal sedimentary structures of the Huatung Ridge, the offshore equivalent of the Lichi Mélange interpreted as a syntectonic mélange formed during the collision of the Luzon Arc with EU [*Sibuet and Hsu*, 1997]. The Huatung Ridge is thrusting over the Luzon Arc. Thus, according to Malavieille et al. [2002], the Southern Longitudinal Trough is the offshore equivalent of the collisional basin in which the Lichi Mélange was deposited earlier to the north and the Huatung Ridge consists of folded and faulted sediments belonging to this basin.

In summary, Point B lies in the prolongation of the base of the northeastern SCS continental margin. We suggest that the oceanic SCS slab might be of different nature (continental versus oceanic) north and south of Point B, the Luzon intra-oceanic arc being formed only south of Point B, above the subducting EU oceanic lithosphere. With a PH/EU convergent motion in the N307° direction, i.e. with a westward component of the PH Sea plate with respect to EU, the Manila subduction zone would progressively move in the southwestward direction with respect to EU. We now discuss the nature and geographical extension of the portion of subducted slab located north of Point B.

3.3. Nature and Origin of the Portion of East Dipping Slab Located East of the Coastal Range

What is the nature of the slab located north of Point B? It is now widely accepted that the lower part of the Eurasian continental lithosphere subducted since the beginning of collision while the sediments and the upper part of the crust were deformed and uplifted to form the Taiwan orogen (*e.g.* Suppe et al. [1981], Brusset et al. [1999]) and then were eroded. Several alternatives to this model have been proposed: slab break-off to account for a flipping of subduction polarity in northern Taiwan [*Teng et al.*, 2000] or rapid subduction and retreat of the continental lithosphere to explain the high pressure, low temperature terranes of the Tananao Complex [*Malavieille et al.*, 2002]. This last mechanism implies the subduction of buoyant continental lithosphere as a consequence of several factors. Subduction of the Indian plate beneath the Himalaya, dragged continental lithosphere into the subduction zone due to the force of previously subducted oceanic lithosphere, eclogitized at 70-90 km, and resulting in a density increase and scraping off the upper part of the continental crust. Hsu and Sibuet [1995] have also shown that the portion of the Ryukyu Trench presently located east of 126°E longitude (before OT opening) is located along the same regular trend as the former Ryukyu Trench located on the northeastern SCS margin, and suggesting a smoothed shape for the Eurasian margin before collision (Figure 6).

Assuming that the EU continental lithosphere started to subduct at the onset of the Luzon Arc-EU collision, there are three different ways to estimate the shape and the surface of continental lithosphere subducted during the Taiwan Orogeny:

1) Sibuet et al. [2002] suggest that the Luzon Arc started to collide with the EU margin at 9 Ma, at 126°E longitude with respect to the present-day position of the Ryukyu subduction zone, where the Ryukyu Trench changes direction from NE-SW to ENE-WSW. Since that time, the Luzon Arc has moved westward at a mean velocity of 4.5 cm/yr [*Sibuet et al.*, 2002]. If the width of the subducted continental lithosphere regularly increased from a zero-length at 9 Ma to 150 km today (the present-day length of the Coastal Range), the whole surface of subducted continental lithosphere would be 30,000 km^2.

2) Using tomographic results, Lallemand et al. [2001] have restored the horizontal shape of the subducted PH Sea and EU (South China Sea) slabs. To estimate the geometries of the EU and PH Sea plates, they infer the existence of a major lithospheric tear, which follows the base of the Ryukyu margin, propagating through the EU continental lithosphere. The fragment of subducted EU continental crust is of triangular shape with its narrow summit located at the base of the Ryukyu forearc (near 126°E longitude), and the opposite side corresponding approximately to the length of the Longitudinal Valley. Its surface is about 30,000 km^2.

3) Figure 6 shows a simplified bathymetric map of the Taiwan area with: i) the approximate boundary of the present-day EU continental crust, which is assumed to be of constant thickness (approximating the 3000-m isobath east of Taiwan

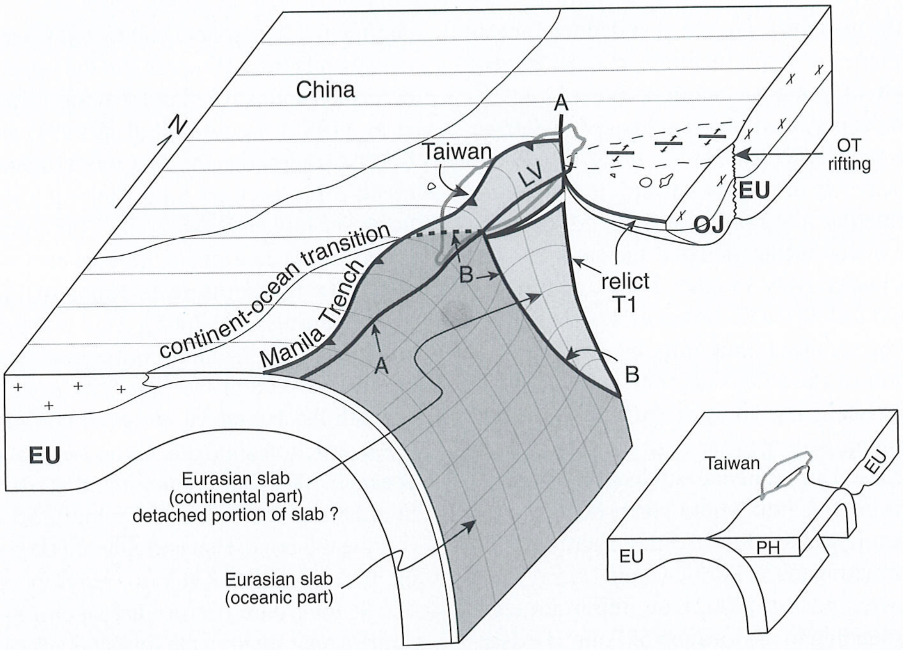


Figure 7. Sketch showing the 3-D shape and extent of the Eurasian (EU) slab after removal of the Philippine (PH) Sea plate (modified from Lallemand et al. [2001]). Thin lines approximately every 100-km give the scale of the EU subducted slab. A is the trace of the western boundary of the PH Sea plate. East of Point A, EU is torn by the westward motion of the PH Sea plate and thus has two boundaries (relict T1). Point B is the boundary between the oceanic (dark gray) and continental (light gray) parts of the EU slab located in the prolongation of the South China Sea ocean-continent transition zone. LV, Longitudinal Valley; OT, Okinawa Trough; OJ, Okinawa-Japan plate.

and 2000-m west of Taiwan, because this margin is highly sedimented (labeled "present-day EU crust boundary" in Figure 6)); ii) the restored position of the EU continental crust boundary east of Taiwan (labeled "EU crust boundary before OT opening" in Figure 6) before OT opening (extension of 75 km, except west of 124°E longitude where the extension regularly decreases to zero west of the Ilan Plain, at the tip of the OT [*Sibuet et al.*, 1995]); iii) the EU continental crust boundary (labeled "EU crust boundary before collision" in Figure 6) before collision of the Luzon Arc and opening of the OT at 6 Ma. The gray area in Figure 6 corresponds to the surface of the EU continental crust subducted since the onset of the Luzon Arc collision (50,000 km^2). This estimate differs from previous works because of the uncertainty in the position of the ocean-continent transition before the OT opening, as well as in the width of the subducted continental lithosphere, which probably increased rapidly during the first 3 m.y. and less during the last 6 m.y. In addition, a possible shortening of the EU plate in Taiwan was not taken into account.

In conclusion, we favor the subduction of a piece of continental lithosphere of 30,000-50,000 km^2 in extent beneath the Coastal Range and the Huatung Basin. The shallow part of the EU crust, located west of the undeformed PH Sea plate and including sediments, was scraped off and accreted in the proto- and present-day Taiwan chain. Even if a major detachment is located at a depth of about 10 km beneath Taiwan [*Carena et al.*, 2002] and merges eastward with the top of the EU slab, the overlying upper crust and sediments are highly deformed and do not belong to the undeformed PH Sea plate. Figure 7 shows a 3-D cartoon of the EU plate established from tomographic results, with the portion of EU slab of continental origin in light gray. Before the collision of the Luzon Arc with EU the northern edge of the EU slab was connected to the base of the Ryukyu forearc.

3.4. Present-day Plate Kinematics

In the Taiwan area, the location of major plate boundaries has long been a matter of debate. The present-day plate kinematic system is presented in the three panels of Plate 5, each panel corresponding to the PH Sea, EU and Okinawa-Japan (OJ) plates, respectively. The overall plate motion in a EU reference frame corresponds to a westward motion of the PH Sea plate into a tear of the EU plate. As backarc extension occurred in the OT, an additional Okinawa-Japan plate is required.

Plate 5 shows the distribution of crustal deformation inferred from the pattern of seismicity, focal mechanisms [*Kao et al.*, 2000; *Kao et al.*, 1998; *Wu et al.*, 1997], and the velocity field calculated from Taiwan GPS stations before [*Yu et al.*, 1997] and after the Chi-Chi earthquake (*e.g.* Yu et al. [2001], and Loevenbruck et al. [2001]). The development of a tear fault within the EU lithosphere is triggered through time by the western motion of the PH Sea plate with respect to EU (Plate 5). East of Point A, the edge of the EU plate is torn and thus has two boundaries (relict T1 in Figure 7). One is the upper plate, trench edge of the Ryukyu forearc, the other is the deeply subducted border of the EU slab, underlying the PH Sea plate. In this scheme the OJ plate is not only increasing in size in the N-S direction due to the OT opening but also in the westward direction. As the tip of the OT axis is moving westward, a portion of the EU plate located west of Point A is progressively incorporated in the OJ plate, becoming part of the Ryukyu Arc and forearc.

The Ryukyu accretionary prism, about 50 km wide, displays numerous slope breaks corresponding to the emergence of thrust ramps and a pronounced thrust front (Figure 5). The Ryukyu accretionary prism cannot be considered as a rigid plate or part of the rigid OJ plate. A major right-lateral strike-slip fault, localized at the rear of the wedge, accommodates the strain partitioning caused by the 40° PH/EU oblique convergence [*Lallemand et al.*, 1999]. The Ryukyu accretionary wedge is moving laterally along this transcurrent fault mapped from 122°E to 124°E longitude and located at the toe of the rigid arc and forearc backstop (relict T1 in Plate 5c). Recently published swath bathymetric data east of 124°E [*Matsumoto et al.*, 2001] suggest that the transcurrent fault likely extends eastward to 126°E, merging with the Ryukyu Trench where it changes direction (Figure 5). We assume that the relict tear fault T1 extends from 122°E to 126°E longitude.

Plate 5 shows synthetic profiles across each of the three main domains and from the OT to the Manila accretionary prism. The nature and geometry of the EU slab differ on each side of Point B: south of Point B, the EU slab is oceanic (Plate 5, profile 3) and the Manila accretionary prism consists of sediments scraped off the EU plate, but also derived from the erosion of Taiwan; north of Point B, the EU slab is continental (Plate 5, profile 2). Thrust faults in the Western Foothills merge with a 10-km deep flat main detachment beneath Taiwan, which dives down to 30-60 km depth beneath eastern Taiwan (*e.g.* Brusset et al. [1999], Lallemand et al. [2000] and Carena et al. [2002]). As for the Manila accretionary prism, the material above this flat ramp initially belonged to the EU plate (Plate 5, profile 2). The Longitudinal Valley represents the contact between the EU and PH Sea plates. The deep portion of the EU slab located east of the Coastal Range is believed to be detached from the rest of the EU continental plate (Figure 7) [*Carena et al.*, 2002; *Lallemand et al.*, 2001]. However, it is not clear whether the slab is completely detached between relict Fault T1 and Point B as suggested by Lallemand et al. [2001] or whether it is partially torn. Solving this question has consequences for the type of motion within the EU slab, along its former OCT. In Figure 7, we have adopted the model of Lallemand et al. [2001]. As Point B is a significant boundary within the EU slab, the notion that the orogen records a continuum from subduction to collision has to be questioned, even if the result is only minor differences in the surface geology.

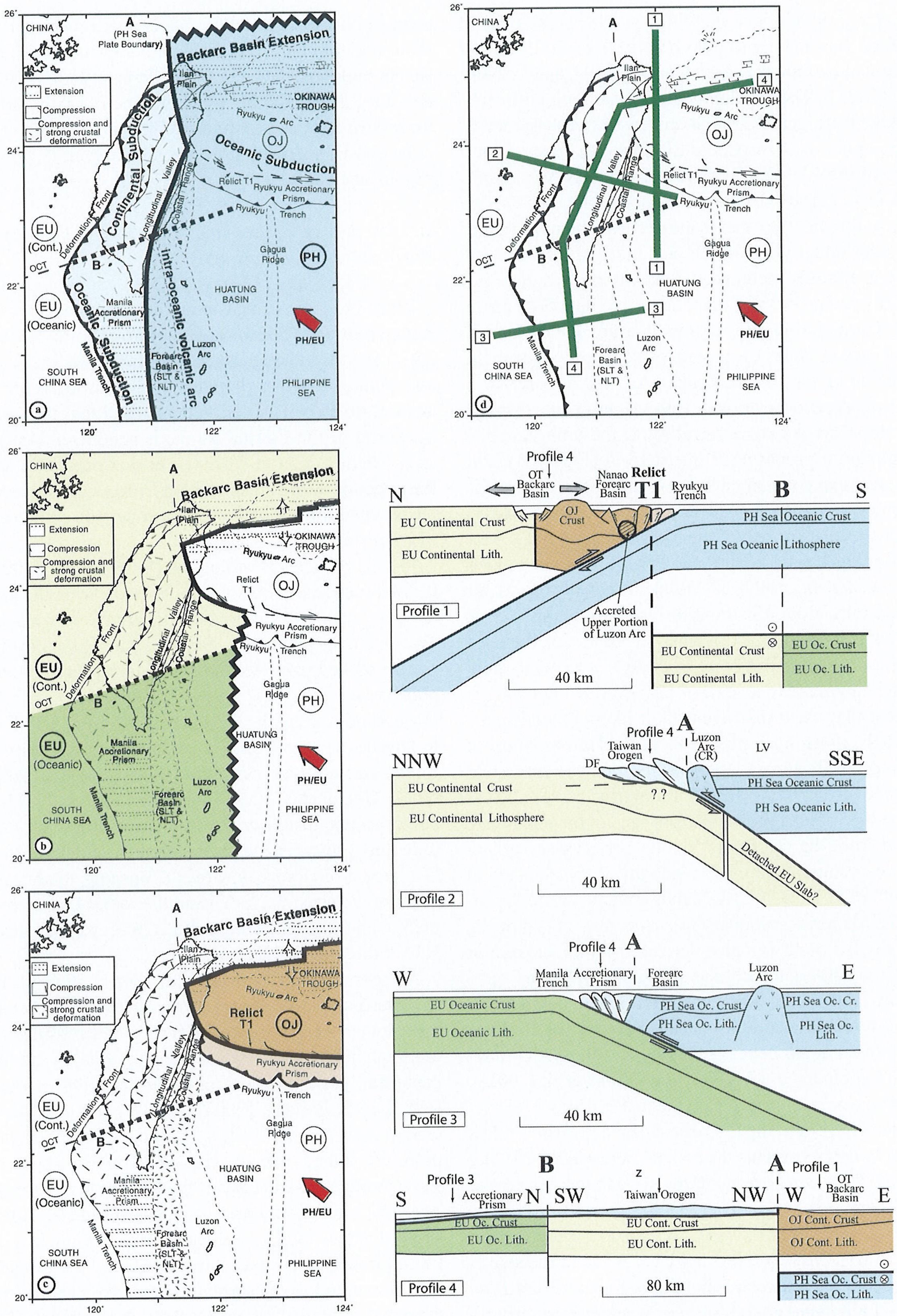
3.5. Arc-arc Collision: What Is the Fate of the Luzon Arc North of the Coastal Range?

There is a large debate about the prolongation of the Coastal Range (emerged portion of Luzon Arc) north of Hualien (Figures 5 and 6). Although some authors [*Cheng et al.*, 1996; *Wang and Chiang*, 1998] suggest that the Luzon Arc is not considered to exist north of Hualien, others argue that the northern prolongation of the Coastal Range is subducted beneath Taiwan (*e.g.* Chemenda et al. [1997]). In this paper, we suggest that the upper portion of the northern Luzon Arc is accreted against the Ryukyu forearc while the lower portion subducted with the PH Sea slab.

PH/EU parameters of rotation [*Sibuet et al.*, 2002] show that the oblique Luzon Arc collision started during the late Miocene (6-9 Ma) near 126°E longitude, where the present-day Ryukyu Trench changes direction. Since that time, the northward component of the Luzon Arc motion with respect to EU totals about 400 km and the colliding point of the Luzon Arc has moved through time along the Ryukyu margin from 126°E to 121.5°E longitude. The uplift of proto-Taiwan is the consequence of Luzon Arc collision with the Eurasian margin [*Hsiao et al.*, 1999]. Huang et al. [1992] and Hsiao et al. [1999] have clearly demonstrated that proto-Taiwan existed between 124°E longitude and Taiwan, beneath the continental shelf and northern OT slope. Therefore, before the OT opening, the easternmost end of proto-Taiwan (at 124°E) and the initial Luzon Arc collision point with the Ryukyu forearc (at 126°E) were aligned along the N307° direction (Figure 6).

As soon as the Luzon Arc collision point with EU shifted westward, the eastern part of the proto-Taiwan chain was no more under compression, was eroded and subsided mostly as a consequence of the westward propagation of the OT backarc basin into proto-Taiwan [*Sibuet et al.*, 1998]. Note that Teng [1996] and Clift et al. [2003] link the erosion of the proto-Taiwan to gravitational collapse, not to the westward propagation of the OT. If the Luzon Arc collision with EU is quite well understood through time, the fate of the 400-km long portion of Luzon Arc, which has disappeared since 9-6 Ma and has not been identified north of the Coastal Range remains unclear. Clift et al. [2003] indicate that it must be accreted to the EU margin.

A decoupled-wedge structure has been recently imaged on a deep multi-channel seismic profile shot across the Hidaka col-



CHINA
A
(PH Sea Plate Boundary)
Backarc Basin Extension
Extension
Compression
Compression and strong crustal deformation
Ilan Plain
OKINAWA TROUGH
Ryukyu Arc
OJ
Oceanic Subduction
Continental Subduction
Longitudinal Valley
Coastal Range
Relict T1
Ryukyu Accretionary Prism
Ryukyu Trench
Deformation Front
EU (Cont.)
OCT
B
EU (Oceanic)
Oceanic Subduction
Manila Accretionary Prism
Intra-oceanic volcanic arc
Gagua Ridge
PH
HUATUNG BASIN
PH/EU
Luzon Arc
Forearc Basin (SLT & NLT)
SOUTH CHINA SEA
Manila Trench
PHILIPPINE SEA
a
b
c
d
26°
24°
22°
20°
120°
122°
124°
1
2
3
4
Profile 4
OT Backarc Basin
Nanao Forearc Basin
Relict T1
Ryukyu Trench
N
S
EU Continental Crust
EU Continental Lith.
OJ Crust
PH Sea Oceanic Crust
PH Sea Oceanic Lithosphere
Profile 1
Accreted Upper Portion of Luzon Arc
40 km
EU Continental Crust
EU Continental Lith.
EU Oc. Crust
EU Oc. Lith.
Taiwan Orogen
Luzon Arc (CR)
LV
NNW
SSE
DF
??
EU Continental Lithosphere
PH Sea Oceanic Lith.
Detached EU Slab?
Profile 2
W
E
Manila Trench
Accretionary Prism
Forearc Basin
EU Oceanic Crust
EU Oceanic Lith.
PH Sea Oc. Crust
PH Sea Oc. Lith.
PH Sea Oc. Cr.
Profile 3
Profile 3
Accretionary Prism
SW
NW
Z
Taiwan Orogen
Profile 1
OT Backarc Basin
EU Oc. Crust
EU Oc. Lith.
EU Cont. Crust
EU Cont. Lith.
OJ Cont. Crust
OJ Cont. Lith.
PH Sea Oc. Crust
PH Sea Oc. Lith.
80 km
Profile 4

lision zone, which was created by the convergent motion of the Kuril Arc with respect to Central Hokkaido [*Tsumura et al.*, 1999]. The upper half (upper crust and upper portion of the lower crust, 5.5 to 6.0 km/s) of the Kuril Arc was thrust westward over the northeast arc while the lower half (lower portion of the lower crust (6.6 km/s) and the upper mantle) descended [*Ito et al.*, 2000]. The geometry of reflectors is thus consistent with the decoupling of the lower crust of the colliding Kuril Arc during active collision with Japan [*Tsumura et al.*, 1999]. This example of arc-arc collision supports a model of continental growth by arc accretion with decoupling and subduction of the lower crust. Continental material is believed to form by arc magmatism. However, the geochemical composition of the continental crust requires both the loss of lower crustal cumulates in the mantle and crystal fractionation in the upper arc crust [*Draut et al.*, 2002]. Such a geochemical model is in agreement with the Tsumura et al. [1999] model of arc accretion.

P-wave arrival times obtained from seismic land stations in Taiwan and in the western Ryukyu islands were inverted to derive the velocity structure down to 50 km beneath Taiwan and the western Ryukyu subduction zone [*Hsu*, 2001]. A 220-km long linear feature was shown east of Taiwan beneath the upper Ryukyu forearc slope and the Nanao Basin (Plate 6), but not further east due to a lack of data. This feature is about 50 km wide and is located at a depth between 15 and 45 km. Its thickness of 30 km is similar to that seen in the accreted portion of the Hidaka collision zone [*Tsumura et al.*, 1999]. With respect to the adjacent velocity structure of the Ryukyu subduction zone, this feature presents a slight high-velocity zone in its upper part (15-20 km) and a low velocity zone in its lower part (30-45 km), as seen in the Hidaka collision zone.

A low-velocity zone (5.5 km/s compared to 7.0 km/s in the adjacent terranes) has been also seen at the same location on a N-S wide angle and refraction profile shot at 122.5°E longitude [*Wang et al.*, 2002]. The low velocity zone has been identified within the upper plate beneath the upper Ryukyu forearc slope and the Nanao Basin. Though this light body is less well located on tomographic results, both tomographic and refraction data suggest the existence of such a body beneath the Nanao Basin and upper Ryukyu forearc slope.

An elongated -180 mGals free-air gravity anomaly low exists in this region [*Hsu et al.*, 1998]. Its minimum axis is located 10-km north of the Nanao Basin axis and extends to 125°E longitude, where it abuts against a 200-km long positive feature located north of the Ryukyu Trench (Plate 2). Forward modeling suggests that the gravity low might be associated both with the 10-25 km deep low-density body (accreted upper part of the Luzon Arc) and the overlying Nanao forearc Basin, a topographic depression partly filled with sediments, and located seaward of the accreted upper Luzon Arc and north of the Ryukyu accretionary wedge. Thus, both the Nanao Basin and the underlying low-density body are parallel features, which disappear east of 125°E longitude.

In summary, we propose to apply the lithospheric model of Tsumura et al. [1999] since the beginning of the collision of the Luzon Arc with the Eurasian margin. Brittle deformation occurred in the Luzon Arc until decoupling took place in the upper lower crust, similar to that seen in northern Hokkaido [*Ito et al.*, 2000; *Tsumura et al.*, 1999] and in the Aleutian islands [*Klemperer and Fliedner*, 2000]. As soon as the northern portion of the Luzon Arc started to collide with EU, the upper part of the arc (25-30 km), which seems to be there less dense than the adjacent continental crust [*Hsu*, 2001; *Wang et al.*, 2002], accreted against the Ryukyu forearc. The lower part of the arc subducted beneath EU together with the underlying PH Sea oceanic crust. The Nanao Basin is thus a partly filled depression created seaward of the accreted upper Luzon Arc and north of the Ryukyu accretionary prism.

Profiles 1 and 2 (Plate 5) show the difference in fate of the Luzon Arc: accretion of the upper part of the Luzon arc against

Plate 5. (a, b and c): Present-day plate tectonic context and types of deformation for each of the three Philippine Sea, Eurasia and Okinawa-Japan plates [*Sibuet and Hsu*, 2004]. EU, Eurasia; NLT, Northern Luzon Trough; OCT, ocean-continent transition; OJ, Okinawa-Japan; PH, Philippine; SLT, Southern Longitudinal Trough. Point A is the western boundary of the PH Sea plate and Point B is the OCT zone within the EU slab (in the prolongation of the South China Sea OCT). Relict fault T1 is located beneath the transcurrent fault localized at the rear of the present-day Ryukyu accretionary prism. Based on the distribution and magnitude of earthquakes [*Kao et al.*, 2000] and GPS displacements [*Yu et al.*, 1997], areas of diffuse extension exist in northern Taiwan, OT and in the Manila accretionary prism. The portion of Taiwan Island west of Point A and the northern part of the Manila accretionary prism are moderately deformed by compressive stress. The northern Luzon, the Coastal Range and the Manila forearc basins (SLT and NLT) are severely deformed by compressive stress; the Okinawa Trough, northern Taiwan (east of Point A) and the southern portion of the Manila accretionary prism are under extension [*Kao et al.*, 2000; *Teng and Lee*, 1996]. The PH Sea plate is in blue and the Philippines islands to northern Taiwan accretionary prism is in light blue. The EU plate is in green for its oceanic part and in yellow for its continental part. The OJ plate is in orange and the Ryukyu accretionary prism in light orange. (d) Green dashed lines are locations of the four synthetic profiles, which show the nature and boundaries of EU, PH Sea and OJ plates [*Sibuet and Hsu*, 2004]. In green, EU oceanic lithosphere; in yellow, EU continental lithosphere; in blue, PH Sea oceanic lithosphere, in orange, OJ lithosphere. In light blue, Manila to northern Taiwan accretionary prism; in light orange, Ryukyu accretionary prism. The scale of profile 4 is half that of the other profiles.

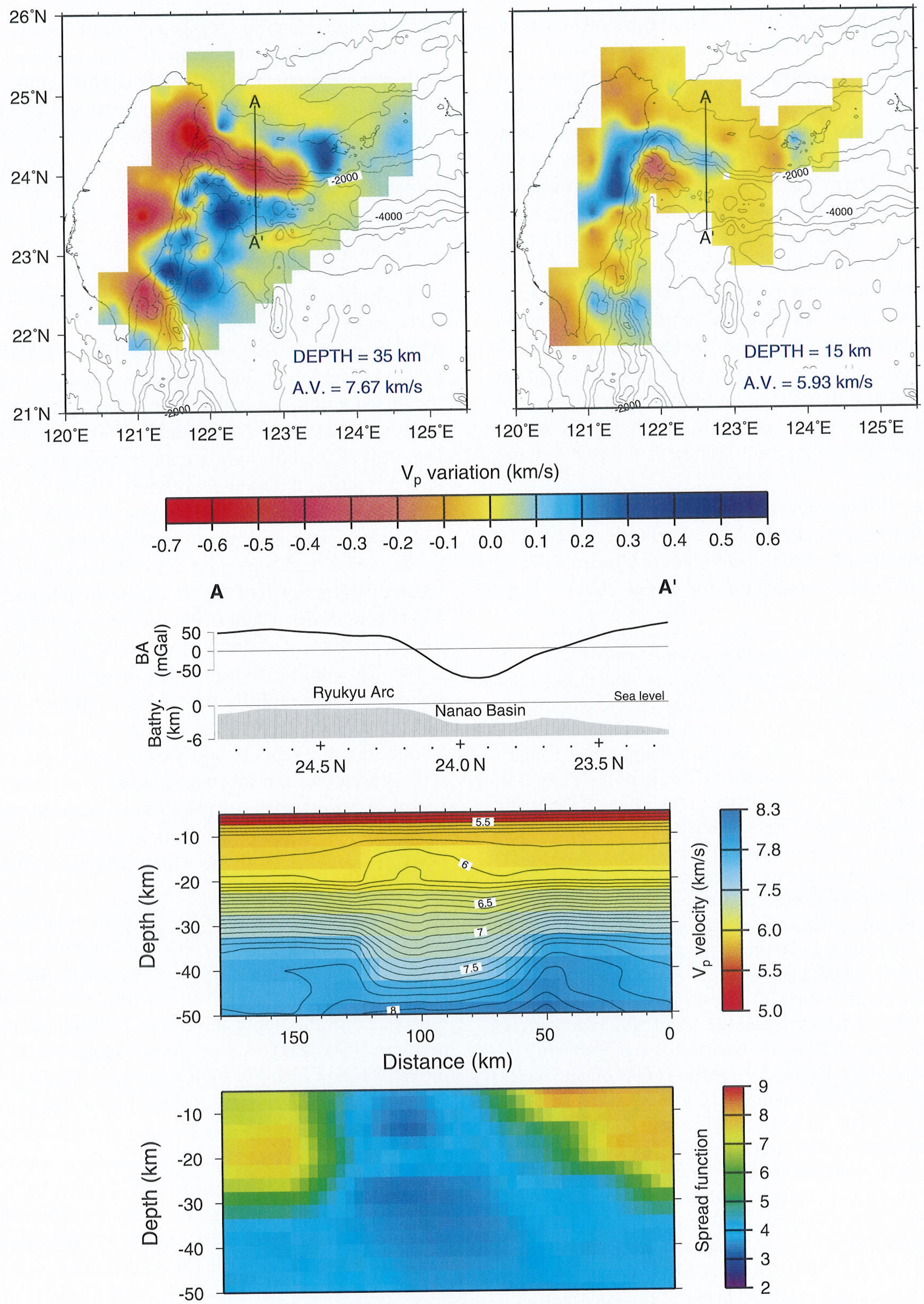


Plate 6. P-wave seismic tomography in the Taiwan-Ryukyu region adapted from Hsu [2001] showing the presence of a light crustal body parallel to the Nanao forearc basin. A.V., averaged velocity for the two top figures. This light body is interpreted as the relict of the upper portion of the Luzon Arc accreted against the Ryukyu forearc.

the Ryukyu forearc (profile 1) and collisional stage of the Luzon arc with EU in Taiwan, and subsequent shortening and uplift of the upper crust and overlying sediments initially belonging to the EU plate (profile 2). The Suao Basin (Figure 5), an offshore basin located 50 km northeast of Hualien, has been recently tilted and uplifted [*Font et al.*, 2001]. Lately, the depression created during the tilting of the Suao Basin was partly filled with sediments coming from the Lanyanghsi River through the Ilan Plain, giving rise to the Hoping Basin, lying unconformably on top of the Suao Basin [*Font et al.*, 2001]. We suggest that the uplift of the basement located beneath the tilted Suao Basin could be the result of the recent collision and accretion of the northernmost segment of the Luzon Arc.

3.6. Geodynamic Context of the Formation of Taiwan

Although the process of formation of proto-Taiwan is now quite well understood, the geological time constraints remain poor. Collision started at 12 Ma for Lu and Hsü [1992], 8 Ma for Chang and Chi [1983], Huang [1984] and Delcaillau et al. [1994], 6.5 Ma for Huang et al. [1997] and Lin and Watts [2002], 6-4 Ma for Barrier and Angelier [1986], 5-3 Ma ago for Teng [1996], 4 Ma for Suppe [1984] and even more recently for Malavieille et al. [2002]. On the northern OT margin, the youngest rock strata deformed by thrust faulting are late Miocene, with the collapse linked to OT opening occurring during the late Pliocene [*Hsiao et al.*, 1999; *Teng et al.*, 2000]. In the following geodynamic context, collision is assumed to start at 9 Ma (Figure 8), which seems reasonable in light of the 6.5 Ma age of emplacement of the foreland basin created by the Taiwan mountain load [*Lin and Watts*, 2002].

In the conceptual geodynamic context discussed below, we make the following assumptions:

- The PH/EU convergent rate in the N307° direction has been constant (5.6 cm/yr) since the beginning of collision (9 Ma).
- The western PH Sea plate boundary has moved continuously westward at a constant 4.5 cm/yr westward velocity with respect to EU. Simultaneously, the Manila Trench continuously moved with respect to EU, in the direction of the northern SCS margin (N250°).
- The main Taiwan collision occurred between Points A and B. Simultaneously, the EU continental lithosphere subducted eastward between the relict T1 and B. Since 9 Ma, the length of subducted EU continental lithosphere (length of the Coastal Range) has increased from 0 to 150 km.
- As soon as the northern portion of the Luzon Arc started to collide with the Ryukyu subduction zone, the upper part of the Luzon Arc accreted against the Ryukyu forearc and its lower part was subducted with the underlying PH Sea plate.
- As soon as the OT started to open, its western tip has moved westward with Point A.

3.6.1. Arc-continent collision at 9 Ma. About 15 Ma, the Manila Trench became active, extending northwards to the Ryukyu subduction zone. Since that time, the PH Sea plate, including the Ryukyu slab, the Luzon Arc and the PH Sea oceanic domain simultaneously moved westward with respect to EU.

From 15 to 9 Ma, extension did not occur in the OT. Point B was located at the base of the Ryukyu continental slope, in the southwestern prolongation of the active Ryukyu Trench. As soon as the EU subducted slab reached a depth of 100-150 km, the Luzon Arc started to form as an intra-oceanic arc. However, due to its minor initial relief, the Luzon Arc was at first subducted beneath EU as part of the PH Sea plate.

By 9 Ma, the topography of the Luzon Arc was sufficiently developed and buoyant to resist subduction and thus started to collide with the Eurasian margin near 126°E longitude.

3.6.2. Arc-continent collision at 6 Ma. Because the Luzon Arc continuously moved westward along the Chinese margin, the northern part of the already formed mountain belt (proto-Taiwan) progressively shifted to the eastern side of Point A, becoming an inactive part of the mountain chain. Since 9 Ma, the subduction of EU oceanic lithosphere continued south of Point B, simultaneous with the formation of the intra-oceanic Luzon Arc. Simultaneously, the eastward subduction of EU continental lithosphere occurred beneath the Luzon Arc between the relict Fault T1 and Point B features. Thrust faults in western Foothills were probably merging with a deep flat ramp beneath Taiwan. The western portion of the Ryukyu Arc and forearc started to slightly bend northward as a consequence of the subduction of an increased width of Eurasian continental lithosphere between the divergent relict Fault T1 and Point B. At the same time, the northern portion of the Luzon Arc was subducted, but with its upper part being accreted against the Ryukyu forearc.

3.6.3. Arc-continent collision at 3 Ma. The EU continental subduction process continued between the relict Fault T1 and Point B with the length of the Coastal Range increasing between. A continental accretionary wedge continued to develop due to thin-skinned tectonics. Since 6 Ma, the proto-Taiwan chain (northeast of Point A) increased in length but continuously subsided due to both erosional and extensional processes in the incipient OT backarc basin, which started to open at 6 Ma [*Kimura*, 1996; *Sibuet et al.*, 1998]. The western tip of the OT moved together with Point A. Because bathymetric data show a V-shape for the western termination of the OT [*Sibuet et al.*, 1998], we assumed that the westward motion of the PH Sea plate might have occurred at a constant velocity. The bending of the western Ryukyu Trench is now not only due to the oblique propagation of the tear fault inside the EU continental domain but also due to the simultaneous westward propagation and opening of the OT. As Taiwan uplift progressed, erosion increased, with some of the erosional products being deposited on the continental margin south of Point B and in the Manila Trench. These sediments were later incorporated into the Manila accretionary wedge as a consequence of the westward motion of the PH Sea plate with respect to EU.

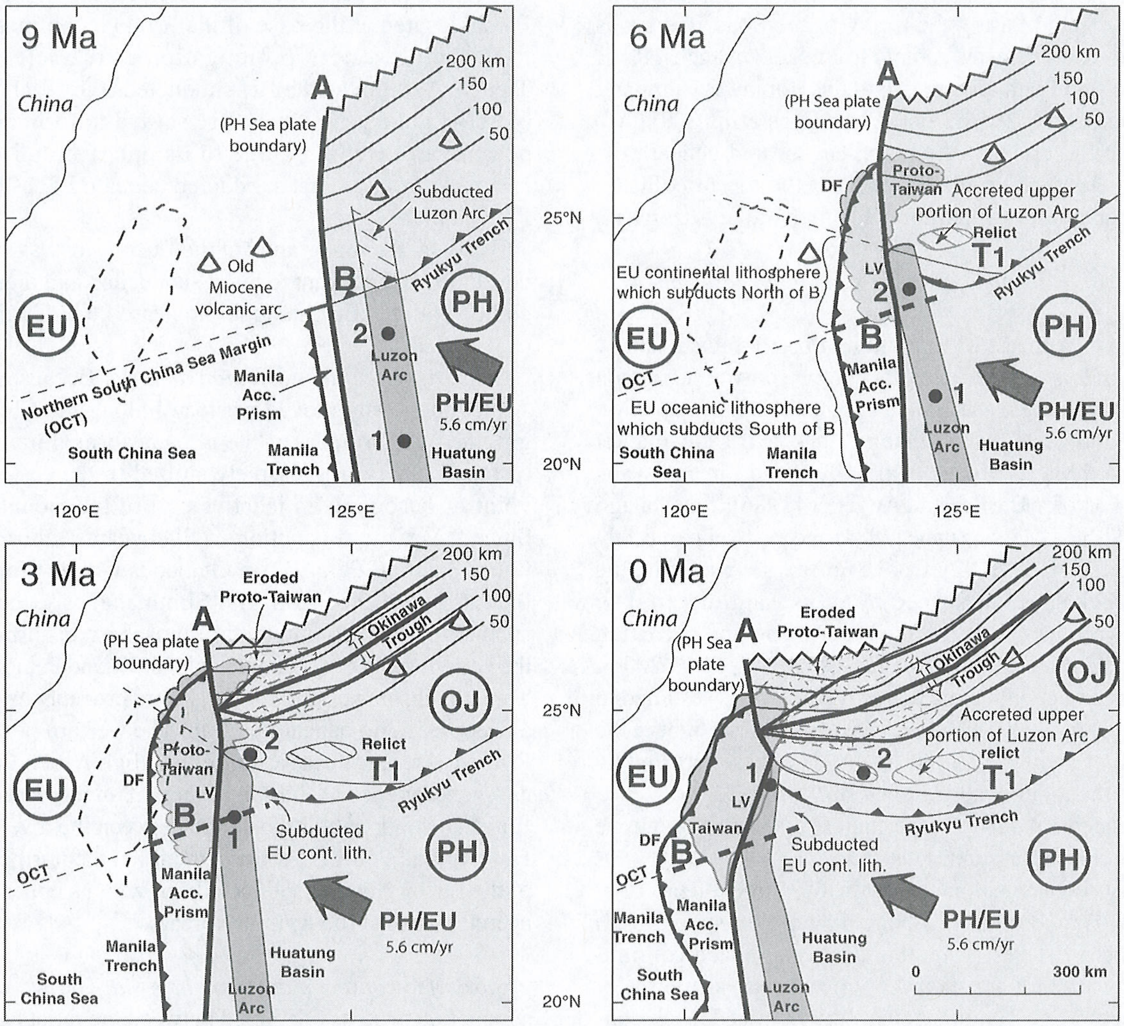


Figure 8. Sketch of the geodynamic evolution of Taiwan. Eurasia (EU) is fixed (adapted from Sibuet and Hsu [2004]). Point A is the western boundary of the Philippine (PH) Sea plate. Since 9 Ma, the PH Sea plate has moved in the N307° direction at a constant 5.6 cm/yr velocity. The uplift of proto-Taiwan is controlled by the westward motion of the PH Sea plate conveying the Luzon Arc. The westward displacement of the PH Sea slab with respect to EU triggers the development of a tear fault inside the EU lithosphere. East of Point A, EU is torn and has two boundaries (relict fault T1). B is the boundary between the oceanic and continental parts of the EU slab (in the prolongation of the South China Sea ocean-continent transition (OCT) zone). Thin continuous lines indicate depths of the Ryukyu slab (in km). Since 6 Ma, the beginning of Okinawa Trough (OT) opening, the OT tip is located above and move along with Point A, explaining the development of the southwestern OT within proto-Taiwan and the triangular shape of its western extremity. 1 and 2 are two fixed points of the Luzon Arc whose tracks can be followed through time. As soon as a portion of the Luzon Arc enters in the Ryukyu subduction zone, its upper part is accreted against the Ryukyu forearc while the lower part is subducted beneath EU. Volcanoes of the old Miocene volcanic arc located south of relict fault T1 have subducted simultaneously with the underlying EU continental lithosphere. DF, deformation front; LV, Longitudinal Valley; OJ, Okinawa-Japan plate. PH Sea plate in dark gray and Manila to northern Taiwan accretionary prism in light gray.

3.6.4. Present-day configuration. The width of the subducted EU continental lithosphere continued to increase between the relict Fault T1 and Point B. The proto-Taiwan chain has been severely eroded, although it may have partly subsided in response to OT opening. The backarc opening process enhanced the subsidence of the eastern part of the mountain chain, leaving only 50 km of the proto-Taiwan chain above sea-level (present-day portion of Taiwan located east of Point A) (Figure 5). The curvature of the western Ryukyu Trench is now more than 90° and the accreted portion of the Luzon Arc present in the Ryukyu forearc runs 400 km along strike. The amount of erosional products coming from Taiwan and deposited on the margin south of Point B and in the Manila Trench is such that 60 km of Taiwan was built south of Point B.

3.7. Discussion and Geodynamic Implications of the Proposed Model

The aim of this work was to present a conceptual model of the formation of Taiwan, not a detailed geological model of the emplacement and deformation of uplifted terranes. However, simple implications come from our model.

1) The notion of continuum from subduction to collision and finally to collapse might be taken with caution. Crustal strain shows a change from extension to compression in the EU crust on each side of Point A. However, compression occurs on both sides of Point B, even if the strain direction slightly changes at that place. In addition, the Southern Longitudinal Trough (Figure 5) is not the offshore equivalent of the Longitudinal Valley [*Malavieille et al.*, 2002], implying that geological features differ north and south of Point B. On land, if compression was considered by most field geologists to be continuous along the whole Central Range to south of the Hengchun peninsula, one exception was provided by Lin and Chen [1998] who suggested that the geometry of folds and thrusts differs on each side of a boundary located near 22.7°N latitude (our Point B). Our model suggests that the accretionary prism is continuous from the Philippines to northern Taiwan. However, the slight change in strain and geometry of folds and thrusts across Point B might be linked to the oceanic-to-continental change in nature of the underlying EU slab.

2) A second implication of our model concerns the way in which the southwestern OT formed. By simply looking at the triangular shape of the Ilan Plain, it has been already proposed that the OT propagated westward [*Sibuet and Hsu*, 1998; *Wang et al.*, 1999] but most of the authors suggest that the southwestern OT results from the OJ/EU rotation around a pole located near the tip of the Ilan Plain [*Lee et al.*, 1998; *Miki et al.*, 1990]. Miki et al. [1990] tried to avoid too much extension in the northern OT due to a location of the pole of rotation at the tip of the Ilan Plain, by decoupling the opening of the northern and southwestern OT, although no evidence for this has been advanced. In our model, the tip of the OT is always located above Point A and moves at a constant velocity with Point A during the formation of the entire OT (6 m.y.). This mechanism provides a simple explanation of the triangular shape of the Ilan Plain and southwestern OT by regular westward rifting propagation. In addition, our model explains the curved shapes of the southwestern Ryukyu Arc and forearc by two factors: the subduction of an increased width of EU continental lithosphere through time and the westward propagation of OT opening. In this fashion the 150-km bending of the western extremity of the Ryukyu subduction zone could be explained without requiring a pre-existing bayonet shape of the Eurasian margin, as initially proposed by Pelletier and Stéphan [1986].

3) What was the fate of the 150-km wide PH Sea oceanic domain located between the Manila Trench and the Luzon Arc? This crust does not exist west of the Coastal Range, although some may have been incorporated into the Lichi Mélange, a mixture of ophiolitic blocks and Miocene sediments formed about 15 Ma and uplifted during the Pliocene [*Jahn*, 1986; *Suppe et al.*, 1981]. For some authors, the basalts of the Lichi Mélange represent near-ridge seamount products whose origin was within the SCS oceanic crust [*Chung and Sun*, 1992; *Liou et al.*, 1977]. In contrast we suggest that the PH Sea oceanic crust located between the Manila Trench and the Luzon Arc was progressively incorporated into the EU plate as a consequence of the N307° PH/EU motion and then subducted north of Point B, beneath the Coastal Range, along with the adjacent EU continental crust. In this scheme, the subducted portion of this oceanic domain would lie as a stripe located between the continental and oceanic parts of the EU slab. Finally, the Lichi Mélange can be viewed as a peculiarity, which corresponds to the accretion against the Coastal Range (Luzon Arc) of a small piece of oceanic crust, which belonged to the Taiwan Sea, a former oceanic plate that was located between the PH Sea plate and the SCS [*Sibuet et al.*, 2002].

4) Our model addresses the issue of why there is no Miocene arc volcanism west 124°E. Because no arc volcanic rocks have been reported in the Yeyaema islands, most models infer that the Ryukyu subduction zone was only active east of 126°E during Miocene and earlier. West of 126°E, the margin is considered to have been passive (*e.g.* Lallemand et al. [2001]). In contrast, Sibuet et al. [1997; 2002] demonstrated that this portion of margin was active because, prior to 15 Ma, the Ryukyu subduction zone extended from Japan to west of Taiwan. Our model reconciles these two opposite interpretations. We suggest that the portion of EU continental crust, which subducted east-

ward between the relict Fault T1 and Point B included the portion of Ryukyu Miocene volcanic arc previously formed west of 126°E (Figure 8).

4. CONCLUSIONS

The main conclusions of this study are as follows:

1) Basins of the East and South China Sea continental shelves are located within four belts, which run parallel to the Chinese margin. Rifting occurred at the same time within the basins belonging to each of these four belts, and the belts become younger oceanwards. The four rifting phases occurred during Paleocene, Eocene, Oligocene to early Miocene and early middle Miocene to Present. The basins are separated by volcanic ridges that are the same age as the basins located immediately northwest. We interpret these basins and associated ridges as backarc basins and portions of relict arcs linked to the Ryukyu subduction zone. The Ryukyu subduction zone extended from Japan to southwest Taiwan until the early middle Miocene (about 15 Ma) where the southeast portion of the subduction zone facing the Tainan Basin and future Taiwan became inactive. Southwest of the Tainan Basin, the Pearl River Basins formed during the rifting of the northern SCS margin.

Subsidence curves show that rifting ceased 17–18 Ma in the west Taiwan Basins. The new shear wave velocity model suggests that the Ryukyu slab extended in the past southwest of Taiwan, beneath the northern China Sea margin. A deep seismic line shot across the northeastern South China Sea margin also suggests that this margin was active in the past. We conclude that about 15-20 Ma, the southwestern extremity of the Ryukyu subduction zone jumped from 118°E (southwest of the Tainan Basin) to 126°E (where the trend of the Ryukyu subduction zone changes direction). Since that time, the southwestern extremity of the Ryukyu subduction zone has moved progressively westwards to its present-day location at 122°E.

2) The plate kinematic reconstructions of East Asia since about 15 Ma takes into account the following constraints: spreading in the Taiwan Sea (northern Huatung Basin and oceanic domain beneath the Manila accretionary prism) was 2 cm/yr from chron 23 to 20 (51-43 Ma) and 1 cm/yr from chron 20 (43 Ma) to 5b (15 Ma); spreading ceased at the same time (about 15 Ma) in the Taiwan Sea, SCS, northeastern SCS, Sulu Sea, Japan Sea, and Shikoku and Parece Vela Basins; since about 15 Ma, the mean PH/EU velocity has been 5.6 cm/yr in a N307° direction.

3) Since 9 Ma, the formation of Taiwan has been controlled by the westward migration of the PH Sea plate into a tear in the fixed EU plate. The PH Sea plate boundary (Point A), which bounds the Ryukyu subduction zone to the west, corresponds to a diffuse crustal boundary feature which limits present-day extensional processes east of Point A (northern Taiwan and OT) from compressional processes west of Point A. The westward motion of the PH Sea slab triggered the development of a tear fault inside the Eurasian continent. East of Point A, EU is torn and thus has two boundaries. One is the upper plate, trench edge of the Ryukyu forearc; the other is the northern boundary of the deeply subducted EU slab underlying the PH Sea plate. Point B corresponds to the limit between the oceanic and continental portions of the EU slab. The intra-oceanic Luzon Arc formed south of Point B above the oceanic portion of the subducting Eurasian lithosphere. North of Point B, the Luzon Arc collided with EU, driving the subduction of a continental portion of Eurasian lithosphere. East of Point A, we suggest that the upper part of the Luzon Arc accreted against the Ryukyu forearc while its lower part subducted beneath EU.

4) One of the consequences of such a geodynamic context is that the widely accepted notion of continuum from subduction to collision and finally to collapse has to be questioned. The crustal strain changes from extension to compression across Point A might be linked to the presence of the underlying PH Sea slab edge. However, only faint strain and tectonic changes occur on each side of Point B, the ocean-continent transition of the EU slab. Furthermore, during the formation of the OT, its western tip moved simultaneously with Point A. The OT propagated westward in the proto-Taiwan orogen at the assumed 4.5 cm/yr westward velocity, giving rise to the triangular shape of its southwestern extremity. The bending of the southwestern Ryukyu subduction zone may be due to the combination of two processes: the subduction of a portion of Eurasian continental lithosphere whose width increased through time and the westward propagation of the OT backarc basin. Miocene arc volcanic rocks were never found in the southwestern part of the Ryukyu Arc. We suggest that the portion of EU continental lithosphere, which subducted south of relict T1, included the portion of Miocene Ryukyu volcanic arc previously formed west of 126°E longitude.

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