

HOW WAS TAIWAN CREATED?

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Abstract. Since the beginning of formation of proto-Taiwan during late Miocene (9 Ma), the subducting Philippine (PH) Sea plate moved continuously through time in the N307° direction at a 5.6 cm/yr velocity with respect to Eurasia (EU), tearing the Eurasian plate. Strain states within the EU crust are different on each side of the western PH Sea plate boundary (extensional in the Okinawa Trough and northeastern Taiwan versus contractional for the rest of Taiwan Island). The B feature corresponds to the boundary between the continental and oceanic parts of the subducting Eurasian plate and lies in the prolongation of the ocean-continent boundary of the northern South China Sea. Strain rates in the Philippines to northern Taiwan accretionary prism are similar on each side of B (contractional), though with different strain directions, perhaps in relation with the change of nature of the EU slab across B. Consequently, in the process of Taiwan mountain building, the deformation style was probably not changing continuously from the Manila to the Ryukyu subduction zones. The Luzon intra-oceanic arc only formed south of B, above the subducting Eurasian oceanic lithosphere. North of B, the Luzon arc collided with EU simultaneously with the eastward subduction of a portion of EU continental lithosphere beneath the Luzon arc. In its northern portion, the lower part of the Luzon arc was subducting beneath Eurasia while the upper part accreted against the Ryukyu forearc. Among the consequences of such a simple geodynamic model: i) The notion of continuum from subduction to collision might be questioned. ii) Traces of the Miocene volcanic arc were never found in the southwestern Ryukyu arc. We suggest that the portion of EU continental lithosphere which has subducted beneath the Coastal Range might include the Miocene Ryukyu arc volcanoes formed west of 126°E longitude and which are missing today. iii) The 150-km wide oceanic domain located south of B between the Luzon arc and the Manila trench, above the subducting oceanic EU plate (South China Sea) was progressively incorporated into the EU plate north of B.

Keywords: Taiwan, kinematics, tectonics, seismicity, tomography, geodynamic model

1. Introduction

The Taiwan mountain belt is one of the youngest orogenies in the world. Whatever are the proposed models to explain the formation of Taiwan, all authors agree that the uplift of Taiwan results from the collision of the Luzon arc with the Eurasian (EU) margin. The Luzon arc is an intra-oceanic volcanic arc which belongs to the Philippine (PH) Sea plate, outcrops in the Coastal Range and whose contact with the EU margin is the Longitudinal Valley. Numerous discussions are relative to the nature of the collided margin: sedimentary monocline belonging to the EU plate and deformed by the collision of the Luzon arc (Suppe, 1981; Suppe, 1984), lithospheric collision between the Luzon arc and the EU lithosphere (Wu et al., 1997), exotic block accreted against the Eurasian margin (Lu and Hsü, 1992) or previously formed Ryukyu arc and backarc basin system (Hsu and Sibuet, 1995; Sibuet and Hsu, 1997) subsequently deformed by the Luzon arc collision. Davis et al. (1983) and Suppe (1984) demonstrate 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 Ma earlier. Thus, most authors (*e.g.* Teng (1990) and Malavieille et al. (2002)) 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). A major issue in the tectonics of Taiwan is still to define the basic local plate kinematic configuration, which is partly hidden by extensive deformation. In order to understand the geodynamic context of Taiwan, we start from the recent East Asia plate kinematic reconstructions established by Sibuet et al. (2002) since 15 Ma. Then, we will define the location of the western boundary of the PH Sea plate (A) which bounds the 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 and then east of the Manila accretionary prism. The PH Sea plate is moving westward through time with respect to the EU plate. In northern Taiwan, there is no surface feature associated with the western end of the Ryukyu slab but geological provinces lying on both sides of A are characterized by different stress patterns. The purpose of this paper is to understand the process of Taiwan mountain building within a simple plate tectonic context which includes a comprehensive history of the westward migration of the collision. For that purpose, we first need to 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. Geological Setting

Throughout the whole paper, plate motions are relative to EU supposed to be fixed. Latitudes and longitudes are also given in the EU frame. About 15 Ma ago, a major plate reorganization occurred in East Asia (Sibuet et al., 2002). Seafloor spreading ceased in the South China Sea (SCS) (Briais et al., 1993), Japan Sea (Jolivet et al., 1994), Sulu Sea (Rangin and Silver, 1991), and Shikoku and Parace Vela basins (Chamot-Rooke et al., 1987; Okino et al., 1998; Okino et al., 1994). According to Sibuet et al. (2002), the complex system of a dozen of plates suddenly evolved to a simple three-plate system (Eurasia (EU), Philippine (PH) Sea and Pacific (PA)) which continued until today (Figure 1). Since 8 Ma, the PH/EU pole of rotation is located northeast of Japan and we have adopted for this period the present-day PH/EU motion determined by Seno et al. (1993) (48.2°N , 152°E , $-1.09^{\circ}/\text{Ma}$). In the Taiwan area, the PH/EU direction of convergence and velocity given by Seno et al. are $\text{N}307^{\circ}$ and 7.1 cm/yr , respectively. Since 8 Ma, the PH Sea plate rotated clockwise about 9° with respect to EU (Hall, 2002). GPS data acquired in the Taiwan area show that the present-day PH/EU motion is 8 cm/yr (Yu et al., 1997), close to Seno et al. velocity. However, a mean 5.6 cm/yr convergent velocity during the last 8 Ma was suggested (Sibuet et al., 2002).

15 Ma ago, the vergence and location of the subduction zone shifted from east to west of the Philippine archipelago (Maletierre, 1989; Wolfe, 1981), giving rise to the Manila subduction zone. Later on, the Luzon arc started to form as an intra-oceanic arc. At the beginning, 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 is only 9 to 6 Ma ago that the young Luzon arc acquired a significant topographic expression to resist 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). On land, 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. The proto-Taiwan chain was located at the position of the present-day southwestern Okinawa Trough and adjacent northern continental shelf (Hsiao et al., 1999; Sibuet et al., 1998). Its eastern boundary was carefully mapped (Hsiao et al., 1999). In this scheme, the eastern limit of thrust faults is located in the $\text{N}307^{\circ}$ direction with respect to the location of the first impact of the Luzon arc collision relative to the EU plate, about 400 km east of Taiwan (Figure 2). The impact point is located where the Ryukyu trench changes direction from NE-SW to E-W. From the PH/EU kinematic parameters, the mean westward motion of the Luzon arc is 4.5 cm/yr for the last 8 Ma and the colliding point of the Luzon arc with EU continuously migrated westwards for a total amount of 400 km (Figures 1 and 2).

3. Location of the Western Philippine Sea Plate Boundary (A)

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 3). The boundary of the PH Sea plate follows the subducted plate edge. If 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 high velocity anomaly on tomographic images. Obviously, 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, A is not well constrained in northern Taiwan but corresponds to the western boundary of the deep seismicity (Font et al., 1999) (Figure 3) and extends approximately from Hualien to somewhere between Chungli and the Taitun volcano (Figure 4).

Northeast of 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. The subducted PH Sea plate is in downdip extension between 80 and 120 km and in downdip compression at 270 km (Kao et al., 1998). Above the Benioff zone, within the EU plate, shallow crustal extensional earthquakes occur in the southwestern Okinawa Trough and Ilan Plain. On land, several active faults with a normal component have been identified (Figure 4). 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 Taitun volcano is cut by a set of Quaternary normal and strike-slip faults (Lee and Wang, 1988) and the modern Chinshan fault is a series of normal faults with a strike-slip component (Chen and Yeh, 1991; Lee and Wang, 1988).

West of A, earthquakes show a consistent orientation of P-axes in the direction of collision, with crustal thrust faulting mechanisms mostly in the N290° direction, except near Hualien, where thrust faulting mechanisms change direction to N330° (Kao et al., 1998). 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. According to the above observations, there are only compressive earthquakes mechanisms and reverse faults which testify to solely contractional motions within the upper plate, west of A.

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

South of Hualien, the Philippine Sea plate boundary follows the Longitudinal Valley (Figure 4), as suggested by numerous authors (*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 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 and possibly upper crust material belonging to the EU plate (South China Sea), the rest of the EU plate subducting beneath the PH Sea plate. The Manila trench is prolonged 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. To summarize, a major accretionary prism have been identified west of the boundary of the rigid PH Sea plate and consists of deformed sediments and upper crust previously belonging to the EU plate, whose nature is continental in the north and oceanic in the south.

4. 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 could be followed down to 500-600 km (Rangin et al., 1999). However, the deep seismicity associated with the Manila slab occurs south of B (Figure 3) and is almost absent north of B. 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 B, as well as an increase of the slab steepness north of B. South of B, the distribution of seismicity at depths >80 km represents the geometry of the Manila subducted slab (Figure 3). Earthquake focal mechanisms show typical subduction related features: normal faulting earthquakes near the Manila trench axis, low-angle thrust events over depths between 15 and 35 km underlining the seismogenic portion of the plate interface. The subducted EU plate is in downdip extension between 100 and 150 km and in downdip compression between 150 and 200 km (Kao et al., 2000). South of B, there is a good correspondence 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 (Figure 3) suggests that the portion of the slab imaged on tomographic data might be detached (Lallemand et al., 2001). B is approximately located in the northeastward prolongation of the northern continental margin of the South China Sea, suggesting that the slab might be of continental and oceanic nature north and south of B, respectively. In other words, within the EU slab, B would follow the ocean-continent transition and might underline the southern limit of the possible northward detached portion of the slab. The considerable intraplate deformation occurring north of 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 B exists offshore, south of the Coastal Range (Figure 3). 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 melange interpreted as a syntectonic melange formed during the collision of the Luzon arc with EU. 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 melange was deposited earlier to the north and the Huatung Ridge consists of folded and faulted sediments belonging to this basin.

In conclusion, B lies in the prolongation of the base of the northeastern South China Sea continental margin. We suggest that the slab might be of different nature (continental versus oceanic) north and south of B, the Luzon intra-oceanic arc being formed only south of 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. The cartoon of Figure 5 shows a 3-D view of the EU and PH Sea plates in south Taiwan. In the next section, we will discuss the nature and geographical extension of the portion of subducted slab north of B.

5. 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 B? In the Pelletier and Stéphan (1986) hypothesis, a bayonet shape of the margin was required to initiate the subduction of a piece of oceanic lithosphere north of B. In this hypothesis, as long as the oceanic lithosphere is subducting, the proto-collision in Taiwan would have to be postponed. 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, uplifted to form the Taiwan orogen (*e.g.* Suppe et al. (1981), Brusset et al. (1999)) and then eroded. Several alternatives to this model have been proposed: slab breakoff 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 a buoyant continental lithosphere as a consequence of several factors as suggested by the subduction of the Indian plate beneath the Himalaya: dragging of the continental lithosphere by already subducted oceanic lithosphere, eclogitization process occurring at 70-90 km depth 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 Okinawa Trough opening) is located along the same regular trend that the one of the northeastern South China Sea margin, suggesting a smoothed shape

for the Eurasian margin before collision (Figure 2). In this hypothesis, it seems difficult to infer, north of B, the presence of a large piece of oceanic lithosphere.

Assuming that the EU continental lithosphere started to subduct at the onset of the Luzon arc 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 entered in collision with the EU margin 9 Ma ago, 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 moved westward at a 4.5 cm/yr mean velocity (Sibuet et al., 2002). If the width of the subducted continental lithosphere regularly increased from a zero-length 9 Ma ago to 150 km today (the present-day length of the Coastal Range), the whole surface of subducted continental lithosphere would be 30,000 km².

2) Using tomographic results, Lallemand et al. (2001) have restored the horizontal shape of the unfolded PH Sea and EU (South China Sea) slabs. To explain 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 portion of subducted continental crust presents a 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 km².

3) Figure 2 shows a simplified bathymetric map of the Taiwan area with: i) the approximate boundary of the present-day EU continental crust supposed of constant thickness (3000-m isobath east of Taiwan and 2000-m isobath west of Taiwan because this margin is highly sedimented (present-day EU crust boundary in Figure 2)); ii) the restored position of the EU continental crust boundary east of Taiwan (EU crust boundary before Okinawa Trough opening in Figure 2) before the Okinawa Trough 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 Okinawa Trough (Sibuet et al., 1995)); iii) the EU continental crust boundary (EU crust boundary before collision in Figure 2) before collision of the Luzon arc and Okinawa Trough (OT) opening (6 Ma). The gray area in Figure 2 corresponds to the surface of the EU continental crust of constant thickness subducted since the onset of the Luzon arc collision (50,000 km²). This estimate differs from previous ones because of the uncertainty in the position of the ocean-continent boundary (OCT) before the OT opening and the width of the subducted continental lithosphere, which probably increased rapidly during the first 3 Ma and less during the last 6 Ma. 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 to 50,000 km² 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 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 6 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.

6. Present-day Plate Kinematics

In the Taiwan area, the location of major plate boundaries were always a matter of debate. A scheme of the present-day plate kinematic system is presented in the three panels of Figure 7, each panel corresponding to the geographical extent of the PH Sea, EU and Okinawa-Japan (OJ) rigid plates, respectively. Globally, the overall plate motion corresponds to the westward motion of the PH Sea plate inside the teared EU plate supposed to be fixed. As backarc extension occurred in the Okinawa Trough, an additional plate (OJ) is required. The simplicity of the three-plate pattern geometry was never properly understood in the past because of the presence of accretionary prisms resulting from the deformation of scraped sediments and upper crustal material previously belonging to the EU plate in western Taiwan and to the PH Sea plate in the Ryukyu accretionary prism.

Figure 7 shows the distribution of crustal deformation established 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)). In northern Taiwan, in the Okinawa Trough as well as in the Manila accretionary prism, areas of diffuse extension are shown. Areas where compressive deformation occurs or is particularly intense (Northern Luzon Trough, Southern Longitudinal Trough, Huatung Ridge, Coastal Range and Luzon arc) are also underlined.

The development of a tear fault within the EU lithosphere is triggered by the western motion of the PH Sea plate with respect to EU through time (Figures 7 and 8). East of A, the edge of EU plate is torn and thus has two boundaries (relict T1 in Figure 6). 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. West of A, the tear fault will develop within the EU plate, but not along A which is continuously moving westward with respect to EU.

Figures 6 and 7 shows the configuration of EU and PH Sea slabs. For clarity, Figure 7 is divided into three panels. Figure 7a shows in blue the extension of the PH Sea plate, which partly overlies the EU plate and is overlain by the OJ plate. The accretionary prism, with its oceanic (Manila) and continental (Taiwan uplift) portions adjacent along B, appears in light blue. It cannot be considered as a rigid plate or part of the rigid PH Sea plate because internal

deformations occur inside the accretionary prism. Figure 7b shows the extension of the EU plate (in yellow for the continental part and green for the oceanic part). The EU plate partially overlies the PH Sea plate. Figure 7c shows in orange the extension of the OJ plate. 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 4). It 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 accretionary wedge is moving laterally along this transcurrent fault identified from 122°E to 124°E longitude and located at the toe of the rigid arc and forearc backstop (relict T1 in Figure 7c). Recently published swath bathymetric data east of 124°E longitude (Matsumoto et al., 2001) suggest that the transcurrent fault is likely extending eastward to 126°E, merging the Ryukyu trench where the trench changes direction (Figure 2). We assume that the relict tear fault T1 extends from 122°E to 126°E longitude.

Figure 8 shows synthetic profiles across each of the three main domains (east of A, north and south of B) and from the Okinawa Trough to the Manila accretionary prism (locations in Figure 7). The nature and geometry of the EU slab differ on each side of B: South of B, the EU slab is oceanic (Figure 8, profile 3) and the Manila accretionary prism consists of crust (?) and sediments scraped off the EU plate but also coming from the erosion of Taiwan. North of B, the EU slab is continental (Figure 8, profile 2). Thrust faults in western Foothills merge a 10-km deep flat main detachment beneath Taiwan which dives down to 30-60 km 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 (Figure 8, profile 2). The Longitudinal Valley is the contact between the EU and PH Sea plates. The deep portion of the EU slab is supposed to be detached east of the Coastal Range (Figure 8b) (Carena et al., 2002; Lallemand et al., 2001). Is the slab completely detached between relict T1 and B as suggested by Lallemand et al. (2001) or partially torn? It is a pending question with a consequence on the type of motion within the EU slab, along its former OCT. In figure 6, we have adopted Lallemand et al. (2001) suggestion. As B is a significant boundary within the EU slab, the notion of continuum from subduction to collision has to be raised, even if they are only faint differences in the surface geology across it.

7. 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 (Figure 4). Except for some authors (Cheng et al., 1996; Wang and Chiang, 1998), the Luzon arc does not exist north of Hualien. For others, the northern prolongation of the Coastal Range subducted beneath Taiwan (*e.g.* Chemenda et al. (1997)). In

this paper, we suggest that the upper portion of the northern Luzon arc 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 late Miocene near 126°E longitude with respect to EU, where the present-day Ryukyu trench changes direction. Since that time, the northward Luzon arc motion with respect to EU was about 400 km and the colliding point of the Luzon arc with EU moved through time along the Ryukyu margin from 126°E to 121.5°E longitude. The surrection of proto-Taiwan is the consequence of the Luzon arc collision with the Eurasian margin (Hsiao et al., 1999). Huang et al. (1992) and Hsiao et al. (1999) have 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 (124°E) was lying approximately in the N307° prolongation of the initial Luzon arc colliding point with EU (126°E) (Figure 2).

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 as a consequence of the westward propagation of the Okinawa Trough backarc basin within proto-Taiwan (Sibuet et al., 1998). If the geometry of the Luzon arc collision with EU is quite well understood through time, the remaining question is: what is the fate of the 400-km long portion of Luzon arc, which have disappeared since 9 M.y. and was never identified north of the Coastal Range?

A decoupled-wedge structure has been recently imaged on a deep multi-channel seismic profile shot across the Hidaka collision zone which was created by the convergent motion of the Kuril arc with respect to the Japan arc in Central Hokkaido (Tsumura et al., 1999). Numerous east-dipping reflectors are observed in the upper part of the Kuril lower crust at a depth of 14 to 23 km, while west-dipping reflectors are observed in the lower part of the lower crust at a depth of 23 to 33 km. 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 Japan arc while the lower half (lower portion of the lower crust (6.6 km/s) and the upper mantle) descends (Ito et al., 2000). The geometry of reflectors is thus consistent with the decoupling of the lower crust of the colliding Kuril arc at a depth of 23 km during the active collision with the Japan arc (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.

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 evidenced east of Taiwan beneath the upper Ryukyu forearc slope and the Nanao Basin (Figure 9), but not further east by lack of data. This feature is about 50 km wide and is located at a depth comprised between 15 and 45 km. Its thickness is similar to the one of the accreted portion of the Hidaka collision zone. 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 for the Hidaka collision zone.

A low-velocity zone (5.5 km/s with respect to 7.0 km/s in the adjacent terranes) has been also evidenced at the same location on a N-S wide angle and refraction profile shot at 122°30'E longitude. It has been identified within the upper plate, beneath the upper Ryukyu forearc slope and the Nanao basin (Wang et al., 2002). Though this light body is less well located on tomographic results, both tomographic and refraction data suggest the existence of a light body beneath the Nanao basin and upper Ryukyu forearc slope.

An elongated -180 mGal free-air gravity anomaly low (Hsu et al., 1998) exists in this region. Its minimum axis is located 10-km north of the Nanao basin axis and extends to 125°E longitude. Forward modelling suggests that the gravity low might be associated both with the 10-25 km deep light body (accreted upper part of the Luzon arc against the Ryukyu forearc) and the overlying Nanao forearc basin, a topographic depression partly filled with sediments, and located seaward of the upper Luzon arc accreted body and north of the Ryukyu accretionary wedge. Thus, both the Nanao basin and the underlying light body are parallel features which disappear to the east at 125°E longitude.

In summary, we propose to apply the lithospheric models 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 of the Luzon arc, as shown 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 the Ryukyu subduction zone, the upper part of the arc (25-30 km), which is less dense than the adjacent continental crust, accreted against the Ryukyu forearc and the lower part of the arc subducted beneath Eurasia simultaneously with the underlying PH Sea oceanic crust. The Nanao basin is considered there as a partly filled depression created seaward of the upper Luzon arc accreted body and north of the Ryukyu accretionary prism.

Profiles 1 and 2 (Figure 8) show the difference in fate of the Luzon arc: accretion of the upper part of the Luzon arc against 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 4), an offshore basin located 50 km northeast of Hualien, have 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 rise located beneath the tilted Suao basin could be the result of the recent collision and accretion of the last segment of the Luzon arc.

8. Formation of Taiwan: a Geodynamic Model

If the process of formation of proto-Taiwan is now quite well understood, unfortunately geological time constraints are poor. Collision started 12 Ma ago for Lu and Hsü (1992), 8 Ma ago for Chang and Chi (1983), Huang (1984) and Delcaillau et al. (1994), 6.5 Ma ago for Huang et al. (1997), 6-4 Ma ago for Barrier and Angelier (1986), 5-3 Ma ago for Teng (1996), 4 Ma ago for Suppe (1984) and even more recently for Malavieille et al. (2002). On the northern Okinawa Trough margin, the youngest rock strata deformed by thrust faulting are late Miocene, with the collapse linked to the OT opening occurring during the late Pliocene (Hsiao et al., 1999; Teng et al., 2000). In the following geodynamic model, collision is assumed to start 9 Ma ago (Figure 10). A 5.6 cm/yr PH/EU convergence rate is assumed to be constant since that time in the N307° direction (4.5 cm/yr in the E-W direction) (Sibuet et al., 2002). However, if the beginning of collision had occurred more recently (*e.g.* 6 Ma), the convergent rate would have been higher. In that case, the four reconstructions would have corresponded to 6 Ma, 4 Ma, 2 Ma and Present-day, with a mean convergent rate close to the present-day one (7.1 cm/yr) since 6 Ma.

In the geodynamic model discussed below, we make the following assumptions:

- The PH/EU convergent rate in the N307° direction is constant (5.6 cm/yr) since the beginning of collision (9 Ma). The westward component of this motion is 4.5 cm/yr.
- The western PH Sea plate boundary A continuously moved westward at a constant 4.5 cm/yr westward velocity with respect to EU. Simultaneously, the Manila subduction continuously moved with respect to EU, in the direction of the northern South China Sea margin (N250°).
- The main Taiwan collision occurred between 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) increased from 0 to 150 km.
- As soon as the northern portion of the Luzon arc entered in collision with the Ryukyu subduction zone, the upper part of the Luzon arc accreted against the Ryukyu forearc and its lower part subducted with the underlying PH Sea plate.
- As soon as the Okinawa Trough started to open, its western tip simultaneously moved westward with A.

8.1. 9 Ma Reconstruction (Figure 10)

About 15 Ma ago, 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 Eurasia.

From 15 to 9 Ma, extension did not occur in the Okinawa Trough. B was located at the base of the Ryukyu continental slope, in the southwestern prolongation of the 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. Due to its minor initial relief, the Luzon arc subducted beneath EU as part of the PH Sea plate.

9 Ma ago, the topography of the Luzon arc was sufficiently developed and buoyant to resist subduction and started to collide with the Eurasian margin near 126°E longitude with respect to EU.

8.2. 6 Ma Reconstruction (Figure 10)

Because the Luzon arc continuously moved westward, the northern part of the already formed mountain belt (proto-Taiwan) progressively shifted on the eastern side of A, becoming an inactive part of the mountain chain. Since 9 Ma, the subduction of EU oceanic lithosphere continued south of B, simultaneously with the formation of the intra-oceanic Luzon arc while the eastward subduction of the EU continental lithosphere occurred beneath the Luzon arc between the relict T1 and B features. Thrust faults in western Foothills were probably merging 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 T1 and B features. Simultaneously, the northern portion of the Luzon arc subducted with its upper part being accreted against the Ryukyu forearc.

8.3. 3 Ma Reconstruction (Figure 10)

The EU continental subduction process continued between the relict T1 and B with an increasing length of the Coastal Range in between. The continental accretionary wedge continued to develop by thin-skinned tectonics. Since 6 Ma, the proto-Taiwan chain (northeast of A) increased in length but continuously subsided due to both erosional and extensional processes in the incipient Okinawa Trough backarc basin, which started to open 6 Ma ago (Kimura, 1996; Sibuet et al., 1998). The western tip of the Okinawa Trough was moving simultaneously with A. Because bathymetric data show a V-shape for the western termination of the Okinawa Trough (Sibuet et al., 1998), the westward motion of the PH Sea plate might have occurred at a constant velocity, as previously assumed. 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 to the simultaneous westward propagation and opening of the Okinawa Trough. As Taiwan uplift progressed, erosion increased and part of the products of erosion deposited on the continental margin south of B and in the Manila trench. These sediments were incorporated later on into the Manila subduction system, folded and uplifted in the accretionary wedge as a consequence of the westward motion of the PH Sea plate with respect to Eurasia.

8.4. Present-day Configuration (Figure 10)

The width of the subducted EU continental lithosphere continued to increase between the relict T1 and B. The proto-Taiwan chain was severely eroded, though partly subsiding in response to the Okinawa Trough opening. The backarc opening process increased the disappearance 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 A) (Figure 4). The bending of the western Ruykyu trench is now more than 90° and the accreted portion of the Luzon arc against the Ruykyu arc and forearc reaches 400 km. The amount of erosional products coming from Taiwan and deposited on the margin south of B and in the Manila trench is such that 60 km of Taiwan Island was built south of B, and additional 200 km of sediments mostly originated from the erosion of Taiwan were deformed south of Taiwan, in the Manila accretionary wedge.

9. 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 this model.

1) The notion of continuum from subduction to collision and finally to collapse might be taken with caution. The crustal state of strain shows a reasonable tectonic change from extension to compression in the EU crust on each side of the PH Sea plate boundary A. However, compression occurs on both sides of B, even if the strain direction slightly changes across B. In addition, the Southern Longitudinal Trough is not the offshore equivalent of the Longitudinal Valley (Malavieille et al., 2002) implying that geological features differ north and south of B. On land, if compression was considered by most field geologists to be continuous along the whole Central Range to south to the Hengchun peninsula, one exception was provided by Lin and Chen (1998) who suggested that the geometry of folds and thrusts was different on each side of a boundary located near 22.7°N latitude (our B boundary). 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 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 Okinawa Trough formed. By simply looking at the triangular shape of the Ilan Plain, it has been already proposed that the Okinawa Trough propagated westward (Sibuet and Hsu, 1998; Wang et al., 1999) but most of the authors suggest that the southwestern Okinawa Trough 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). In Miki et al. (1990) model, to avoid a too large amount of extension in the northern Okinawa Trough due to a location of the pole of rotation at the tip of the Ilan plain, a decoupling of the opening between northern and southwestern Okinawa Trough was inferred but never evidenced. In our model, the tip of the Okinawa Trough is

always located above A and moves at a constant velocity with A during the formation of the entire Okinawa Trough (6 Ma). This mechanism provides a simple explanation of the triangular shape of the Ilan Plain and southwestern Okinawa Trough 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 the Okinawa Trough opening. By this way, 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 is the fate of the 150-km wide PH Sea oceanic domain located between the Manila trench and the Luzon arc? It does not exist west of the Coastal Range except for the Lichi melange, a mixture of ophiolitic blocks and Miocene sediments which was formed about 15 Ma ago and uplifted during Pliocene (Jahn, 1986; Suppe et al., 1981). For some authors, the basalts of the Lichi melange represent near-ridge seamount products whose origin was related to the South China Sea oceanic crust (Chung and Sun, 1992; Liou et al., 1977). 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 B, beneath the Coastal Range, simultaneously 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 melange is a peculiarity which corresponds to the accretion against the Coastal Range of a small piece of oceanic domain which belongs to the Taiwan Sea, a former oceanic plate which was located between the PH Sea plate and the South China Sea (Sibuet et al., 2002).

4) Why there is no Miocene arc volcanic activity west of 124°E longitude? As no arc volcanic rocks were reported in the Yeyaema islands, current interpretations infer that the Ryukyu subduction zone was only active east of 125°E during Miocene and earlier. West of 125°E, the margin was considered as passive (*e.g.* Lallemand et al. (2001)). On the contrary, Sibuet et al. (1997; 2002) demonstrated that prior to 15 Ma the Ryukyu subduction zone extended from Japan to west of Taiwan. A deep seismic profile shot west of Taiwan, across the northeasternmost South China Sea margin, shows that the margin was an active margin before 15 Ma (Sibuet et al., 2002). Our model reconciles these two opposite interpretations. We suggest that the portion of EU continental crust which subducted eastward between the relict T1 and B included the portion of Ryukyu Miocene volcanic arc previously formed west of 126°E (Figure 10).

10. Conclusions

The main conclusions of this study are as follows:

1. Since 9 Ma, the formation of Taiwan is controlled by the westward migration of the PH Sea plate within the teared EU plate supposed to be fixed.

2. The PH Sea plate boundary A, which bounds to the west the Ryukyu subduction zone, corresponds in the overlying terranes to a diffuse crustal boundary feature which limits present-day extensional processes east of A (northern Taiwan and Okinawa Trough) from contractional processes west of A. The westward motion of the PH Sea slab triggered the development of a tear fault inside the Eurasian continent. East of A, EU is torn and thus has two boundaries (relict T1). 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.

3. B corresponds to the limit between the oceanic and continental portions of the EU slab. The intra-oceanic Luzon arc formed south of B above the oceanic portion of the subducting Eurasian lithosphere. North of B, the already formed Luzon arc collided with Eurasia, simultaneously with the subduction of a continental portion of Eurasian lithosphere. East of A, we suggest that the upper part of the Luzon arc accreted against the Ryukyu forearc while its lower part subducted beneath Eurasia.

4. The consequences of such a geodynamic model are:

- The widely accepted notion of continuum from subduction to collision and finally to collapse might be taken with caution. The crustal strain changes from extension to compression across 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 B, the ocean-continent boundary of the EU slab.

- During the formation of the Okinawa Trough, its western tip moved simultaneously with A. The Okinawa Trough 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 is due to the combination of two factors: the subduction of a portion of Eurasian continental lithosphere whose width increased through time and the westward propagation of the Okinawa Trough 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.

- The 150-km wide oceanic domain located south of B between the Luzon arc and the Manila trench was progressively incorporated into the Eurasian plate north of B and then subducted beneath the Coastal Range simultaneously with the adjacent EU continental lithosphere.

- The collision of the Luzon arc with EU triggered the development of a tear fault within the EU lithosphere and the initiation of continental subduction. Today, the slab pull force exerted by the continental subduction might be the fundamental mechanism giving rise to the Taiwan uplift, mostly by thinned-skinned tectonics.

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Legend of figures

Figure 1: Plate kinematic reconstructions with respect to EU supposed to be fixed, of the East Asia region since 8 Ma after Sibuet et al. (2002), showing at a regional scale the westward motion through time (4.5 cm/yr) of both the Luzon arc and the Manila subduction zone. Plate name abbreviations: EU, Eurasia plate; PA, Pacific plate; PH, Philippine Sea plate. Light gray, continental domains; white, oceanic domains; dark gray, intra-oceanic arcs; thick black lines, active plate boundaries; large black arrow, PH/EU motion. Fracture zones, main magnetic lineations and their identifications are shown. The Taiwan Sea (TS) is located between the Luzon-Ryukyu transform plate boundary (LRTPB) and the Gagua ridge (GR). Signification of symbols: C, Celebes Sea; J, Japan; L, Luzon island; LA, Luzon arc; MT, Manila trench; OT, Okinawa Trough; P, Palawan; PB, Philippine Sea basin; PVB, Parece Vela basin; S, Sulu Sea; SCS, South China Sea; SB, Shikoku basin; T, Taiwan.

Figure 2: Simplified bathymetry (isobath spacing, 1 km) near Taiwan with the portion of EU continental crust (in gray) subducted beneath the Luzon arc, during its collision with Eurasia. Proto-Taiwan thrust faults extension (northeast of Taiwan) from Hsiao et al. (1999). The position of EU crust boundary before the Okinawa Trough opening is with respect to EU 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.

Figure 3: Seismicity map of Taiwan with earthquakes recorded during the 1991-1997 period by the Central Weather Bureau in Taiwan. Deeper earthquakes are plotted above shallower earthquakes in order to underline the Ryukyu and Manila slabs. A is the western boundary of the PH Sea plate and B the ocean-continent transition zone within the EU subducted slab. Isobaths of subducted slabs are every 50 km. South of B, the Luzon arc is forming; between 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 A, the upper portion of the Luzon arc is probably accreted against the Ryukyu forearc.

Figure 4: 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 Okinawa Trough, Lallemand et al. (1999) for the Ryukyu accretionary wedge and forearc basins, Malavieille et al. (2002) for the Coastal Range (CR), Luzon arc, southern Longitudinal Trough (SLT) and North Luzon Trough (NLT) forearc basins, Liu et al. (1997) for the Manila accretionary prism. A is the western boundary of the PH Sea plate and B is the ocean-continent transition (OCT) zone within the EU subducted slab (in the prolongation of the

South China Sea OCT). East of A, in northern Taiwan and on the northeastern continental shelf, normal faults are reactivated thrust faults (Hsiao et al., 1999). Between A and B and south of B, all features are contractional. 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; IP, Ilan Plain; LF, Lishan fault and its extension in the Okinawa Trough. 1, Chinshan fault; 2, Shanjiao fault; 3, Nankan fault (Teng and Lee, 1996).

Figure 5: 3-D Sketch of southern Taiwan. A is the western boundary of the PH Sea plate (thick lines). B is the boundary between the oceanic and continental parts of the EU slab located in the prolongation of the South China Sea ocean-continent transition (OCT) zone. The Manila oceanic accretionary prism and the Taiwan continental accretionary prism are in light and dark gray, respectively. CR, Coastal Range; HP, Hengchun peninsula; LA, Luzon arc; LV, Longitudinal Valley.

Figure 6: 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 A, EU is torn by the westward motion of the PH Sea plate and thus has two boundaries (relict T1). 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 (OCT) zone. LV, Longitudinal Valley; OT, Okinawa Trough; OJ, Okinawa-Japan plate.

Figure 7: Simplified present-day plate tectonic context and types of deformation for each of the three Philippine (PH) Sea, Eurasia (EU) and Okinawa-Japan (OJ) plates. A is the PH Sea plate boundary and B is the ocean-continent transition (OCT) zone within the EU slab (in the prolongation of the South China Sea OCT). Relict 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), the portion of Taiwan Island south of A and the northern part of the Manila accretionary prism are moderately deformed by compressive stress; the northern Luzon and Coastal Range as well as the Manila forearc basins (Southern Longitudinal Trough (SLT) and North Luzon Trough (NLT)) are severely deformed by compressive stress; Northern Taiwan (east of A), the southern portion of the Manila accretionary prism and the Okinawa Trough are under extension (Kao et al., 2000; Teng and Lee, 1996). Green dashed lines are locations of the four synthetic profiles shown in Figure 8. a) The PH Sea plate is in blue and the Philippines to northern Taiwan accretionary prism is in light blue. b) EU oceanic (in green)

and continental (in yellow) plates; c) OJ plate in orange and Ryukyu accretionary prism in light orange.

Figure 8: Synthetic profiles located in Figure 7 (dashed green lines) showing the nature and boundaries of Eurasia (EU), Philippine (PH) Sea and Okinawa-Japan (OJ) plates. In green, EU oceanic lithosphere; in yellow, EU continental lithosphere; in blue, PH Sea oceanic lithosphere, in orange, OJ lithosphere. In light blue, Philippines to northern Taiwan accretionary prism; in light orange, Ryukyu accretionary prism. The scale of profile 4 is half the one of other profiles.

Figure 9: 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.

Figure 10: Sketch of the geodynamic evolution of Taiwan. Eurasia (EU) is fixed. A is the western boundary of the Philippine (PH) Sea plate. Since 9 Ma, the PH Sea plate 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 A, EU is torn and has two boundaries (relict 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 A and moves simultaneously with 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 T1 have subducted simultaneously with the underlying EU continental lithosphere. DF, deformation front; LV, Longitudinal Valley; OJ, Okinawa-Japan plate. PH Sea plate in blue and Philippines to northern Taiwan accretionary prism in light blue.

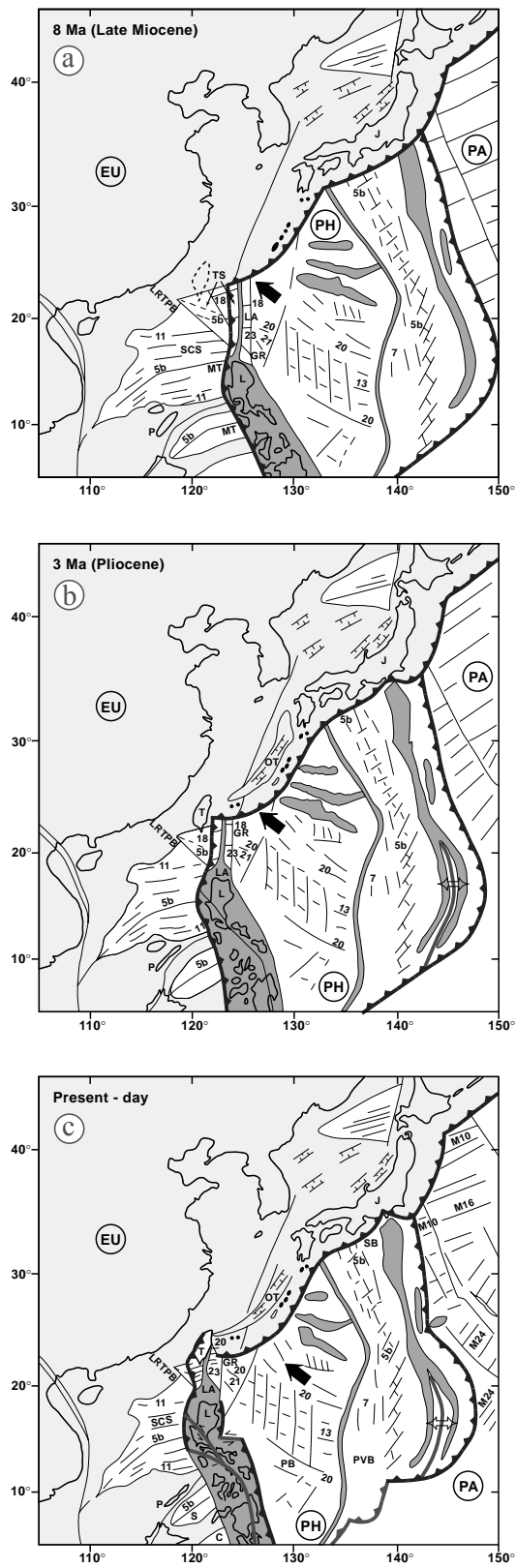


Figure 1

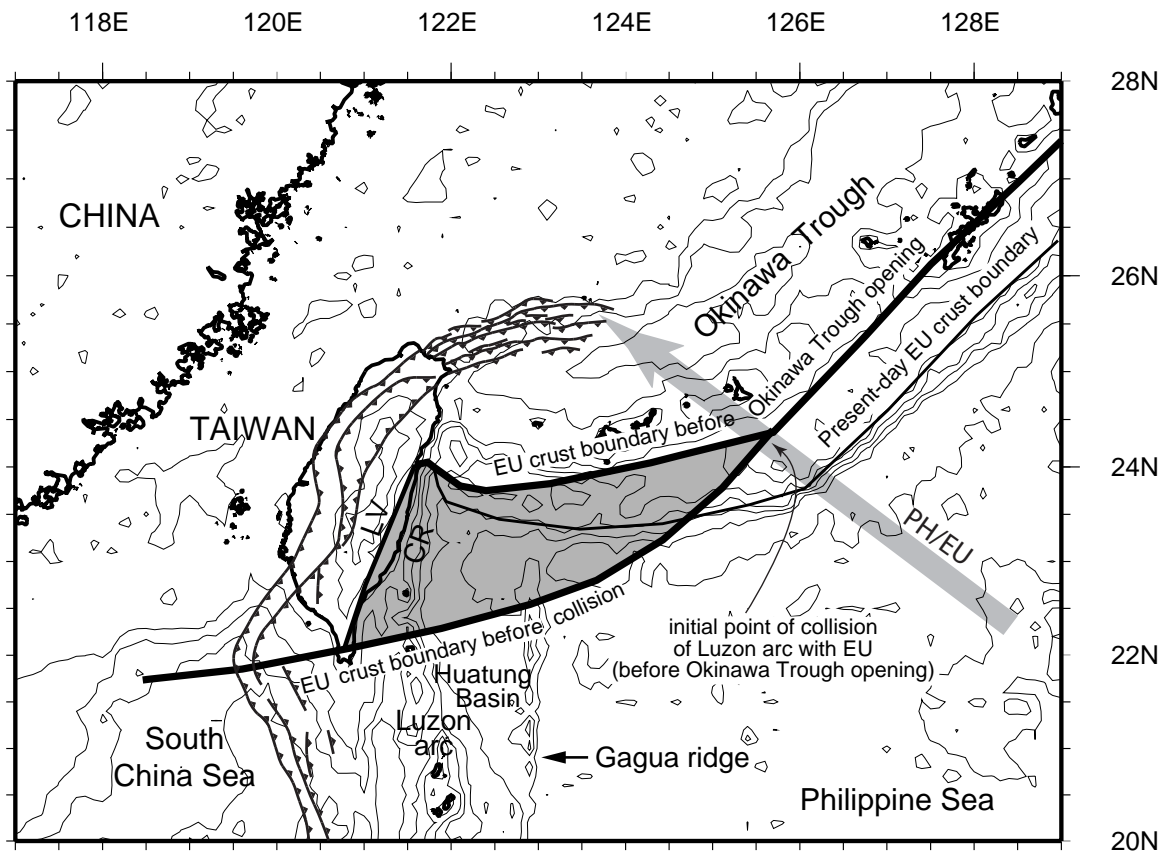


Figure 2

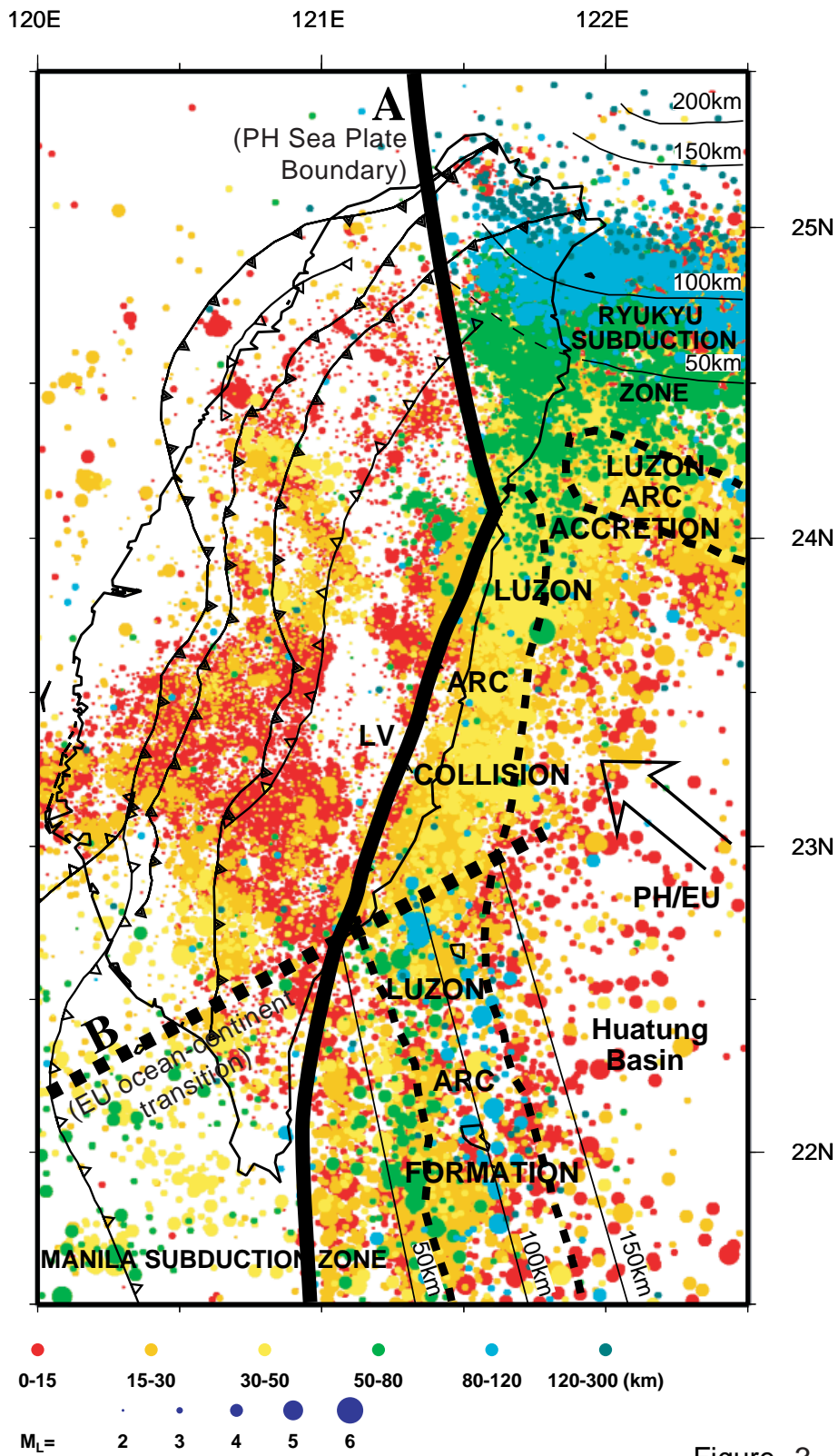


Figure 3

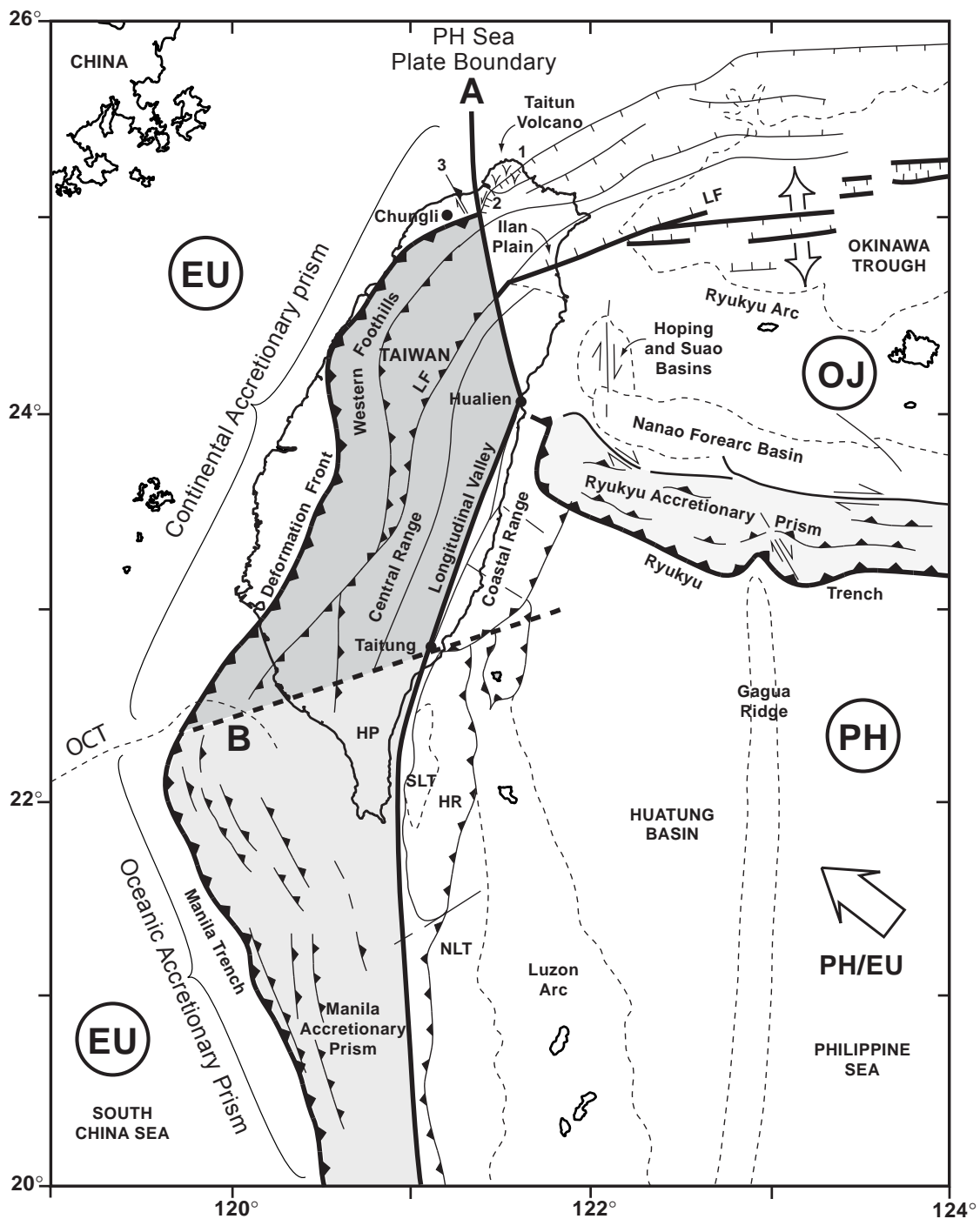


Figure 4

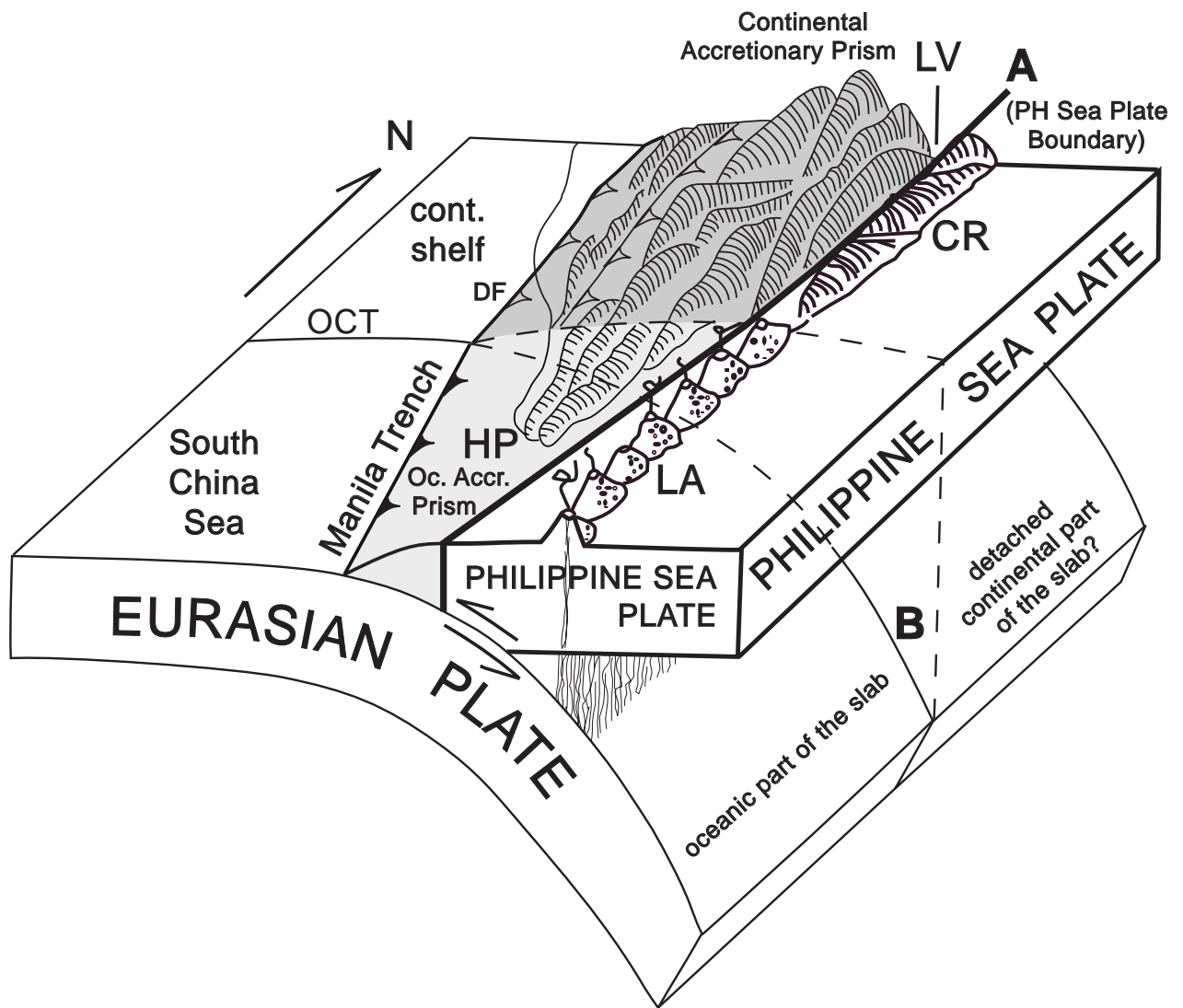


Figure 5

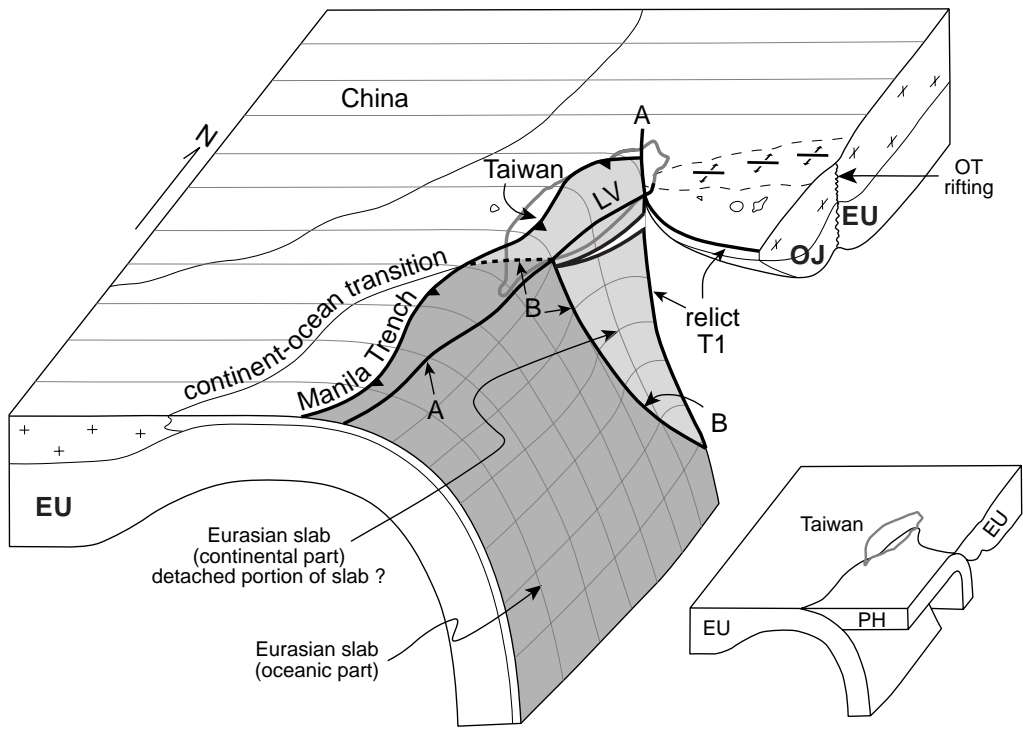


Figure 6

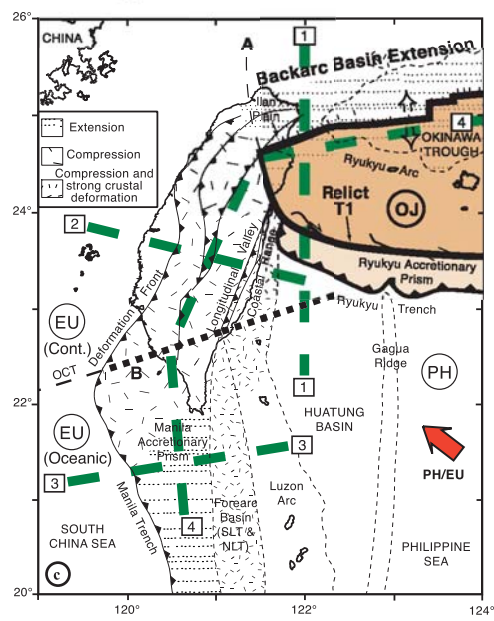
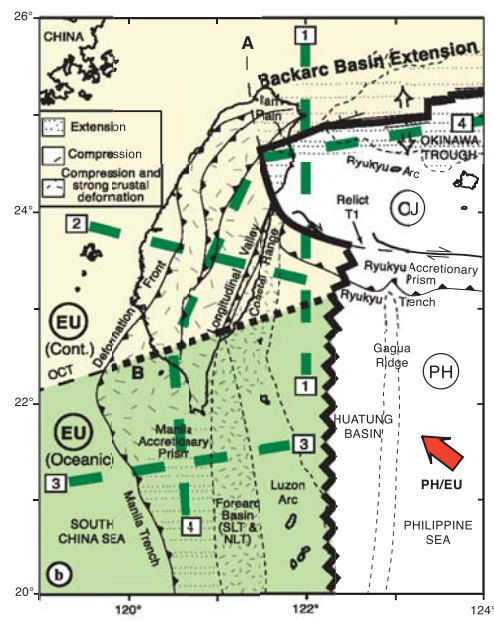
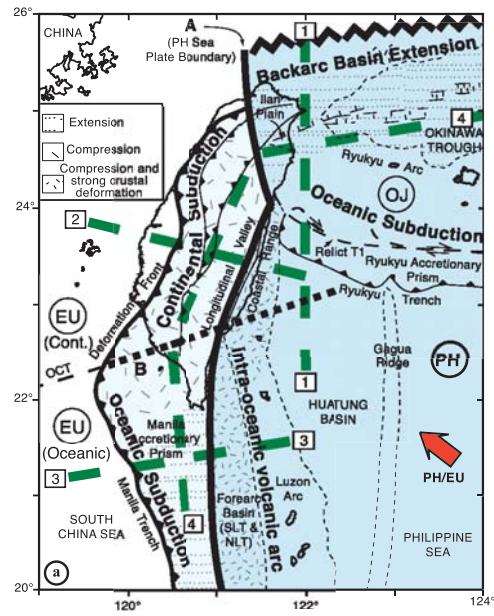


Figure 7

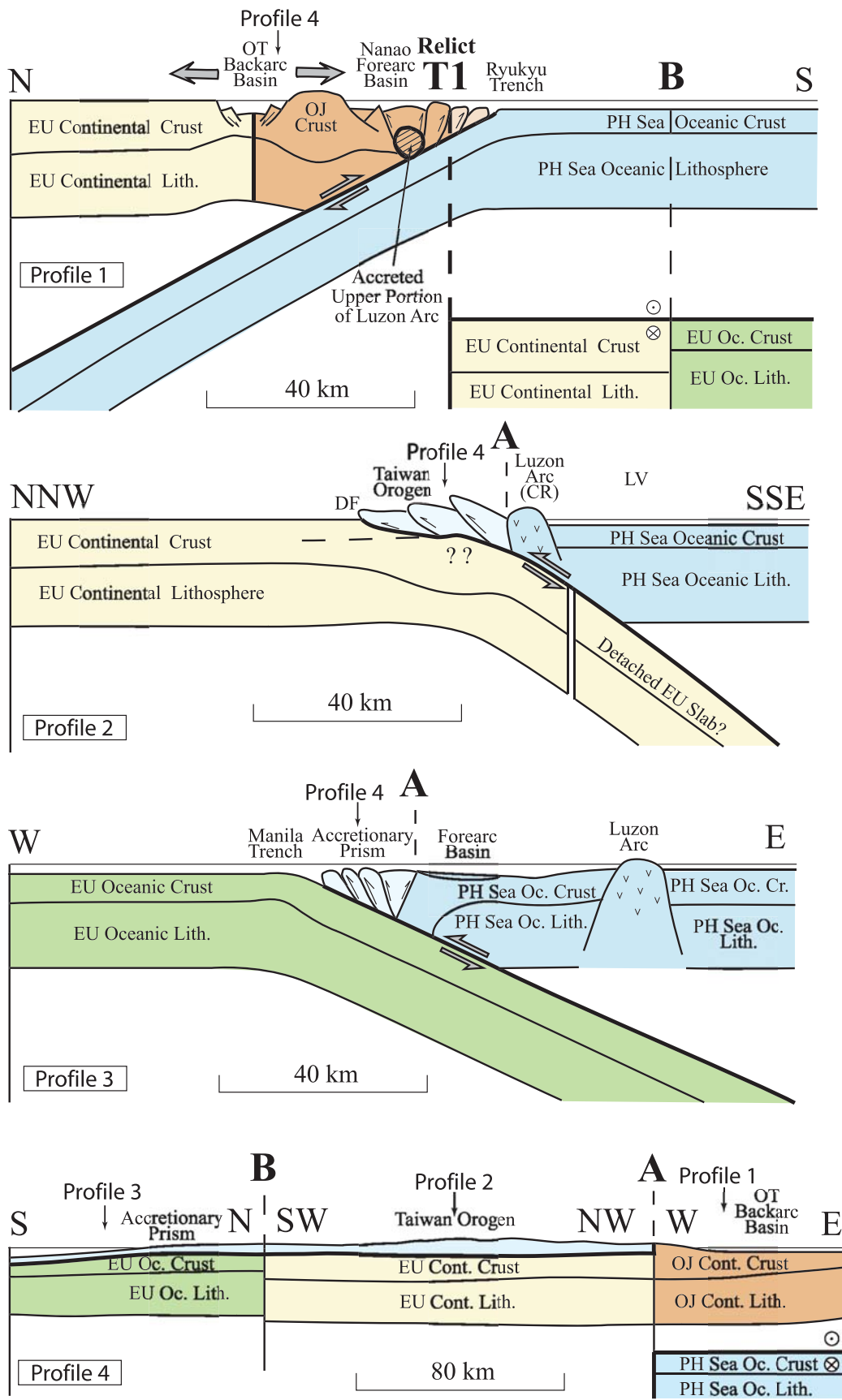


Figure 8

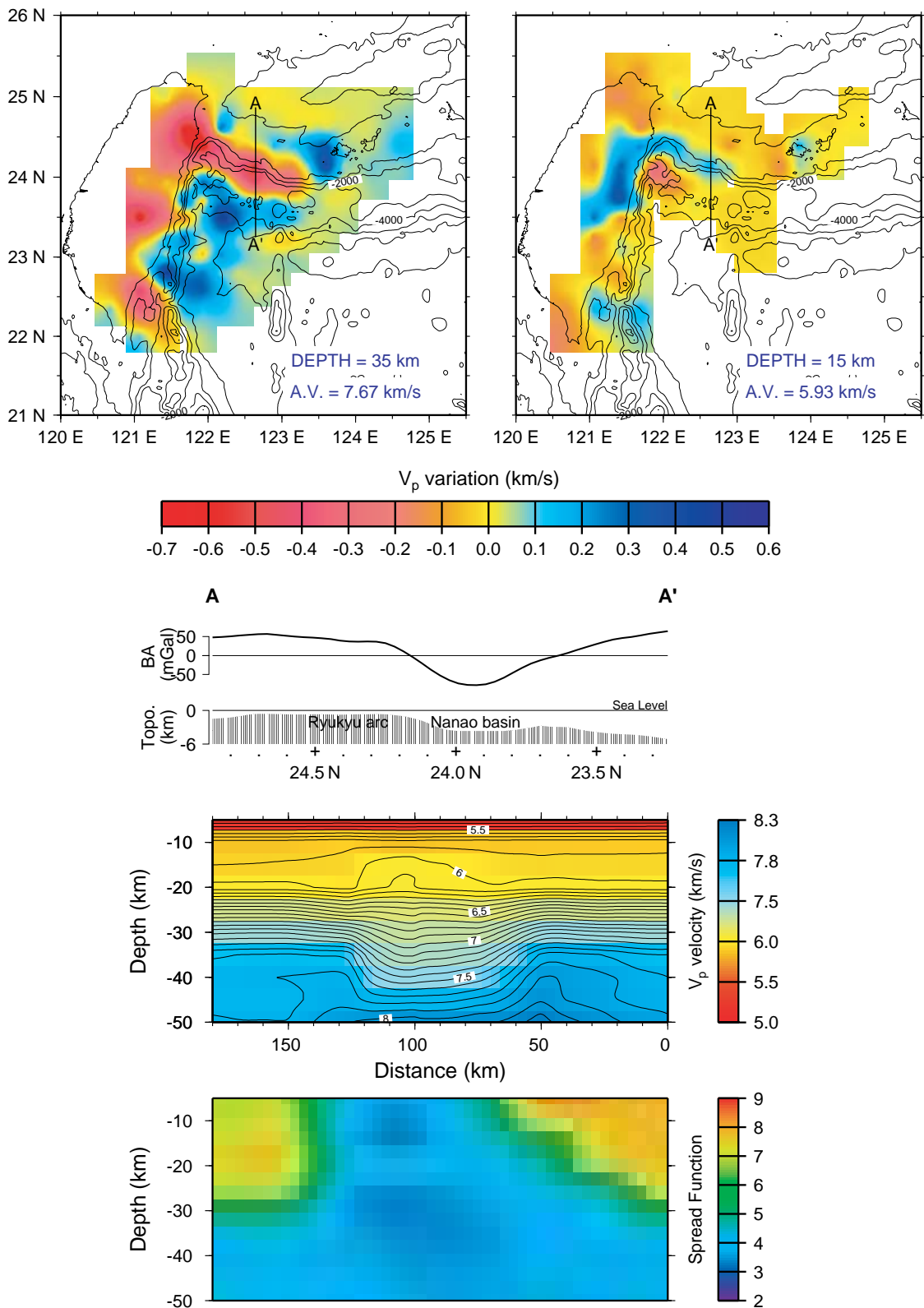


Figure 9

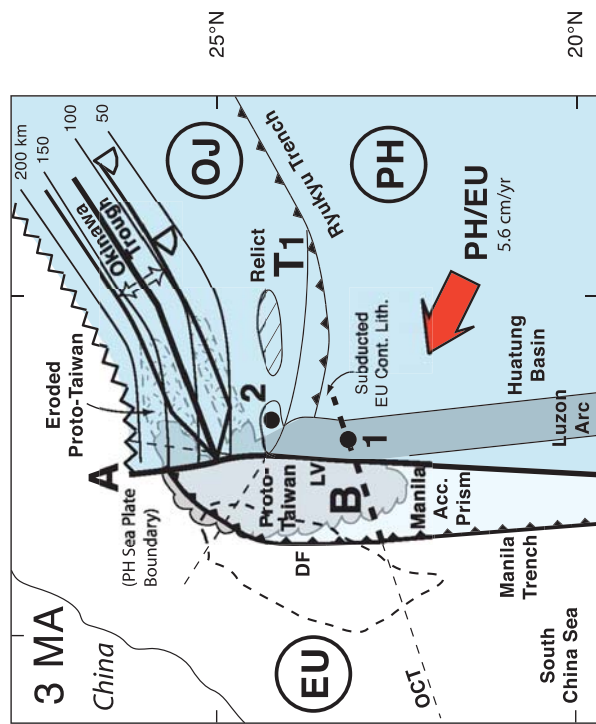
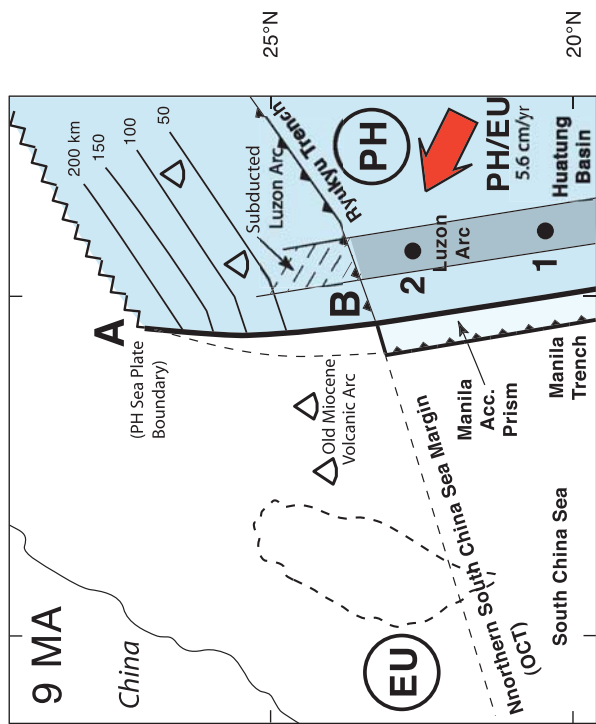
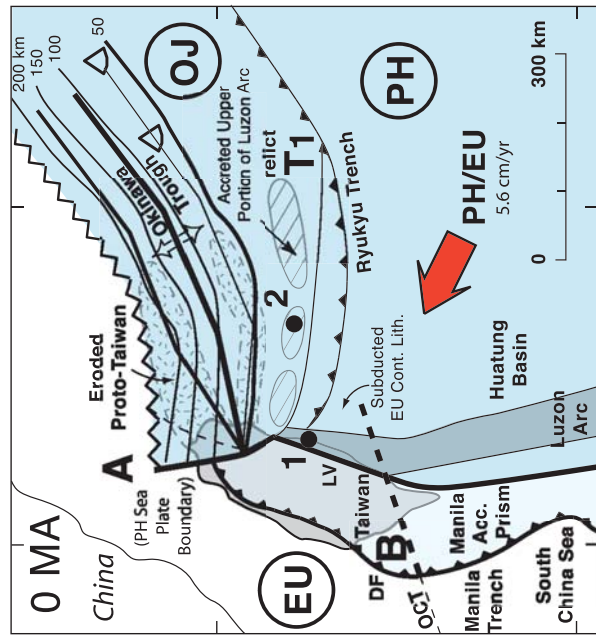
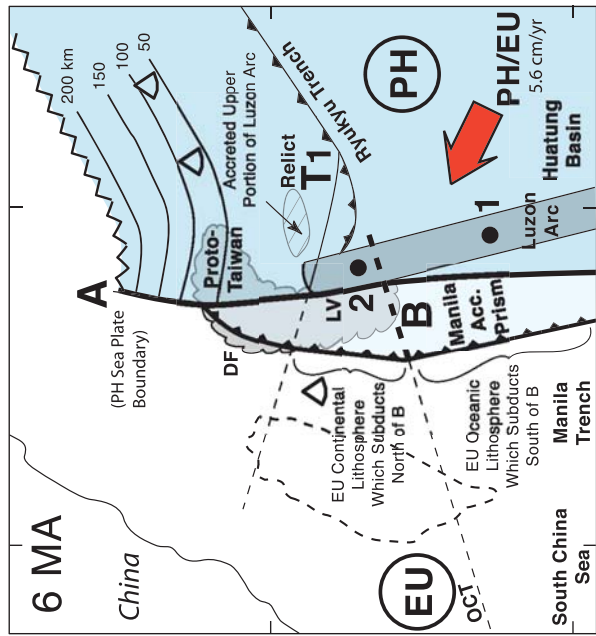


Figure 10