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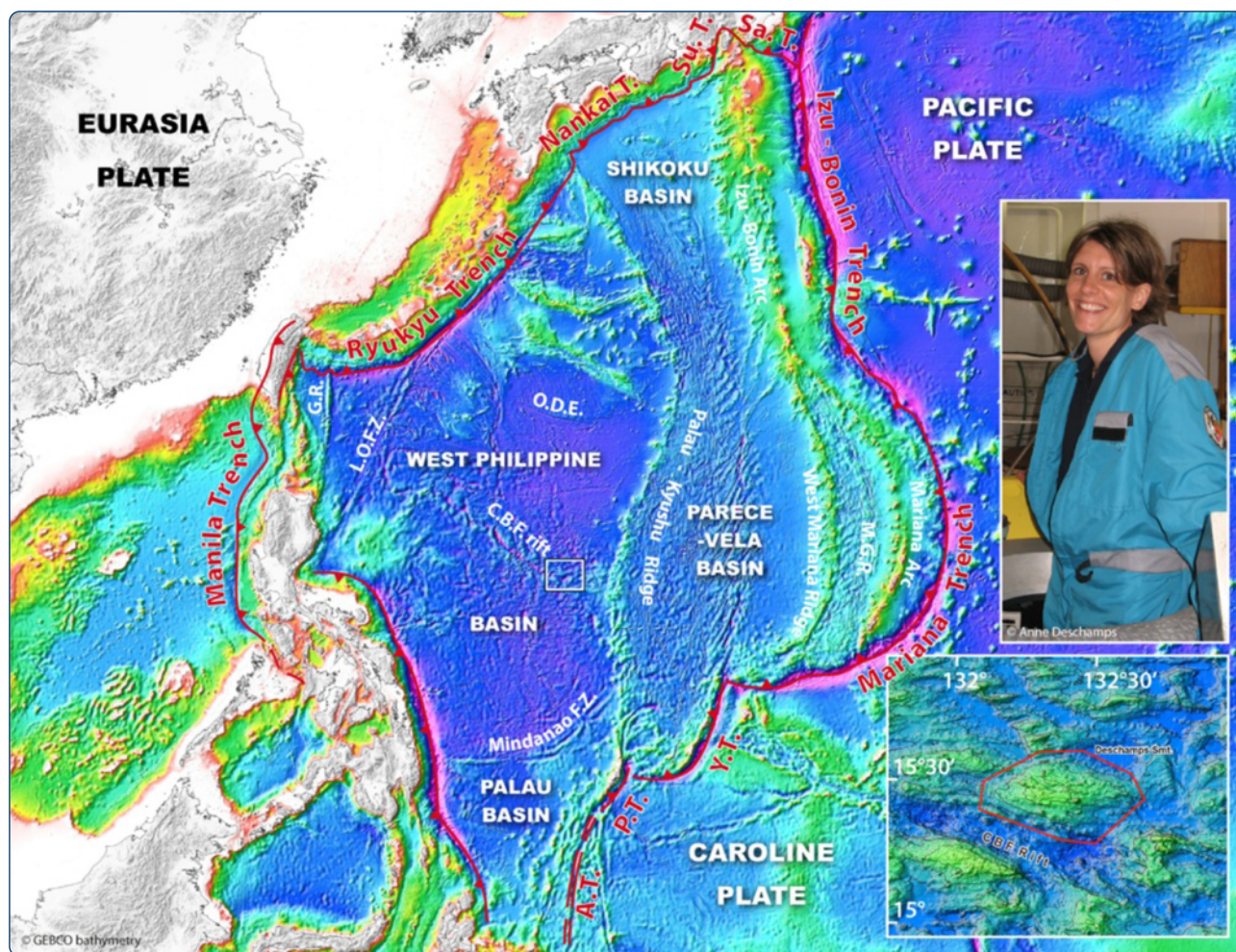
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Philippine Sea plate inception, evolution and consumption : focus on the newly named Deschamps seamount fringing the Central Basin Fault Rift in memory of Anne Deschamps

Lallemand

REVIEW

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Philippine Sea Plate inception, evolution, and consumption with special emphasis on the early stages of Izu-Bonin-Mariana subduction

Serge Lallemand^{1,2}

Abstract

We compiled the most relevant data acquired throughout the Philippine Sea Plate (PSP) from the early expeditions to the most recent. We also analyzed the various explanatory models in light of this updated dataset. The following main conclusions are discussed in this study. (1) The Izanagi slab detachment beneath the East Asia margin around 60–55 Ma likely triggered the Oki-Daito plume occurrence, Mesozoic proto-PSP splitting, shortening and then failure across the paleo-transform boundary between the proto-PSP and the Pacific Plate, Izu-Bonin-Mariana subduction initiation and ultimately PSP inception. (2) The initial splitting phase of the composite proto-PSP under the plume influence at ~54–48 Ma led to the formation of the long-lived West Philippine Basin and short-lived oceanic basins, part of whose crust has been ambiguously called “fore-arc basalts” (FABs). (3) Shortening across the paleo-transform boundary evolved into thrusting within the Pacific Plate at ~52–50 Ma, allowing it to subduct beneath the newly formed PSP, which was composed of an alternance of thick Mesozoic terranes and thin oceanic lithosphere. (4) The first magmas rising from the shallow mantle corner, after being hydrated by the subducting Pacific crust beneath the young oceanic crust near the upper plate spreading centers at ~49–48 Ma were boninites. Both the so-called FABs and the boninites formed at a significant distance from the incipient trench, not in a fore-arc position as previously claimed. The magmas erupted for 15 m.y. in some places, probably near the intersections between back-arc spreading centers and the arc. (5) As the Pacific crust reached greater depths and the oceanic basins cooled and thickened at ~44–45 Ma, the composition of the lavas evolved into high-Mg andesites and then arc tholeiites and calc-alkaline andesites. (6) Tectonic erosion processes removed about 150–200 km of frontal margin during the Neogene, consuming most or all of the Pacific ophiolite initially accreted to the PSP. The result was exposure of the FABs, boninites, and early volcanics that are near the trench today. (7) Serpentinite mud volcanoes observed in the Mariana fore-arc may have formed above the remnants of the paleo-transform boundary between the proto-PSP and the Pacific Plate.

Keywords: Philippine Sea Plate, Izu-Bonin-Mariana, Subduction initiation, Boninite, Fore-arc basalt, Serpentinite mud volcano, Back-arc basin, Transform fault, Arc terrane, Plume-ridge interaction

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Introduction

The visible part of the Philippine Sea Plate (PSP) has a diamond shape with a maximum north-south length of ~3400 km and a maximum east-west width of ~2600 km. Its extent below southwest Japan, the Ryukyu Arc, and the Philippine Arc (see Fig. 1) is outlined by the Benioff zones and tomography anomalies (Kao and Chen 1991; Bijwaard et al. 1998; Wang et al. 2008). Removal of the PSP edges through tectonic erosion has also been documented, especially along the Izu-Bonin-Mariana (IBM) Trench (Hussong and Uyeda 1981; Bloomer 1983; Fryer et al. 1992; Lallemand 1995). Considering that the eastern Shikoku, Parece Vela and Mariana basins, which opened as back-arc basins (Karig 1971a; Uyeda and Ben Avraham 1972) during the Neogene, it is clear that the shape and area of the PSP has changed over time. The PSP subducts under southwest Japan and the Ryukyu Arc along its northwestern side, and under the Philippine Arc along its southwestern side. Conversely, it overrides the Pacific

Plate along its eastern edge (fringed by the IBM Arc) and its central-western edge (fringed by the Luzon Arc). To summarize, most of its boundaries consist of subduction zones except a small divergent segment in the south, called the Ayu Trough (Fujiwara et al. 1995).

Many authors have contributed to our understanding of the tectono-magmatic evolution of the PSP (e.g., Uyeda and Ben Avraham 1972; Hilde and Lee 1984; Seno and Maruyama 1984; Hickey-Vargas 1991; Stern and Bloomer 1992; Hall et al. 1995a; Okino et al., 1999; Deschamps and Lallemand 2002). From among this group, I would like to dedicate this review to Anne Deschamps who recently departed.

The main objectives of this review paper are as follows:

- to scan the nature and age of the various parts of the PSP including the subducted and eroded parts that represent about one third of its total area.

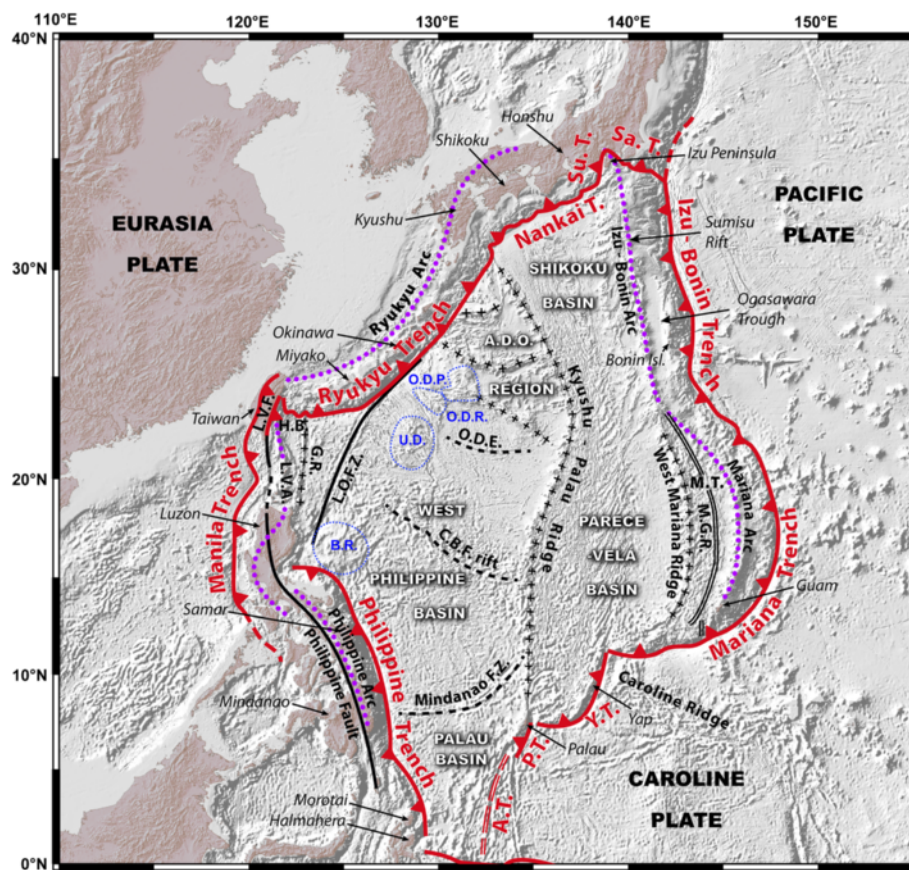


Fig. 1 Philippine Sea Plate boundaries and toponymy. Plate boundaries (in red) are reported on a bathymetric map based on GEBCO data. Main faults are represented by thick black lines and active spreading centers by double lines. Red triangles are located on the overriding plates along the subduction boundaries. Blue dotted lines outline the OIB plateaus. From north to south: Su.T. Suruga Trough, Sa.T. Sagami Trough, Nankai T. Nankai Trough, A.D.O. Region Amami-Daito-Oki-Daito Region, Bonin Isl. Bonin Islands, O.D.P. Oki-Daito Plateau, O.D.R. Oki-Daito Ridge, U.D. Urdaneta Plateau, O.D.E. Oki-Daito Escarpment, L.V.F. Longitudinal Valley Fault, H.B. Huatung Basin, G.R. Gagua Ridge, L.V.A. Luzon Volcanic Arc, L.O.F.Z. Luzon-Okinawa Fracture Zone, B.R. Benham Rise, C.B.F. rift Central Basin Fault rift, M.T. Mariana Trough, M.G.R. Malaguana-Gadao Ridge, Mindanao F.Z. Mindanao Fracture Zone, Y.T. Yap Trench, P.T. Palau Trench, and A.T. Ayu Trough

- to portray the major tectonic events that have shaped the PSP such as its inception, subduction initiation, intraplate deformation including shortening, rifting and spreading, or tectonic erosion and subduction along some of its boundaries.
- ultimately, to give rise to new insights into the PSP tectonic evolution and serve as a guide for new investigations.

Review

Overview of the non-subducted part of the present-day PSP

The boundaries of the PSP have changed tremendously through time. Indeed, most of the so-called “Philippine mobile belt” still belonged to the PSP during the Miocene (e.g., Moore and Silver 1982; Lewis and Hayes 1983; Hall 1987; Rangin et al. 1990; Lallemand et al. 1998). In the past, the transpressive plate boundary was located either west of or within the archipelago. The reader can refer to Pubellier et al. (2004) among others for accounts of the huge continental and oceanic sliver motions of this area during the Neogene. The southeastern boundary, i.e., Mariana and Yap Trenches in particular, considerably lengthened through time to accommodate the spreading of the West Philippine Basin, PareceVela Basin, and Mariana Trough (Fig. 1; Deschamps and Lallemand 2002; Kobayashi 2004; Okino et al. 2009; Ribeiro et al. 2013). In this section, we will focus on the visible part of the present-day PSP, aiming to trace the various pieces of the puzzle that result from a multi-stage tectonic evolution.

Present-day boundaries of the PSP

The PSP was the fastest-moving plate during Cenozoic times (Zahirovic et al. 2015). Today, the absolute motion of its southern part (~10 cm/year) in classical reference frames (DeMets et al. 2010; Seton et al. 2012 or Kreemer et al. 2014) is comparable to that of the Pacific Plate, which is presently the fastest plate in the world. The high speed of these two plates is primarily due to the slab pull force exerted along their northwestern boundaries (i.e., Forsyth and Uyeda 1975; Spence 1987; Pacanovsky et al. 1999; Faccenna et al. 2007). The PSP subducts toward the northwest along ~2500 km of the Ryukyu Trench and Nankai Trough and 1500 km of the Philippine Trench (Fig. 1). This dominant force controls the northwest motion of the plate. The plate is surrounded by subduction zones. It is the down-going plate along most of its western edges, and the overriding plate along its eastern edge (marked by the Izu-Bonin-Mariana, Yap and Palau trenches) and the central part of its western edge (marked by the Manila Trench). There, between Luzon and Taiwan, the plate boundary is not a single distinct unit because the eastward-dipping Manila Trench overlaps the westward-

dipping Philippine Trench at the latitude of Luzon island and the northward-dipping Ryukyu Trench at the latitude of Taiwan. In the vicinity of both islands, the relative convergence between the Eurasian Plate and the PSP is accommodated along several faults, which include the Longitudinal Valley Fault and deformation front in Taiwan (Angelier 1986) and the Philippine Fault and Manila Trench in Luzon (Pubellier et al. 2004). In the south, the short Palau Trench extends over a distance of ~500 km into a slow-spreading ridge, called the Ayu Trough, which is supposed to have been active since 25 Ma (Weissel and Anderson 1978; Fujiwara et al. 1995). The southernmost E-W-trending boundary corresponds to an area of diffuse strike-slip deformation, which includes the Sorong Fault (Hall 1987). There are two major collisions between active volcanic arcs carried by the PSP and the adjacent margins. One is in the west with the Luzon volcanic arc. It has resulted in the Taiwan orogeny since the Pliocene (e.g., Angelier 1986; Lallemand et al. 2001; Malavieille et al. 2002). The other is in the north with the IBM arc. It has resulted in the development of the Izu Collision Zone since the Upper Miocene (e.g., Taira et al. 1989; Aoike 2001; Arai et al. 2009). Lallemand (2014) described the deformation modes of the subducting PSP associated with these two arc-continent collisions.

The PSP itself appears as a mosaic of oceanic basins, aseismic ridges, plateaus, fracture zones, volcanic arcs and fore-arcs, fossil, and active spreading centers. Despite some controversy on the age of the small westernmost Huatung Basin (see Fig. 1), oceanic basins are younger from west to east.

Oceanic basins of the PSP

The West Philippine Basin (WPB) is the largest of the PSP oceanic basins at ~1500 km long and ~1100 km wide. It occupies more than one third of the plate's surface. The main scar, which is east of Luzon and trends WNW-ESE, was initially called the Central Basin Fault (CBF). Mrozowski et al. (1982) and then Hilde and Lee (1984) demonstrated that symmetric spreading occurred on both sides of that feature, which was thus identified as a fossil spreading center and renamed the CBF rift. Its unusual depth, which reaches a maximum of 7900 m, has been explained by post-spreading anagmatic extension (Fujioka et al. 1999; Deschamps et al. 1999, 2002; Okino and Fujioka 2003). Several ridge jumps, plateau emplacements and overlapping spreading centers were interpreted by Deschamps et al. (2002, 2008) as the expression of plume-ridge interaction in the image of the spreading pattern in the North Fiji Basin (Lagabriele et al. 1997; Faccenna et al. 2010). Taylor and Goodliffe (2004), mainly based on a single line acquired south of the CBF rift, suggested that the opening direction of the basin rotated 100° clockwise in the period from 48–49

to 33 Ma. Another striking difference from the initial model proposed by Hilde and Lee (1984) concerns the northernmost and southernmost parts of the WPB. Deschamps and Lallemand (2002) showed that magnetic lineations trend N-S north of the Oki-Daito escarpment (ODE, Figs. 1 and 2), i.e., perpendicular to the main trend on both sides of the CBF rift. They postulate that this piece of oceanic crust was the oldest in the WPB based on a dated basalt collected at site 294/5. According to their interpretation, rifting may have started as soon as 55 Ma ago behind the proto-Philippine Arc.

Recently, Sasaki et al. (2014) provided new insights on the southernmost part of the WPB including the Palau Basin, south of the Mindanao Fracture Zone (FZ, see Fig. 1). Based on three short magnetic profiles across the Palau Basin and a dolerite sampled along the Mindanao FZ dated at ~40 Ma (Ishizuka et al. 2015), they concluded that the Palau Basin exhibits N-S magnetic lineations, as it does north of the ODE. Their magnetic model, even poorly constrained, provides an age of 35–40 Ma, which is quite young compared to its counterpart in the north. Whatever the age of this basin, it is now clear that it is made of normal oceanic crust and spreading occurred E-W (in its present position) south of the Mindanao FZ. The first estimate of the spreading period based on magnetic lineations

by Hilde and Lee (1984) of between 60 and 35 Ma was revised by Deschamps and Lallemand (2002) to from 54 to 33/30 Ma with a short, late extension episode between 30 and 26 Ma. Other authors (Taylor and Goodliffe 2004; Sasaki et al. 2014) refute the oldest ages proposed by Hilde and Lee (1984) in the southern part of the WPB because it has been proved that the seafloor fabric changes drastically south of the Mindanao FZ. Unlike most previous authors, Sasaki et al. (2014) consider a constant spreading rate for the opening of the WPB and a progressive cessation of spreading from 37.5 Ma in the southeast to 35.5 Ma in the northeastern part of the CBF rift.

Hilde and Lee (1984) also supposed that the WPB extended westward in the Huatung Basin across the north-south trending Gagua Ridge, whereas Deschamps et al. (2000) argued that dredged gabbros, dated as Early Cretaceous in the Huatung Basin, invalidated this hypothesis. The age of this small basin is highly controversial. Estimates vary from 131 to 119 Ma (Deschamps et al. 2000) to 52–43 Ma for the northern part and Early Cretaceous for the southern part (Sibuet et al. 2002), to 44–33 Ma (Hilde and Lee, 1984), 42–33 Ma (Doo et al. 2015), or even ~30–15 Ma (Kuo et al. 2009). Further details and an interesting discussion about this controversy are available in Eakin et al. (2015).

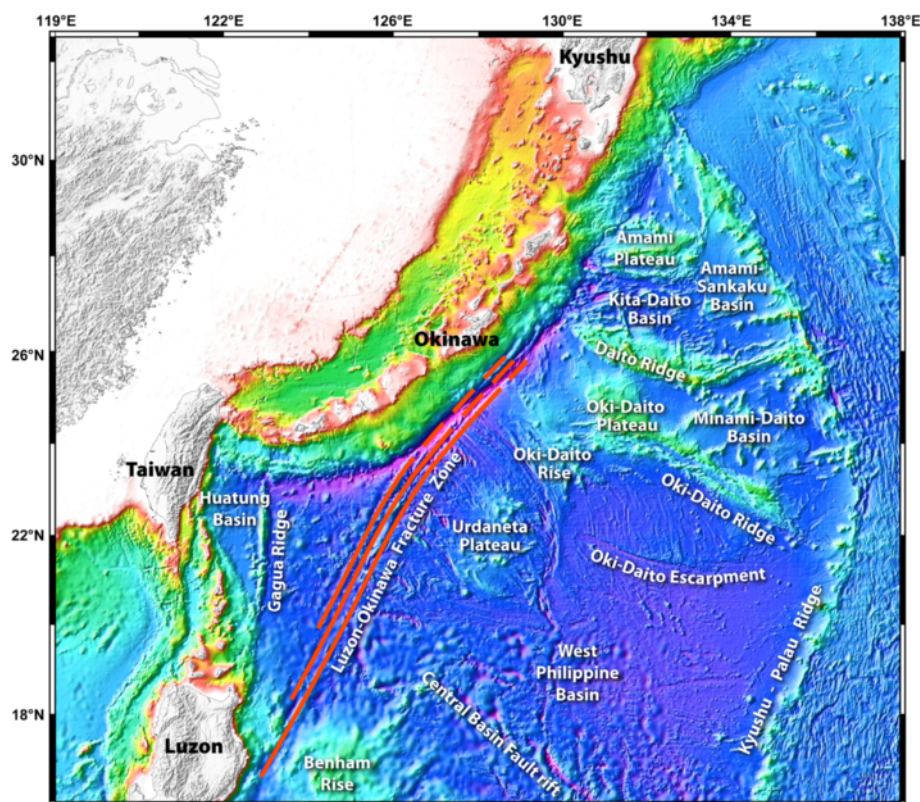


Fig. 2 Close-up view of the bathymetry of the northern half of the WPB. Details of the Amami-Daito-Oki-Daito region, Gagua Ridge and Luzon-Okinawa Fracture Zone are outlined

East of the WPB lies a north-south elongated (~2500 km long by ~200–600 km wide) pair of oceanic basins called Shikoku and Parece Vela. Both basins rifted right after the cessation of spreading in the WPB, i.e., ~30 Ma ago, then spreading occurred between 29–26 and 15 Ma (Chamot-Rooke et al. 1987; Okino et al. 1994, 1998). Sdrolias et al. (2004) further detailed the kinematics of this “twin-back-arc opening,” characterizing it as the consequence of Pacific slab rollback and PSP clockwise rotation. The back-arc region ruptured simultaneously in the north on the Shikoku rift and in the south on the Parece Vela rift about 2500 km apart. Spreading centers propagated toward each other and merged at ~23 Ma, forming an R-R-R triple junction that accommodated the difference in the spreading orientations of the two basins. The dramatic change in spreading orientation and rate after 20 Ma is interpreted by Sdrolias et al. (2004) as the expression of PSP rotation.

The youngest oceanic basin, the Mariana Trough, which is ~1000 km long and ~200 km wide, has opened behind the Mariana Arc since ~6–8 Ma (Husson and Uyeda 1981; Fryer 1995; Stern et al. 2003; Kato et al. 2003). Many authors have studied the morphological, geophysical, and petrological characteristics of this basin (Karig 1971b; Baker et al. 1996; Yamazaki and Murakami 1998; Stüben et al. 1998; Martinez et al. 1995, 2000; Deschamps and Fujiwara 2003; Deschamps et al. 2005). The asymmetric (full) spreading rate varies from 20 mm/year in the north to 40 mm/year in the south (Asada et al. 2007). This is partly responsible for the curvature of the arc and its stretching, especially south of Guam where a rift crossing the arc and the fore-arc accommodates the spreading (Ribeiro et al. 2013; Stern et al. 2013).

The Izu-Bonin Arc, north of the Mariana Arc, has also been affected by trench-perpendicular extension since 2 Ma but the neo-formed intra-arc basin, ~120 km long and ~40 km wide, called the Sumisu Rift (Fig. 1), is still in the rifting stage (Taylor et al. 1990, 1991).

Aseismic ridges and plateaus of the PSP

The PSP oceanic basins are separated by aseismic ridges and plateaus, except along the eastern edge where active volcanic arcs bound the Mariana Trough and Sumisu Rift (see Fig. 1). The longest (~2600 km) aseismic ridge, called the Kyushu-Palau Ridge (KPR), trends north-south in the middle of the plate. It is a remnant arc (Karig 1972) separated from the IBM active volcanic arc 25 Ma ago (Ishizuka et al. 2011a) during the opening of the Shikoku and Parece Vela back-arc basins. It consists of aligned volcanoes over a width of ~50–150 km, presenting a marked change in the orientation from NNW-SSE to NNE-SSW around ~23° N. Similarly, the aseismic West Mariana Ridge is a remnant arc that separated

from the Mariana active volcanic arc ~7 Ma ago (Karig and Glassley 1970; Gardner 2010). Both the KPR paleo-arc on one side and the Izu-Bonin arc and the West Mariana Ridge on the other side are conjugate margins of the Shikoku and Parece Vela basins.

West of the KPR, there are two regions with aseismic ridges, a large complex triangular-shaped region, called the Daito Ridges region or Amami-Daito-Oki-Daito (ADO) region, that lies north of the WPB, and a sharp linear ridge trending north-south, west of the WPB close to Taiwan, called the Gagua Ridge (Fig. 2).

The ADO region comprises three ridges from north to south: the Amami Plateau, the Daito Ridge, and the Oki-Daito Ridge and Plateau (Fig. 2). Tokuyama (1985, 2007), based on petrology, concluded that the Amami Plateau, composed of three 200-km-long E-W-trending ridges, was an active island arc before the Eocene (Nishizawa et al. 2014). It is presently subducting beneath the Ryukyu Arc. South of the Amami Plateau, the Daito Ridge extends E-W for more than 500 km. It also subducts on its western side and connects with the KPR on its eastern side. Basalts collected there also indicate an island arc setting (Tokuyama et al. 1986). Further south, the Oki-Daito Ridge s.l., which is 700 km long, is composed, from west to east, of the Oki-Daito Rise, the Oki-Daito Plateau, and the Oki-Daito Ridge s.s. Volcanic edifices on the Oki-Daito Ridge, as well as in the Minami-Daito Basin north of it, have similar ocean island basalt (OIB) signatures and an estimated age of 44–48 Ma (Ishizuka et al. 2013). Most of the Daito ridges probably originated from island arcs, based on their similar velocity structures (Nishizawa et al. 2014) and their petrology (Hickey-Vargas 2005; Ishizuka et al. 2013). The Oki-Daito Rise, southwest of the main ridge, is an exception in terms of velocity structure, as it has a thinner crustal section. Nishizawa et al. (2014) concluded that it is quite clear from velocity structure and magnetism that the Kita-Daito Basin resulted from the splitting of the Amami Plateau and Daito Ridge, but there are no distinctive features indicative of subduction along the southern edge of the Daito Ridge as earlier suggested by Tokuyama et al. (1986).

In the northwestern part of the WPB, we distinguish two broad highs at equal distances from the CBF rift (Fig. 2), the Benham Rise east of Luzon and the Urdaneta Plateau south of Okinawa. Ocean Island Basalt (OIB)-like lavas dated at ~36 Ma were sampled on the Benham Rise (Hickey-Vargas 1998), and Deschamps et al. (2008) described overlapping spreading centers near the Urdaneta Plateau. All these observations strongly support the influence of a mantle plume called the “Oki-Daito mantle plume” contemporaneous with the first stages of the WPB rifting and spreading (Deschamps and Lallemand 2002; Ishizuka et al. 2013). The three

small basins sandwiched between the Daito ridges, i.e., from south to north the Minami-Daito Basin, the Kita-Daito Basin and the Amami-Sankaku Basin, are poorly known. Their crustal thickness is typical of oceanic crust (Nishizawa et al. 2014), and the few available samples, obtained from drilling in the Minami-Daito Basin, give ages for OIB sills of ~51 and ~43 Ma (Hickey-Vargas, 1998). E-W-trending magnetic lineations in the Kita-Daito Basin are indicative of N-S spreading (Nishizawa et al. 2014). The Amami-Sankaku Basin was drilled during the last International Ocean Discovery Program (IODP) expedition 351. Preliminary results suggest a ~50–55 m.y. old basaltic basement geochemically similar with “fore-arc basalts” from the IBM arc (see discussion in section 1.5, Expedition 351 Scientists 2015; Ishizuka et al. 2015). Finally, the small oceanic basins in the ADO region are roughly contemporaneous with the first stages of WPB spreading and could be considered to be the result of short-lived multiple spreading centers originating from the same regional magmatic event (Deschamps and Lallemand 2003).

East of Taiwan, the 300 km-long, 4 km-high Gagua Ridge trends north-south along 123° E. This high separates the small Huatung Basin to the west from the main WPB to the east. Magnetic lineations of the Huatung Basin intersect the ridge at an angle of 90°, which strongly suggests that the Gagua Ridge was a former fracture zone (Mrozowski et al. 1982; Hilde and Lee 1984; Deschamps et al. 1998). On the west side of the ridge, magnetic lineations of the WPB are highly oblique, so the Gagua Ridge is a non-transform discontinuity (NTD). Velocity and gravity models across the ridge suggest an episode of westward underthrusting of the WPB oceanic crust beneath the Huatung Basin (Deschamps et al. 1998; Eakin et al. 2015). The Gagua Ridge thus represents failed subduction along a former fracture zone or transform boundary between two oceanic basins. That episode of short-lived subduction occurred in the Oligocene according to Deschamps and Lallemand (2002), whereas Eakin et al. (2015) favor a younger Miocene age.

Non-transform discontinuities and fracture zones of the PSP

Among the NTDs and fracture zones observed in the PSP, some are prominent like the Luzon-Okinawa Fracture Zone (LOFZ), or the numerous curved fracture zones that offset the short segments of the Parece Vela spreading center. Others are suspected based on their orthogonality with the seafloor fabric such as the ODE, the Mindanao FZ, the Gagua Ridge or even the Oki-Daito Ridge. The LOFZ consists of a bundle of two to four parallel NE-SW-trending strike-slip faults offsetting the westernmost piece of the WPB. It extends from the mid-point between Luzon Island and Benham Rise to

the Ryukyu Trench south of Miyako Island and then parallels the trench until at least the fore-arc area off Okinawa Island (see Figs. 1 and 2 and map in Hsu et al. 2013). It is difficult to estimate the total offset along the LOFZ because the fossil spreading center west of the fracture zone is entirely subducted beneath the Ryukyu fore-arc and because its location between the northern termination of the CBF rift and the LOFZ is not clear. Based on magnetic identifications, the age of the oceanic crust west of the LOFZ ranges from 54 Ma in the south to younger than 47.5 Ma in the north (Doo et al. 2015). Shinjo and Ishizuka (2015) noticed that all samples collected east of the LOFZ have geochemical plume signatures, contrary to those reported west of the fracture zone. This is a strong argument for a large offset of perhaps several hundred kilometers along that fault zone.

IBM, Yap, and Palau volcanic arcs

The IBM arc is a 3000 km-long intra-oceanic island arc dated at ~50 Ma (Cosca et al. 1998). It experienced a succession of magmatic episodes accompanying, and recording, subduction initiation as well as several periods of seafloor rifting and spreading. The first spreading phase, at the origin of the West Philippine Basin, was contemporaneous with its initiation and building, probably in interaction with the Oki-Daito mantle plume (Ishizuka et al. 2013). The second phase of spreading, resulting in the Shikoku and Parece Vela basins, split the arc ~25 Ma ago into a remnant arc to the west (the present KPR) and the present IBM arc that remained continuously active. The last spreading episode occurred ~6–8 Ma ago in the Mariana section isolating the West Mariana Ridge to the west from the active Mariana Arc to the east. The oldest evidences of the initial IBM arc consist of gabbros and basalts dated ~51–52 Ma, which were found both in the Izu-Bonin fore-arc east of the Bonin Islands (Ishizuka et al. 2008, 2011b) and the Mariana fore-arc south of Guam (Reagan et al. 2013). The stratigraphic section of the fore-arc crust consists, from bottom to top, of peridotites, gabbroic rocks, a sheeted dike complex, basaltic lava flows, lavas and dikes of boninite and their differentiates, transitional high-Mg andesites, and tholeiitic and calc-alkaline arc lavas (Ishizuka et al. 2014). Reagan et al. (2010) named the oldest volcanic products “fore-arc basalts (FABs)”, mainly because they are exposed in the present IBM fore-arc. This name is ambiguous because (1) these lavas have geochemical affinities with mid-ocean ridge basalts (MORBs), and (2) because similar lavas of the same age (~50–52 Ma) were drilled during IODP Expedition 351 in the Amami-Sankaku Basin (ADO region) west of the KPR (Arculus et al. 2015). This discovery indicates that the area of initial seafloor spreading contemporaneous with subduction initiation

extended from the present-day fore-arc to the region west of the present-day KPR (Ishizuka et al. 2015). Observations made in the central Mariana Arc indicate that transitional, and then boninitic lavas, dated ~44–48 Ma, overlie the FABs, constituting a coherent suite of magmatic rocks characteristic of ophiolite assemblages (Reagan et al. 2010). Variations in Cr/(Cr + Al) atomic ratios of spinels in dunites from the IBM “present-day fore-arc” probably reflect changing melt compositions from MORB-like to boninitic melts due to an increase of slab-derived hydrous fluids and/or melts (Morishita et al. 2011). Similarly, transitional suites of high-Mg andesites, dated ~44–45 Ma, preceded the eruption of normal tholeiitic to calc-alkaline arc magmatism (Ishizuka et al. 2011b).

Fewer data are available for the Yap and Palau arc systems, but it has been established that the Yap arc is no longer active volcanically since the Late Oligocene or Miocene, right after its collision with the Caroline Ridge (Hawkins and Batiza 1977; McCabe and Uyeda 1983). Fujioka et al. (1996) observed ultramafic and gabbroic rocks along the inner trench slope of the Yap Trench with a paleo-Moho discontinuity around 6000 m deep. Metamorphic and gabbroic rocks from the Yap Arc appear quite similar to those exposed in the Parece Vela Basin (Ohara et al. 2002). The islands and submerged arc of Palau are thought to be the southern extension of the KPR (Kobayashi 2004). They consist of 20–38 Ma arc-type volcanic rocks including basalt, andesite, and dacite fringed by coral limestones (Meijer et al. 1983).

Constraints and speculations on the subducted portions of the PSP

The slab beneath the Ryukyu and SW Japan arcs

A significant part of the PSP has been subducted beneath the Ryukyu Arc as attested by Wadati-Benioff zone seismicity and regional scale tomographic studies (e.g., Bijwaard et al. 1998; Widiyantoro et al. 1999; Wang et al. 2008; Li and van der Hilst 2010; Wei et al. 2012, 2015). A continuous slab-like high-velocity zone (HVZ), dipping roughly 45°, extends beneath the 1200 km-long Ryukyu Arc down to ~300 km depth. This is consistent with the edge of the seismicity around 250 km according to Wei et al. (2015), whereas Bijwaard et al. (1998) imaged the high-velocity anomaly deeper, down to at least a ~500–600 km depth where it merges with the flat-lying Pacific slab. Lallemand et al. (2001) suspected that the deepest portion of the PSP slab detached during the Upper Miocene - Lower Pliocene east of the Gagua Ridge (Fig. 3). Wei et al. (2015) mentioned that the HVZ extends into the uppermost upper mantle near Taiwan, but only in V_s tomography, which could suggest that an old and more rigid part of the PSP has penetrated into the lower mantle there. That deep slab could be

connected with the detached slab east of the Gagua Ridge (see Fig. 3). Zhao et al. (2012), based on high-resolution P-wave tomography, follow the aseismic slab down to 430 km under Kyushu and 370 km under southwest Honshu, though the intraslab seismicity ends at 180 km deep. We thus estimated the downdip length of the PSP aseismic slab beneath SW Japan to be ~700 km (surprisingly, the authors measure 800–900 km). At the scale of the Ryukyu Trench, we will thus hypothesize that the subducted portion of the PSP extends ~700 km northwestward (Fig. 3). PSP imagery beneath southwest Japan is complicated by the fact that the subducting Shikoku Basin is very young (30–15 Ma) and lies above the old subducting Pacific slab. Cao et al. (2014) confirmed the observation by Zhao et al. (2012) that the P-wave high-velocity anomaly is not continuous beneath southwest Japan. It is apparently interrupted by a low-velocity anomaly north of Kyushu, extending northwestward from 80 km to greater depths. Huang et al. (2013) and Cao et al. (2014) suggest that the PSP is tearing and forms a slab window corresponding to the KPR. In their model, both the buoyancy of the KPR and the directional change in the motion of the PSP play a role in the tearing process. The corner flow produced by the subducting Pacific slab, together with its dehydration, probably heats the PSP slab from below, thus producing low-velocity anomalies (Zhao et al. 2012), and hiding the PSP slab dimensions. We will later assume that about 700 km of PSP slab has been subducted beneath the Ryukyu and southwest Japan arcs.

High intraslab velocity parallel to the trench is consistent with the main spreading direction of the WPB (see grey arrows in Fig. 3). This is also parallel to LOFZ, and may reflect fossil anisotropy (Wei et al. 2015). We thus speculate that the subducted part of the WPB, dextrally offset along the LOFZ, extends beneath the Ryukyu Arc up to the latitude of Kyushu. Indeed, if we simply offset the total width of the WPB measured between the Oki-Daito Ridge and the southernmost evidence of WPB oceanic crust (south of the Mindanao FZ), i.e., 2000 km at most, we conclude that the oceanic crust of the WPB may subduct beneath or south of Kyushu if the LOFZ extends to these latitudes. The ADO region could also extend beneath Kyushu (Fig. 3) but the steep slab observed there rather supports the subduction of an “old” oceanic crust. Park et al. (2009) and Yamamoto et al. (2013) inferred the extent of the KPR beneath the toe of the accretionary prism but the LOFZ could plausibly offset the ridge at a distance of ~200 km from the trench. The LOFZ could even offset the KPR northward by ~150–200 km (Fig. 3), providing a geometrical explanation to the following. (1) The change in the trend of the seismic slab isopachs from N20° to N80° occurs beneath the

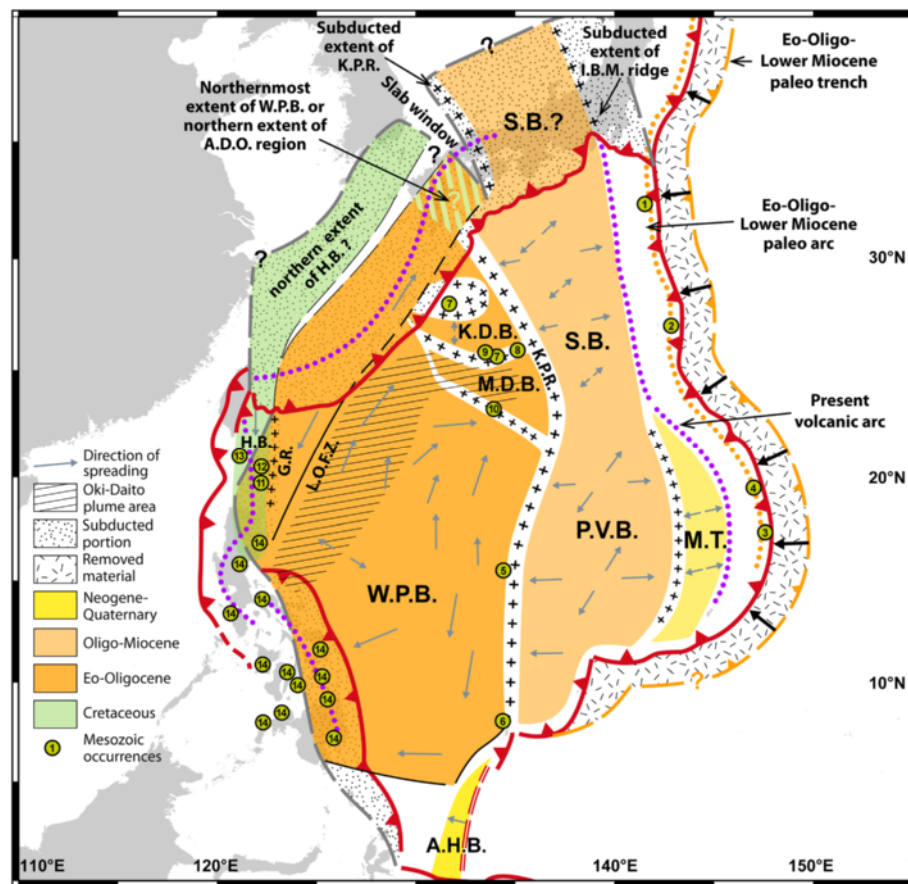


Fig. 3 Oceanic basins with their subducted parts, present and paleo-volcanic arcs and trenches of the PSP. Oceanic basins are outlined with their respective ages and the main directions of spreading (grey arrows). Purple and orange dots indicate the present and Eo-Oligo-Lower Miocene volcanic arcs, respectively, associated with the PSP, either caused by the PSP itself or another plate subducting beneath the PSP. The PSP slabs (dotted areas) have been unfolded and projected at the surface. Note that a small part of the western WPB has subducted beneath the Huatung Basin along the East side of the Gagua Ridge. The location of the Eo-Oligo-Lower Miocene trench (in orange) has been estimated based on the amount of IBM arc consumption by tectonic erosion. Mesozoic terrane occurrences in the PSP are reported with green circles, in which the numbers refer to Table 1 where more details are presented

easternmost part of Honshu and Shikoku islands instead of beneath Kyushu as would be expected if the KPR were not offset. (2) We observe two fore-arc embayments (Fig. 1) that could have resulted from KPR subduction northeast of the present ridge-trench intersection. One is east of Kyushu. It may have been caused by the non-offset ridge segment. The other is south of Shikoku. It might have been caused by the dextrally offset segment. (3) The slab window, described by Huang et al. (2013) and Cao et al. (2014), attributed to a tear along the “weak” KPR is clearly north of the location it would occupy if it were projected along a straight path with no offset. (4) The buoyant KPR might contribute to the shallow-dipping slab beneath southwest Japan (Gutscher and Lallemand 1999) if it is east of the slab window rather than west of it. In this case, the width of the subducted Shikoku Basin would be narrower (Fig. 3).

The slab beneath the Philippine Arc

The western part of the WPB is also subducting beneath the Philippine Trench. Based on the Wadati-Benioff zone, the slab is no longer than 250 km at the latitude of Mindanao Island and is shorter both north and south of there (Lallemand et al. 1998). The subduction can be traced from offshore Mortal Island in the south to Luzon Island in the north (Cardwell et al. 1980). The short length of the subducted slab as well arc volcanism ages support a Pliocene age for the start of PSP subduction offshore Mindano. The subduction then propagated both northward and southward (Lallemand et al. 1998). In detail, that region underwent several episodes of subduction, collision, and terrane accretion as depicted by many authors (McCaffrey et al. 1980; Moore and Silver 1982; Hall 1987; Rangin et al. 1999; Pubellier et al. 2004; Hall and Spakman 2015).

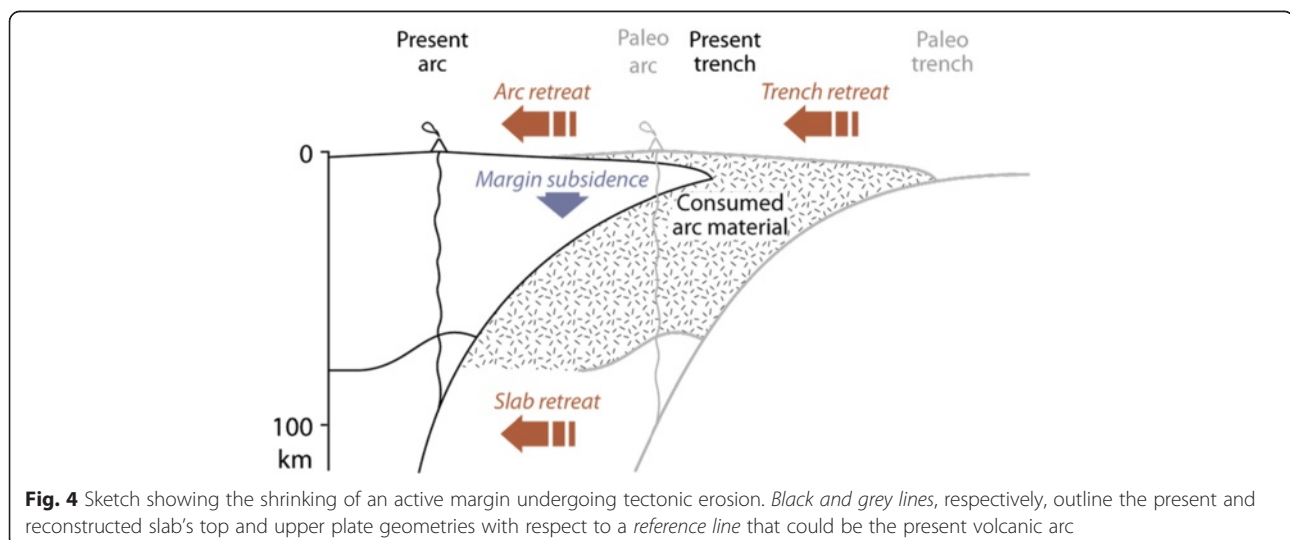
Shortening and even subduction of the WPB probably occurred north of the Philippine Trench along the East Luzon Trough as attested by a negative gravity anomaly and compressional features. Lewis and Hayes (1983) proposed that subduction was active there during the Oligocene. Such short-lived subduction of the WPB along its western margin was also proposed by Deschamps et al. (1998) in the same period with an amplitude decreasing from 200 km in the south to 0 near the present intersection between the Gagua Ridge and the Ryukyu Trench (see Fig. 3). Transpression along the Gagua Ridge has been also proposed by Eakin et al. (2015) but during the lower Miocene rather than the Oligocene.

Estimating margin loss by tectonic erosion along the IBM Arc

Convergent margins are generally the locus of mass transfers from one plate to the other across the plates' interface. Accretion of crust or sediment may occur, contributing to a net growth of the upper plate, but sedimentary or crustal erosion (also called tectonic erosion) is also possible, resulting in a net loss of upper plate material. Von Huene and Scholl (1991) estimated that about half of the convergent margins were accreting and half were erosive. There are two main indicators for a net global loss of crust at a convergent margin: a significant fore-arc subsidence increasing trenchward together with a landward retreat of the active volcanic front (von Huene and Lallemand 1990; Lallemand 1995). Most of the time, the two indicators are observed to be simultaneous (Fig. 4). Since the PSP is subducting along its western boundaries, we will focus on its eastern limit, which includes the IBM, Yap and Palau margins, to determine whether the plate has grown or shrunk.

IBM arc-trench system

The trench-arc distance varies from ~200 km along the Izu-Bonin to less than 200 km in the Mariana subduction zones. The Pacific slab always dips more than 50° and can even reach ~90° beneath the central Mariana Arc. The Pacific plate carries numerous seamounts (Fig. 1) into the subduction zone all along that trench (Fryer and Smoot 1985) favoring the dismantlement, weakening and subsequent consumption of the margin's front (e.g., Lallemand et al. 1994, Dominguez et al. 1998, von Huene et al. 2004). As a consequence, the trench inner slope is steep and no sediment accretion is observed at the toe of the margin. Instead, basement rocks have been drilled close to the trench and even serpentinite mud volcanoes have been observed all along the Mariana fore-arc (Natland and Tarney 1981; Bloomer 1983; Fryer 1992; Lagabrielle et al. 1992b; Fryer et al. 1999). The diversity of metabasic rocks found in one of these seamounts can be explained by recycling of fore-arc materials through tectonic erosion and subduction of the fore-arc (Fryer et al. 2006). The IBM margin has been classified as one of the most erosive margins in the world (e.g., von Huene and Scholl 1991; Lallemand 1995). Subsidence of the fore-arc of more than 2 km since ~40 Ma in the Izu-Bonin fore-arc and ~24 Ma in the Mariana fore-arc has been documented thanks to benthic foraminifers contained in sediment cores that are able to record paleo-bathymetry (Kaiho 1992; Lagabrielle et al. 1992a). In parallel research, Mitchell et al. (1992) reported volcanic rock outcrops from 17 to 41 Ma at a distance of ~50 km from the Izu-Bonin Trench, indicating that the volcanic front has retreated by at least 150 km during the last 41 Ma (Fig. 4). A similar history is documented in the Mariana fore-arc with gabbros and 30–45 Ma volcanic rocks outcropping near the trench, supporting a retreat of the volcanic arc



of at least ~150 km (Hussong and Uyeda 1981; Bloomer 1983) (Fig. 4). Johnson et al. (2014) analyzed plutonic rocks from two dredge sites along a normal fault scarp off Guam very close to the trench. These rocks were created from boninite parental magmas that were modified into tonalites and are now part of a midcrustal layer. Their exposure near the trench is another evidence of tectonic erosion processes.

Volcanic arc retreat sometimes results from a shallowing of slab dip, but in this case, and especially in the Mariana Arc, the present-day slab dip is almost vertical. Assuming that it has been continually vertical, which is unlikely, the margin's front has retreated by at least 150 km through consumption of the material constituting the lower section of the margin following a mechanism depicted in Lallemand et al. (1992) or Lallemand (1998). This appraisal is extremely conservative, and a more realistic estimate would consider the margin's retreat to be at least 200 km because the slab dip was probably shallower in the past and because the oldest record of arc activity recovered in the fore-arc is younger than the presumed arc initiation at ~52 Ma (see section 1.5; Ishizuka et al. 2011a). By extension, the whole margin, from southern Mariana to northern Japan has been severely affected by tectonic erosion processes during the Neogene and maybe earlier (Lallemand 1995, 1998). The Eo-Oligo-Lower Miocene paleo-arc is located a very short distance from the trench, varying from a few kilometers in the Mariana fore-arc to 90 km in northern Japan where the volcanic arc retreated by ~210 km during the last 23 Ma. This retreat occurred simultaneously with a subsidence of the margin by more than 6 km (von Huene and Lallemand 1990). We can thus extend the eastern boundary of the Paleogene PSP by at least 200 km eastward (Fig. 3).

Southernmost Mariana and Yap trench systems

The southernmost Mariana Arc and Yap segments have undergone a different story because the southernmost Mariana arc-trench system stretched E-W in response to the arc indentation of the Caroline Ridge at ~23 Ma, the spreading of the Parece Vela Basin and the Mariana Trough (Fujiwara et al. 2000; Martinez et al. 2000; Stern et al. 2013). The activity of the Yap Arc ceased after collision with the Caroline Ridge (McCabe and Uyeda 1983) and strong tectonic erosion narrowed the trench - paleo-arc distance to ~50 km. As a consequence, lower crustal and even upper mantle sections of the PSP are exposed on the inner slope of the Yap Trench (Fujiwara et al. 2000). Significant arc consumption occurred there at least during the Neogene.

This observation is extremely important for correctly interpreting the petrological and chronological data collected along that arc.

Evidences for Jurassic and Cretaceous relics of the proto-PSP

Most of the PSP consists of Eocene or younger magmatic or sedimentary rocks formed or deposited after its inception, but from the early beginning of PSP exploration, several evidences of older crust have been reported (see Table 1), attesting to the existence of a Jurassic to Early Cretaceous "proto-PSP" (e.g., Ingle et al. 1975; Matsuda et al. 1975; Hussong and Uyeda 1981).

We distinguish between two types of Mesozoic rock occurrences in the PSP depending on their location within, or along, the edges of the PSP. Some were observed in a central position like the ADO region. They necessarily belong to the proto-PSP. Others were observed near the plate boundaries, i.e., along the IBM arc or Huatung Basin. They might have been accreted after the PSP formed.

Starting with the ADO region, Matsuda et al. (1975) reported Late Cretaceous K-Ar ages for whole-rock hornblende tonalites and basalts dredged from the Amami Plateau (7 in Table 1 and Fig. 3). Hickey-Vargas (2005) performed $^{40}\text{Ar}/^{39}\text{Ar}$ dating on those tonalites and obtained an Early Cretaceous age (~115–118 Ma) that could be attributed to subsequent reheating or another Ar-loss event. Based on the geochemical characteristics of the basalts, she concluded that the Early Cretaceous subduction zone that formed the Amami Plateau might have been the site of slab melting compatible with the subduction of a young, hot plate at that time. The Amami tonalites were probably formed by fractional crystallization from the basaltic magma or partial melting of basaltic arc crust. Mizuno et al. (1978) and Okino and Kato (1992) mentioned that the Daito and Oki-Daito ridges (10 in Table 1 and Fig. 3) consisted of pre-Eocene basement rocks overlain by middle Eocene sedimentary rocks. More recently, andesites dredged on the Daito Ridge (8 and 9 in Table 1 and Fig. 3) returned $^{40}\text{Ar}/^{39}\text{Ar}$ ages of ~116–119 Ma (Ishizuka et al. 2011a). Tani et al. (2012) revealed after analyzing samples collected during a diving cruise conducted in 2010 that the ADO region dominantly exposed deep crustal sections of gabbroic, granitic, and metamorphic rocks with possible continental affinities. Jurassic to Cretaceous zircon U-Pb ages have been obtained from the plutonic rocks (7 in Table 1 and Fig. 3).

Today, the KPR occupies a central position but it constituted the proto-IBM arc and rear-arc before the opening of the Shikoku and Parece Vela basins. As discussed in section 3, tectonic erosion processes have consumed the margin's front, at least during the Neogene. We thus consider that Cretaceous ages obtained from a volcanic apron west of the KPR (Ingle et al. 1975; 5 in Table 1 and Fig. 3) or in its southernmost part (Ishizuka et al.

Table 1 List of samples collected in the PSP providing evidence for a pre-Eocene proto-PSP

Region (see n° in Fig. 3)	Sample location	Sample nature and dated material	Estimated age	Source
Izu-Bonin fore-arc				
1	Dive samples (YK11-07) 32° 14' N–141° 39' E	Andesite, diorite	Cretaceous ~100 Ma (U-Pb) + Paleozoic and Proterozoic detrital zircons	Tani et al. 2012 Tani and Myojinsho 2011
2	Sample 6 K1152 27° 19' N– 143° 00' E	Two pillow basalt lavas beneath gabbro	Middle Jurassic 159.4 ± 0.9 Ma ($^{40}\text{Ar}/^{39}\text{Ar}$)	Ishizuka et al. 2011b Ishizuka et al. 2012
Mariana fore-arc				
3	Site 460, DSDP 60 17° 40' N–147° 35' E	Pebble conglomerate with Calpionella alpina	Late Jurassic to Early Cretaceous (~130–140 Ma)	Azéma and Blanchet 1981
3	Site 460 17° 40' N–147° 35' E	Reworked radiolarians	Upper Cretaceous (Campanian, ~72–83 Ma)	Hussong and Uyeda 1981
3	Site 461 17° 46' N–147° 41' E	Reworked radiolarians	“Cretaceous” (~65–135 Ma)	Hussong and Uyeda 1981
4	Dredges 19° 37' N–147° 04' E	Radiolarian cherts	Lower Cretaceous Valanginian ~97–112 Ma and Albian 131–138 Ma	Johnson et al. 1991
4	Dredges 19° 37' N–147° 04' E	Foraminifers in cherts	Lower Cretaceous (Aptian to Albian ~96–113 Ma)	Johnson et al. 1991
4	Dredges 19° 37' N–147° 04' E	MORB metabasalt	Upper Cretaceous 85 Ma (K-Ar)	Johnson et al. 1991
4	Dredges 19° 37' N–147° 04' E	Highly metamorphosed alkali basalt	Upper Cretaceous 71 Ma (K-Ar)	Johnson et al. 1991
WPB				
5	Site 290 east of KPR 17° 44' N–133° 28' E	Reworked foraminifers in a volcanoclastic apron	Cretaceous	Ingle et al. 1975
KPR				
6	Dredge in southernmost part of KPR on a ridge between Palau Trench and KPR	Mafic schists of amphibolite to greenschists facies	May have similar age and origin to Cretaceous Daito Ridge	Ishizuka et al. 2012
ADO region				
7	Dive samples (YK10-04) 25° 54' N–133° 54' E 27° 53' N–131° 51' E 27° 56' N–132° 02' E	Gabbroic, granitic and metamorphic rocks	Jurassic to Cretaceous magmatic zircons	Tani et al. 2012 Tani 2010
Amami Plateau				
7	Dredge 11-17-1 28.07° N–131.63° E	Hornblende tonalite	Lower Cretaceous 69.5 Ma (K/Ar)–117.0 ± 1.1 Ma ($^{40}\text{Ar}/^{39}\text{Ar}$)	Matsuda et al. 1975 Tokuyama 1985 Hickey-Vargas 2005
7	Dredge 11-17-5 28.08° N–132.02° E	Hornblende tonalite	Lower Cretaceous 75.1 Ma (K/Ar)–115.8 ± 0.5 Ma ($^{40}\text{Ar}/^{39}\text{Ar}$)	Matsuda et al. 1975 Tokuyama 1985 Hickey-Vargas 2005

Table 1 List of samples collected in the PSP providing evidence for a pre-Eocene proto-PSP (*Continued*)

7	Dredge 11-9-33	Basalt	Upper Cretaceous 82.4 and 85.1 Ma (K/Ar)	Matsuda et al. 1975 Tokuyama 1985
<i>Daito Ridge</i>				
8	Short core or dredge ~25.9° N–135.2° E	Andesite	Lower Cretaceous 116.9 ± 0.9 Ma ($^{40}\text{Ar}/^{39}\text{Ar}$)	Ishizuka et al. 2011a
9	Short core or dredge ~26.0° N–133.1° E	Andesite	Lower Cretaceous 118.9 ± 0.4 Ma ($^{40}\text{Ar}/^{39}\text{Ar}$)	Ishizuka et al. 2011a
<i>Okai-Daito Ridge</i>				
10	Dredge (unknown exact location)	Alkali basalt	Limit Upper Cretaceous-Paleocene 65 Ma	Okino and Kato 1992
<i>Huatung Basin</i>				
11	Dredged sample RD19 20.40° N–121.47° E	Gabbro	Lower Cretaceous 121.2 ± 4.6 Ma ($^{40}\text{Ar}/^{39}\text{Ar}$)	Deschamps et al. 2000
12	Dredged sample RD20 21.49° N–122.69° E	Gabbro	Lower Cretaceous 123.2 ± 7.8 Ma ($^{40}\text{Ar}/^{39}\text{Ar}$)	Deschamps et al. 2000
13	YehYu Creek sample on Lanyu Island 22° N–121.55° E	Red chert float radiolarian	Lower Cretaceous (Barremian ~113–117 Ma)	Deschamps et al. 2000
<i>Philippine archipelago</i>				
14	Outcrops	Ophiolitic suites	Cretaceous	Yumul 2007

2012; 6 in Table 1 and Fig. 3) are representative of the proto-PSP prior to its isolation.

Closer to the trench, Mesozoic rocks have also been discovered along the IBM fore-arc since Deep-Sea Drilling Project (DSDP) expeditions in 1978 (Hussong and Uyeda 1981; Bloomer 1983). Late Cretaceous reworked radiolarians and Upper Jurassic Calpionellids were observed in cherts drilled at sites 460 and 461 in the Mariana fore-arc (Azéma and Blanchet 1981; 3 in Table 1 and Fig. 3). At distances of ~200 km north of these sites and more than 50 km arcward of the trench (4 in Table 1 and Fig. 3), dredged cherts, mafic and intermediate lavas, and intrusive rocks documented the presence of an ophiolite suite interpreted by Johnson et al. (1991) as accreted fragments of a Cretaceous oceanic plate. The volcanic rocks provided Late Cretaceous K-Ar ages, while the radiolarians and foraminifers supported an early Cretaceous age (~97–138 Ma). More recently, volcanic rocks have been also sampled in the northern part of the IBM fore-arc. East of the Bonin Islands, pillow-lavas were dated at 159.4 ± 0.9 Ma ($^{40}\text{Ar}/^{39}\text{Ar}$, Ishizuka et al. 2011b; 2 in Table 1 and Fig. 3), while andesites and diorites east of Aogashima Island gave Cretaceous (~100 Ma) U-Pb magmatic ages and abundant Paleozoic to Proterozoic detrital zircons (Tani et al. 2012; 1 in Table 1 and Fig. 3). Such “continental” zircons, together with the Indian MORB-like isotopic characteristics of the Jurassic lavas in the Bonin islands region, are substantial evidences for a proto-PSP origin in the vicinity of a continent (Tani et al. 2012; Ishizuka et al. 2012), rather than the allochthonous accreted fragments of Pacific origin proposed in the Mariana fore-arc by Bloomer (1983) or Johnson et al. (1991).

As discussed in section 1.2, the age of the Huatung Basin is controversial but we cannot ignore new datings obtained by Deschamps et al. (2000) on gabbros dredged in 1980 during a Research Vessel Vema cruise (Mrozowski et al. 1982). The two dredge sites are located on basement highs west of the Gagau Ridge (11 and 12 in Table 1 and Fig. 3). They provided Early Cretaceous $^{40}\text{Ar}/^{39}\text{Ar}$ ages of ~115–131 Ma. Furthermore, ~113–117 Ma radiolarians were discovered in several red chert float samples, probably collected during the Tungchin andesitic eruption on Lanyu Island (13 in Table 1 and Fig. 3). These cherts likely belonged to the former oceanic plate (Huatung Basin) on which the Miocene volcanic arc was built (Deschamps et al. 2000). Hickey-Vargas et al. (2008) analyzed the two gabbros and concluded that they have an Indian MORB Hf-Nd isotopic signature, with Pb isotope ratios intermediate but closer to Pacific MORBs than the WPB, which has a strong Indian Pb isotope signature. For comparison, sources for igneous rocks from the Amami Plateau, thought to be an Early Cretaceous island arc, have Indian Hf and Nd (presubduction mantle)

and Pacific Pb (subduction component) isotope characteristics (Hickey-Vargas et al. 2008).

Except in its northernmost part (north Luzon and the Batan islands), the Philippine archipelago does not belong to the present-day PSP, but before the inception of the Philippine Trench in the Late Miocene, it can be considered to have been part of the PSP. The Philippine region has a composite basement containing numerous ophiolite complexes, most of them being of supra subduction origin (Pubellier et al. 2004). Based on gravity data, these authors suggested that the present Miocene volcanic arc in Luzon lies directly on the southern extension of the Huatung Basin. Yumul (2007) provided a compilation of ophiolites all along the mobile belt. Except in a few places, like Zambales or Amnay, most of them are Jurassic to Cretaceous in age (14 in Table 1 and Fig. 2). He proposed a zonation from Late Cretaceous amphibolites with quartz-albite metamorphic soles distributed mainly in the eastern belt, then Early to Late Cretaceous dismembered ophiolites with mostly tectonic melanges in the central belt and finally Late Cretaceous to Oligocene ophiolitic complexes in the western belt. Such a zonation is discussed by Tani et al. (2015), who obtained new zircon ages from the northern ophiolite belts. They showed these to be Eocene in age and thus probably genetically associated with the WPB opening. As for the southern ophiolites, in Cebu, Lagonoy and Dinagat, gabbroic and leucocratic rocks associated with the ophiolitic complexes are Jurassic to Late Cretaceous in age (~90–200 Ma). Tani et al. (2015) then concluded that these Mesozoic ophiolites from the southern Philippines may potentially be correlated with the Mesozoic arc and ophiolitic rocks of the ADO region before the WPB opening.

Early Eocene isolation of the PSP through subduction initiation along the proto-IBM Arc

The geodynamic context of the proto-PSP isolation has long been a matter of debate. The orthogonality of the CBF rift with the KPR was first interpreted as an indication of plate entrapment. It was proposed that the WPB resulted from a segment of the Kula-Pacific Ridge and associated oceanic crust being trapped in the Middle Miocene (Uyeda and Ben Avraham 1972; Uyeda and McCabe 1983; Hilde and Lee 1984). A variant of this model was later proposed by Jolivet et al. (1989), who posited that a piece of the North New Guinea/Pacific Ridge was trapped during the Middle Eocene. Lewis et al. (1982) first proposed a back-arc origin for the formation of the WPB behind the East Mindanao-Samar Arc. Other authors have also suggested that the WPB opened as a back-arc basin either behind the East Mindanao-Samar Arc (Rangin et al. 1990; Lee and Lawver 1995) or the proto-IBM Arc (Seno and Maruyama 1984) but with

poor constraints on the timing and kinematics of these events. Based on considerable geological and paleomagnetic constraints, especially those imposed from ocean drilling data and the Philippine islands, Hall et al. (1995a,b,c) were the first to propose that the proto-PSP was isolated between two opposing-vergent subduction zones, and then rotated clockwise while migrating north from its initial position near the equator. Hall (2002) later refined the plate tectonic evolution of the whole western Pacific region. Deschamps and Lallemand (2002) used Hall's model when providing more details on the WPB spreading processes. In the earlier stages of WPB opening, they have accounted for the presence of a plume (Macpherson and Hall 2001) in the generation of the Benham Rise, Urdaneta Plateau, numerous overlapping spreading centers and ridge jumps. The existence of this plume, called the "Oki-Daito" plume by Ishizuka et al. (2013), is supported by numerous petrologic (e.g., Hickey-Vargas 1998) and structural (e.g., Deschamps and Lallemand 2002; Deschamps et al. 2008) studies. The updated ages and relative symmetry of the WPB opening have allowed us to refute the "entrapment model" and definitely consider the WPB as having been formed between two facing subduction zones, the East Mindanao-Samar, also called the proto-Philippine, and the proto-IBM.

Constraints on the age of PSP inception

The age of the shift from proto-PSP to PSP is reasonably constrained if we accept that the inception of the IBM subduction establishes the birth of the PSP. However, this point needs to be clarified because, if most authors agree that the inception of IBM subduction marked the onset of the PSP, there is still a debate about the best marker for subduction initiation. Ishizuka et al. (2011b) relates the onset of fore-arc basalt (FAB) magmatism, at 51–52 Ma with the age of initiation of (Pacific) slab sinking, because they are the oldest lavas present on the arc that are genetically linked with the magma suite observed from the Mariana to the Izu-Bonin fore-arc. From a petrological point of view, most authors (e.g., Stern and Bloomer 1992; Stern et al. 2012; Reagan et al. 2010; Ishizuka et al. 2011b) consider that the FABs erupted in a fore-arc position close to the paleo-trench, but we proposed in section 3 that ~200 km of the margin's front has been removed since the onset of subduction, putting the FABs at a significant distance from the incipient trench at the time of their eruption. This initial setting was recently confirmed during IODP Exp. 351 with the discovery of rocks similar to the FABs in the Amami-Sankaku Basin, ~250 km west of the present-day trench (~400–450 km west of the paleo-incipient trench), which account for the Shikoku Basin opening (Fig. 2, Ishizuka et al. 2015). The seismic structure of the

KPR indicates that the igneous basement of the Amami-Sankaku Basin, characterized by FABs, continues beneath the ridge, showing that the earliest arc lies above the FABs (Arculus et al. 2015). Moreover, FABs are MORB-like tholeiitic basalts showing little or no mass transfer from a subducting plate (Reagan et al. 2010). Consequently, the argument that the occurrence of FABs should date the inception of IBM subduction no longer stands. By contrast, the first occurrence of boninites, which required a source of volatiles, certainly marks the subduction of a hydrated lithosphere. We thus conclude that the Pacific crust reached a significant depth and metasomatized the overlying mantle at ~49–48 Ma. Considering that the transition from strike-slip movement along a transform boundary to convergent movement would unlikely produce high convergence rates (normal to the boundary) in the earlier stages, subduction inception would reasonably have started a few m.y. earlier, i.e., ~52–50 Ma is probably a minimum age. Thus we finally converge with the estimate by Ishizuka et al. (2011b) using a different marker for the onset of IBM subduction.

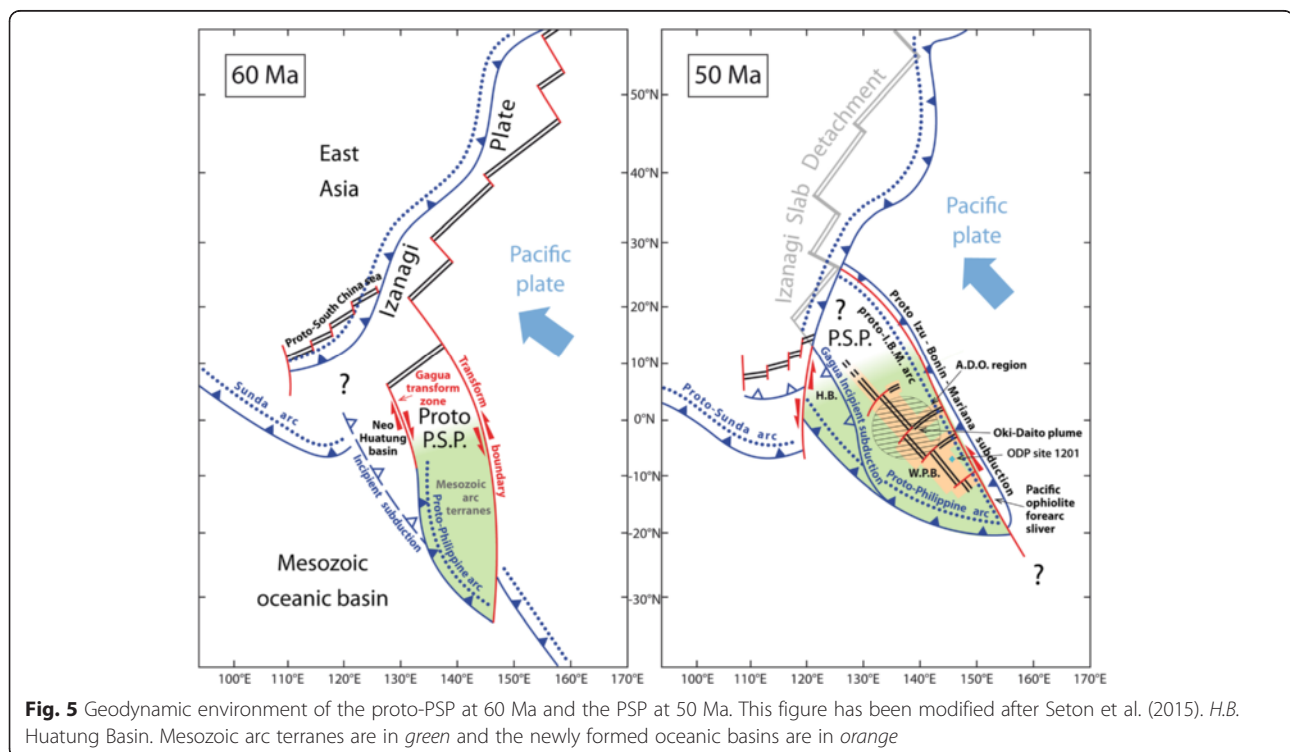
Deschamps and Lallemand (2002), based on the characteristics of the seafloor between the ODE and the Oki-Daito Ridge, estimated the rifting of the WPB to have occurred around 55 Ma with a start of spreading at ~54 Ma. This oldest age for the WPB crust has been confirmed by Doo et al. (2015). Since there was no IBM subduction when the WPB rifted and spreading began, we should consider that it started to open behind the proto-Philippine Arc under the influence of the Oki-Daito plume.

Possible trigger for PSP inception

With respect to what triggered the IBM and thus PSP inception, the following has been documented. (1) The Pacific plate changed its motion at ~43–50 Ma (Steinberger et al. 2004; Sharp and Clague 2006; Wessel et al. 2006; Whittaker et al. 2007). Consequently, the timing of this major event is much younger than the initiation of subduction along the proto-IBM trench prior to 50 Ma. Furthermore, Faccenna et al. (2012) showed that the slab pull force that grew along the IBM (and Tonga-Kermadec) subduction zone largely contributed to the Pacific plate motion change. (2) The *Oki-Daito plume* has also been proposed as another candidate for subduction initiation but again, the oldest rocks with OIB affinities, which were found in the Minami-Daito Basin, were dated ~48.4 Ma (Ishizuka et al. 2013), following the proto-IBM subduction by no less than 3 m.y. Faccenna et al. (2010) proposed that the plume might have resulted from the concomitant return mantle flow generated by the two facing subduction zones following a scenario similar to those of the present North Fiji Basin. Ishizuka et al. (2013) ruled this hypothesis out considering that the mantle 2 OIB enrichment, which persisted over 15 m.y., has the

characteristics of a very depleted mantle and does not show evidence of an enriched plume. Plume-induced subduction initiation has been invoked by Gerya et al. (2015) based on a hotter early Earth, but in a colder Earth where plate tectonics is ongoing, one may consider that the basal drag that the plume exerted below the plates might have altered the small proto-PSP motion. (3) Spontaneous subduction involving the vertical sinking of older, denser lithosphere along a transform-fracture system, as proposed by many authors (e.g., Karig 1982; Stern and Bloomer 1992; Shervais and Choi 2011) can also be ruled out. First of all, reasonable numerical models indicate that compression across the boundary fault is required (e.g., Toth and Gurnis 1998; Gurnis et al. 2004; Leng and Gurnis 2011; Leng et al. 2012) and the classical mode of failure of lithosphere under compression is localization of the deformation along conjugated shear bands dipping at an angle of $\sim 50^\circ$, then failure along one of them and thrust development (e.g., Shemenda 1992). The theoretical conditions used in numerical models to initiate a “spontaneous” gravitational sinking lithosphere require an extremely young (newly formed) upper plate separated from an old, dense oceanic plate by a very weak boundary zone and sometimes even a no-slip condition on the down-going plate (Gerya et al. 2008; Gerya 2012; Leng and Gurnis 2015). Secondly, no present or recent subduction onsets, the Philippine, New Hebrides, Flores or Wetar trench, for example, have involved such gravitational scenarios. All these recent incipient subductions

required compressional tectonic forces to initiate them (Hall et al. 2003; Lallemand et al. 2005). Thirdly, transpression along a former transform boundary between two oceanic plates may allow the younger plate to subduct beneath the older as presently observed along the Hjort Trench (Meckel et al. 2005). (4) The most reasonable trigger is the subduction of the Izanagi-Pacific (IP) Ridge beneath Asia at ~ 60 –55 Ma (Whittaker et al. 2007; O'Connor et al. 2013; Seton et al. 2015). Rapid subduction of the still-spreading IP ridge, over a vast distance, likely triggered a chain reaction of tectonic plate reorganization (Fig. 5). Ridge-push and slab pull forces acting on the Pacific plate changed drastically, resulting in a switch of the absolute motion from northwest to west-directed around 55 Ma. Since the Australian plate absolute motion also changed during the same period (Whittaker et al. 2007), subduction systems initiated along the Tonga-Kermadec and Izu-Bonin-Mariana fracture zones where convergence localized (Hall et al. 2003). The potential giant Izanagi slab detachment is thought to have changed not only plate-driven forces such as slab pull or ridge-push, but also affected the sub-Pacific and sub-East Asia mantle flow. As such, this mega-event could have also contributed to the occurrence of the Oki-Daito mantle plume accompanying the newly formed IBM subduction zone. The mechanism would reasonably be a return flow triggered by a combination of factors including the two facing subduction zones (Faccenna et al. 2010) and the



sinking of the detached Izanagi slab on another side (see Fig. 5).

Reconstructed PSP paleo-kinematics

Based on convergent paleomagnetic results, there is a relative consensus on the paleolatitude of the proto-PSP near the equator (Fig. 5, Hall et al. 1995a,b,c). For many years, interpretation of paleomagnetic results has been controversial because declination data have been obtained only from the eastern margin, where subduction-related tectonic processes may have caused local, rather than plate-wide, rotations (Hall et al. 1995c). Based on ODP site 1201, drilled ~100 km west of the KPR and ~450 km north of the CBF rift, Richter and Ali (2015) produced new data on the PSP drift since the Middle Eocene. They have analyzed the WPB basaltic basement (~47 Ma) and its overlying sedimentary sequence. Paleolatitudes derived from the sedimentary sequence support the model of northward movement of the plate since the Eocene. The basaltic basement lay at $7 \pm 5^\circ$ S in the Middle Eocene (see ODP site 1201 on Fig. 5, to be compared with $19^\circ 18'$ N today). The paleolatitudes determined from site 1201 are consistent with northward movement of the plate as predicted by Hall (1997) or Queano et al. (2007) and as determined by earlier paleomagnetic investigations on drill cores or from onland formations. The PSP decelerated between 50 and 20 Ma with a minimum plate movement at 20 Ma. Any rotational component of the plate is hard to extract since the deep-sea cores are not oriented and most azimuthally oriented paleomagnetic data are from plate boundary zone islands, potentially subject to local rotations. Based on drill cores obtained in the ADO region, Yamazaki et al. (2010) considered that most of the northward shift was accomplished between about 50 and 25 Ma with very little northward movement after 15 Ma. They also present a model in which the PSP rotated 90° clockwise between 50 and 15 Ma around an Euler pole near 23° N 162° E. Deschamps and Lallemand (2002) used Hall's model, which posit a 50° clockwise rotation of the PSP with southward translation between 50 and 40 Ma, no significant movement between 40 and 25 Ma and 40° clockwise rotation between 25 and 0 Ma for the southern part of the plate, and clockwise rotation and northward movement between 50 and 0 Ma for the northeastern part. Despite large uncertainties in rotational motions, they were able, using that model, to account for tectonic interactions with the Huatung Basin along the eastern margin during the Eocene.

Seton et al. (2015), assuming that the Izanagi slab detachment beneath East Asia was responsible for major plate reorganization, have examined the consequences on mantle flow in the Pacific region between ~53 and 47 Ma. Their geodynamic calculations, compared with

tomography data, let them argue that the flow changed from southward before 60 Ma to north-northeastward after 50 Ma, corroborating the paleolatitudes given by paleomagnetic data (Hall et al. 1995c, Richter and Ali 2015).

Cretaceous terranes origin

Around ~60–50 Ma, the proto-PSP was located in the southern hemisphere (Richter and Ali 2015). A wide Mesozoic oceanic basin, called the New Guinea ophiolite back-arc basin, separated the Australian continent from the proto-PSP (Hall 2002; Pubellier et al. 2003). Such a paleo-basin is a good candidate for the origin of Early Cretaceous Huatung Basin rocks with Indian MORB characteristics (Hickey-Vargas et al. 2008). A strip of that New-Guinea Basin was trapped along the complex transform boundary that separated the Indonesia Arc from the East Mindanao-Samar Arc (Fig. 5, Deschamps and Lallemand 2002). Similarly, the Daito ridges show strong petrological, chronological, and geochemical affinities with the southern Philippine ophiolites (Tani et al. 2015). These terranes probably result from the splitting of the same proto-Philippine island arc region at the early stages of WPB rifting under the influence of the Oki-Daito plume (Fig. 5).

Plume-ridge interaction, boninites' and FABs' geodynamic setting

A key point in discussing the early stages of the PSP and IBM subduction concerns the petrologic and geodynamic significance of the boninites, which were initially discovered on the Bonin Islands and then at several locations along the IBM fore-arc up to Palau Island. Boninitic magmatism represents a distinctive style of subduction-related magmatism, thought to result from melting strongly depleted mantle that is variably metasomatized by slab-derived fluids or melts (Crawford et al. 1989; Pearce et al. 1992). Boninites are therefore a rare type of subduction-related magma. They are much richer in H_2O , and require much more refractory sources than normal island arc suites. Their genesis requires a depleted mantle peridotite, a source of (C–O–H) volatiles and an abnormally high geothermal gradient in relatively shallow levels of the mantle wedge.

Based on present-day fore-arc magma records, they are often referred to as suprasubduction-zone ophiolites. Several petrologists (Whattam and Stern 2011; Stern et al. 2012) claim that their eruption follows the “subduction initiation rule” in which the magma source changes progressively in composition due to the combined effects of melt depletion and subduction-related metasomatism. Magmas progress from early decompression melts of unmodified fertile mantle to yield “fore-arc” basalts to younger hydrous flux melts of depleted mantle that has been strongly

modified by subduction-related fluids to yield late high-Mg andesites and boninitic lavas. Finally, normal arc volcanics (tholeiites) cap the subduction initiation sequence (Stern et al. 2012). Metcalf and Shervais (2008), Reagan et al. (2010) and Ishizuka et al. (2011b) described the genetic link in the magmatic suite constituting the “present-day fore-arc” ophiolites along the IBM subduction zone starting from the ~50–52 Ma so-called FABs, then the ~44–48 Ma boninite andesites and their differentiates and/or the ~44–45 Ma, transitional high-Mg andesites, finally topped by the post 44 Ma arc tholeiites and calc-alkaline rocks.

Macpherson and Hall (2001) observed that boninitic magmatism is not recognized in younger subduction zones of similar dimensions (Crawford et al. 1989) like, for example, the Hjort-Macquarie-McDougall-Puysegur incipient subduction (Meckel et al. 2005). Macpherson and Hall (2001) suggest an additional tectonic or thermal factor which could influence the generation of boninites and posit that a mantle plume influenced the magmatic and tectonic evolution of the western Pacific since the Middle Eocene. Occurrences of recent boninites are reported from the North Tonga Ridge (e.g., Falloon et al. 1989), the Valu Fa Ridge in the Lau Basin (Kamenetsky et al. 1997), and the southern New Hebrides arc (Monzier et al. 1993). Deschamps and Lallemand (2003) observed that the geodynamic setting was the same in each of these examples, i.e., eruptions occurred at the intersection between a back-arc spreading ridge and a volcanic arc. Moreover, based on their reconstruction of the WPB in 2002, they noted that the IBM boninites roughly coincided with the intersection between back-arc rifts or spreading centers and the proto-IBM arc. Such geodynamic context is exceptional because most of the time, the rift or back-arc spreading center parallels the arc. Specific conditions are needed to make spreading centers and arcs intersect. Among these is the influence of a mantle plume that disorganizes the mantle flow in the back-arc region.

It is highly probable that the FABs that underlie the boninites not only occur in the IBM fore-arc, but also constitute part of the basement of the short-lived back-arc basins in the ADO region (Ishizuka et al. 2015; Arculus et al. 2015). The occurrence of both FABs and boninites in a modern fore-arc setting is the result of arc consumption by tectonic erosion after their early emplacement. These lavas were initially emplaced far from the incipient trench either in a rear-arc position like the Valu Fa lavas, or in a stretched arc at the termination of a rift like in northern Tonga or in the southern New Hebrides. In section 7, we propose a scenario for the early stages of IBM subduction and associated magmatism that satisfies all previous constraints.

Toward an integrative tectono-magmatic model of the IBM subduction zone

Several paleoreconstructions of PSP evolution have been proposed (e.g., Uyeda and Ben Avraham 1972; Karig 1975; Seno and Maruyama 1984; Jolivet et al. 1989; Hall et al. 1995a; Lee and Lawver 1995; Hall 2002; Deschamps and Lallemand 2002; Pubellier et al. 2004; Honza and Fujioka 2004; Sdrolias and Müller 2006; Gaina and Müller 2007; Zahirovic et al. 2014; Seton et al. 2015). In this study, we do not pretend to provide a complete update of PSP tectonic evolution but rather highlight key phases, especially during the early stages of the evolution and IBM subduction initiation, and connect them into an integrated model of evolution that satisfies the existing geological, geochemical, geochronological and mechanical constraints.

Proto-PSP environment

The proto-PSP consisted of Mesozoic terranes of various origins including island arcs such as the Daito ridges and some Philippine ophiolitic complexes (Hall 2002; Hickey-Vargas et al. 2008; Tani et al. 2015). At the time of PSP inception (Fig. 5), there is no clear evidence whether the Huatung Basin, which probably originated from a larger Mesozoic back-arc basin north of Australia, was already trapped or not. If not, the docking of a strip of that basin through the jump of the plate boundary from the former Gagua fracture zone to another subparallel fracture/transform fault occurred soon because the trapped piece of Huatung crust moved northward with the PSP from the very beginning (Deschamps and Lallemand 2002). The simplest configuration that can be derived from the above observations is that the Mesozoic terranes (~60 m.y. fossil arcs at that time) lay behind the active proto-Philippine Arc. These fossil arc terranes were separated from the Pacific oceanic plate by a transform fault that allowed the Izanagi and Pacific plates to subduct beneath East Asia (e.g., Seton et al. 2015).

Consequences of Izanagi slab detachment

At the end of the Paleocene (~60–55 Ma), the whole of East Asia underwent the subduction of the Izanagi-Pacific active spreading ridge (Whittaker et al. 2007; Zahirovic et al. 2014). Since the ridge segments were subparallel to the former margin, the Izanagi plate detached as a whole, resulting in a drastic change in mantle flow dynamics across the whole West Pacific and East Asia (Seton et al. 2015). This major tectonic event modified not only the driving forces of the Pacific plate and thus the plate motion direction, but also those of the surrounding plates through a change in mantle flow. In such a context, convergence might have localized along the weak fracture zone separating the Pacific plate

from the proto-PSP, which was composed of Mesozoic terranes. Oblique shortening might have started as early as 60–55 Ma ago, then thrusting localized and subduction of the Pacific plate started. The cumulative downwelling action of the former proto-Philippine northeastward-vergent subduction and the mantle flow reorganization following the Izanagi slab detachment may have triggered localized mantle upwelling, further called the “Oki-Daito mantle plume” around 50–52 Ma (Fig. 5).

Impact of the Oki-Daito plume

As in the North Fiji Basin, the proto-PSP split into several “terrane” under the plume’s influence (Hickey-Vargas et al. 2008; Faccenna et al. 2010). Multiple spreading centers with various orientations ruptured the newly formed PSP, especially during the earlier stages at ~52–45 Ma. Some of them, especially in the ADO region, gave rise to short-lived oceanic basins like the Kita-Daito, Amami-Sankaku or Minami-Daito basin or the piece of the WPB north of the ODE. The plume thus likely triggered basins opening as attested by the presence of OIB dikes and sills in the Minami-Daito Basin. The fact that the FAB-like lavas drilled in the Amami-Sankaku Basin do not exhibit a plume-like geochemical signature may indicate that the spreading centers propagated outside of the plume’s area of influence. Plateaus, formed by excess magma, or seamounts with OIB signatures, developed essentially in the northern part of the present WPB, including the Minami-Daito Basin (~43–51 Ma), Oki-Daito Ridge (~48 Ma), Oki-Daito Rise (~40–45 Ma), Urdaneta Plateau (~40–36 Ma) and Benham Rise (~36 Ma) (Ishizuka et al. 2013). Typical failed rifts and propagators characterize the region between the Urdaneta Plateau and the Benham Rise (Figs. 1, 2, 3, and 5, Deschamps et al. 2008).

IBM subduction inception

The IBM subduction zone initiated in the region of the former transform boundary between the splitting Mesozoic terranes that were essentially composed of former island arcs and the Pacific plate. Since the oldest lavas that include subduction components (near the KPR-Daito Ridge intersection) are ~48–49 Ma old (Ishizuka et al. 2011a), we suppose that about 10 m.y. were necessary for the former transform boundary to turn into a subduction, as suggested by numerical models (Leng and Gurnis 2011; Leng et al. 2012). At this stage, it is important to note that subduction did not initiate along a transform fault separating a younger and an older oceanic plate as has often been described (e.g., Shervais and Choi 2011; Stern et al. 2012) but along a transform boundary between Mesozoic terranes characterized by a crust ~20 km thick (Nishizawa et al. 2014) and an

oceanic plate characterized by a ~6–7 km crust but with a lithosphere of highly variable thickness along the strike. Indeed, reconstructions by Zahirovic et al. (2014) and Seton et al. (2015) indicate that the age of the Pacific plate varies from ~0 in the north to ~80 Ma in the south of the proto-IBM arc. Sdrolias and Müller (2006) proposed a more restricted age range between 30 and 70 Ma. In any case, it is quite clear that the oceanic crust (Pacific Plate), whatever its age, will tend to subduct beneath the Mesozoic terranes that are more buoyant if the region undergoes compression. To be complete, we must admit that young oceanic basins formed during subduction initiation, so the proto-PSP ended up consisting of alternating thick Mesozoic arc terranes and newly formed oceanic crust, as also noticed by Leng and Gurnis (2015).

The mode of failure along that transform boundary is unknown, but we can learn from similar known or recent geodynamic contexts. The Australia-Pacific plate boundary south of New Zealand offers an excellent analog to the proto-IBM situation because its former purely transform boundary evolved into a series of incipient subductions from Puysegur in the north to Hjort in the south. At the latitude of the Puysegur Trench, the young Tasman oceanic basin (~10 Ma) began to subduct beneath the Campbell Plateau and Solander Trough in the Miocene (Lebrun et al. 2003). Detailed investigations have shown that the former transform fault is still active in the upper part of the Puysegur Ridge. Failure occurred on the Tasman side at a distance from the transform fault along a shallow-dipping thrust, so that the trench is now located at ~20–50 km from the fault (Collot et al. 1995). Similar observations can be done in the Hjort sector with the young Southwest Tasman Basin (~5–9 Ma) subducting beneath the Macquarie Ridge Complex, which includes the Hjort Plateau, since 6–11 Ma (Meckel et al. 2003). Inspired by these recent examples and by mechanical considerations (e.g., Shemenda 1992), we suggest that the initial thrusting occurred within the Pacific plate at a distance of a few tens of kilometers from the former transform boundary. The former transform fault probably remained active, partitioning the strain if we assume that the convergence was oblique for a long time after subduction initiated. Such geometry favors the lateral migration of a fore-arc sliver (McCaffrey 1992; Lallemand et al. 1999). In the proto-IBM case, the fore-arc sliver was likely composed of Pacific oceanic lithosphere (an ophiolitic wedge of Pacific origin). Considering that the trench-transform distance likely depends on respective lithosphere thicknesses, it may vary along-strike depending on the plate’s age. Since kinematic reconstructions, in Seton et al. (2015), for example, indicate that the age of the Pacific plate increases southward from 0 to 80 Ma, we can reasonably assume

that the Pacific fore-arc sliver will be wider in the south (Fig. 5).

At this stage, it is important to note the difference between spontaneous and forced subduction initiation in a geodynamic context that bears magmatic sources in mind. Indeed, most petrologists (Stern and Bloomer 1992; Shervais and Choi 2011; Stern et al. 2012; Ishizuka et al. 2014) consider that the asthenospheric mantle invades the gap between “sinking old” and “young buoyant” lithospheres, inducing during fore-arc spreading (FABs and boninites), but since spontaneous subduction is unlikely and “compression is the rule” (Hall et al. 2003), there is no available space for the mantle to reach the fore-arc region. Consequently, the first lavas erupted at a significant distance from the incipient trench.

Tectono-magmatic model of PSP inception, IBM subduction initiation and evolution

Chronologically, we propose the following tectono-magmatic events flow at the scale of the whole IBM system (Figs. 5 and 6):

~60–55 Ma: Izanagi slab detachment beneath the East Asia margin triggers diffuse shortening across the region accommodating the transform motion between the Mesozoic composite proto-PSP and the Pacific oceanic plate.

~55–52 Ma: mantle flow reorganization after slab detachment and interaction with local mantle convection (proto-Philippine subduction) contributed to the formation of the Oki-Daito plume.

~54–48 Ma: multiple rifts developed under the influence of the plume, splitting the composite proto-PSP into several ridges separated by short-lived oceanic basins. Those at the head of the plume were intruded by OIB sills like the Minami-Daito Basin but others still show Indian-ocean MORB signatures in their Pb isotopic composition (Ishizuka et al. 2011b; Ishizuka et al. 2013; Hickey-Vargas et al. 2013). The FABs described in the Bonin and Mariana fore-arcs belong to this group. These oceanic basins opened back of the transform fault, probably without or with little contamination from the down-going Pacific plate. One of the spreading centers, subparallel to the proto-Philippine Arc, gave rise to the oldest crust of the WPB. This was abandoned after a few m.y. through a ridge jump.

~52–48 Ma: after several m.y. of diffuse shortening across the transform boundary region, the deformation localized along a thrust fault cutting through the Pacific plate, incorporating a Pacific ophiolite into the newly formed PSP. Behind the transform fault, the PSP lithosphere stretched, creating an alternance of thick island arc remnants and thin oceanic lithosphere (FABs). The width of the Pacific fore-arc sliver was

probably larger in the south where the subducting plate was older. The subducting Pacific plate probably deformed, with undulations, because the thickness of the overriding lithosphere varied along-strike from the Mesozoic ridges to the young, thin oceanic lithosphere. The subducting Pacific crust thus reached the mantle corner sooner beneath the newly formed basins than beneath the ridges.

~49–34 Ma: the first boninites are supposed to have erupted around ~49–48 Ma, persisting for about 15 m.y (Cosca et al. 1998). Following the geodynamic setting proposed by Deschamps and Lallemand (2003), we believe that these lavas did not erupt in a fore-arc position, but rather in an incipient arc and/or rear-arc position mostly at the intersection between the spreading centers of the young oceanic basins and the juvenile arc. These are considered to be the first “arc-related lavas” because they are stratigraphically at the bottom of the volcanic suite, but they were still erupting 15 m.y. after subduction initiation. This clearly indicates, if necessary, that the conditions for their eruption include a specific geodynamic context including an arc-ridge intersection as in the Valu Fa Arc (Tonga), rather than fitting the beginning of subduction.

<45 Ma: high-Ca boninites and high-Mg andesites sometimes appear atop low-Ca boninites on some Bonin or Mariana islands (Ishizuka et al. 2006). They mark the transition between highly depleted mantle sources and the high degrees of fluxed melting at shallow levels that generated boninites and normal arc volcanism (Reagan et al. 2008). Then, typical arc tholeiites and calc-alkaline rocks formed, contributing to the growth of the IBM Arc. The delay of a few m.y. between the oldest boninites and the oldest arc tholeiites is simply explained by the time needed for the Pacific crust to reach the mantle wedge beneath the Mesozoic ridges and the thickening lithosphere associated with the FABs.

<40? Ma: tectonic erosion of the frontal margin has been documented during the Neogene and possibly as soon as the Middle Eocene (Lallemand 1995). It has been accompanied by fore-arc subsidence/consumption and volcanic front retreat (Fig. 7). At least 150 km of frontal margin has been subducted and partly recycled into arc magmatism (Bloomer 1983; Lallemand 1996, 1998). This erosional process permitted the oldest FABs, arc boninites, and tholeiites to outcrop today near the trench. Most or all of the Pacific ophiolite has probably been lost into the subduction zone. Discoveries of Mesozoic cherts and ophiolitic suites in some rare places may represent remnants of the Pacific relics trapped during subduction initiation as well as pieces of seamounts offscrapped from the down-going plate later (Johnson et al. 1991). The widespread serpentinite mud

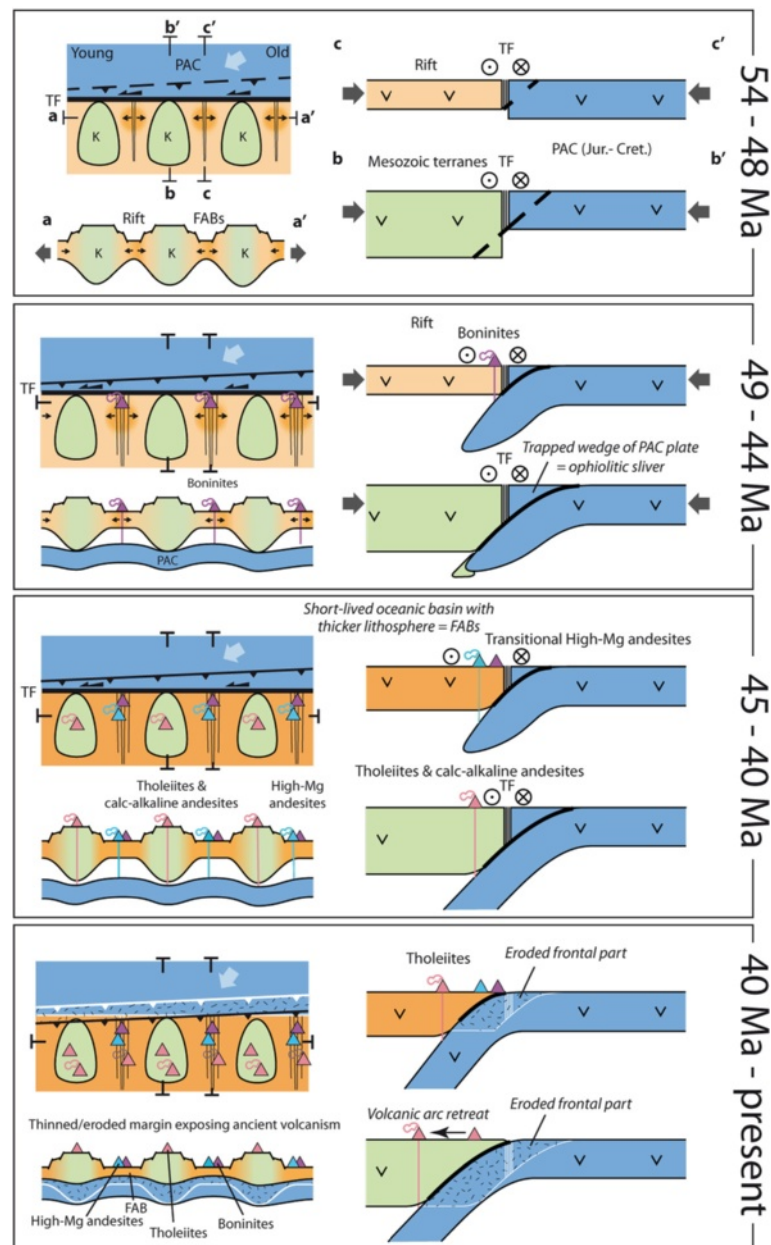


Fig. 6 Schematic tectono-magmatic model of IBM subduction initiation and early evolution. Special emphasis has been done on the geodynamic context of various magma eruptions. The white lines in stage "40 Ma - present" represent the former positions of plate boundaries or the base of the upper plate in stage "45-40 Ma". TF = transform fault, PAC = Pacific, FAB = Fore-arc basalts

volcanoes observed in the Mariana fore-arc (Fryer et al. 1985, 1999) are very interesting because they are the only ones known to be active in this context today. Fluids from the descending plate hydrate the fore-arc mantle and enable serpentinite muds to rise along faults to the seafloor (Fryer 2012). Blueschists found in the mud attest that they originate from depths as great as 25 km and their ascent is probably facilitated by sub-vertical faults cutting through the fore-arc (Fryer et al. 2006). Following our scenario, we propose that they

form along the segments of the initial paleo-transform fault zone that have been preserved from erosion. As suggested above, the Pacific sliver should have been wider in the Mariana fore-arc because it was thought to be much older than in the Izu-Bonin fore-arc. Logically, for a constant amount of tectonic erosion along the IBM fore-arc, the southern part is more likely to preserve Pacific remnants than the northern part. Such a paleo-transform is a good candidate for both a sub-vertical faulted zone and hydration, and thus

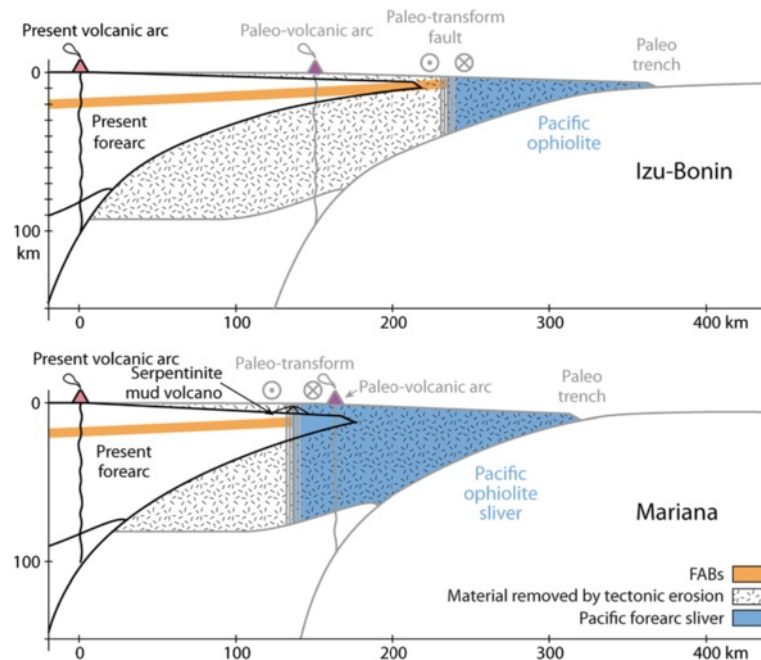


Fig. 7 Schematic cross-sections of Izu-Bonin and Mariana fore-arcs at early stage and at present. This cartoon emphasizes the differences in structural settings that might favor the rise of serpentinite mud diapirs. FABs fore-arc basalts. As in Fig. 4, black and grey lines outline the present and reconstructed slab's top and upper plate geometries with respect to a reference line that could be the present volcanic arc

serpentinite formation. If this hypothesis is correct, we should find remnants of the former Pacific ophiolite trenchward of that zone. The discoveries of Johnson et al. (1991) in the Mariana fore-arc could represent a first sign that this is the case.

Conclusions

Compiling all the data available on the composite PSP, we were able to depict the earlier stages of PSP evolution within a coherent integrative approach.

Based on our calculations, the total PSP area can be roughly estimated to cover $\sim 8.7 \cdot 10^6 \text{ km}^2$, with about 2/3 ($\sim 5.9 \cdot 10^6 \text{ km}^2$) constituting the present-day visible part of the PSP, one fourth ($\sim 2.2 \cdot 10^6 \text{ km}^2$) is already subducted beneath the Ryukyu and Philippine arcs, and the remaining $0.6 \cdot 10^6 \text{ km}^2$ has been eroded along its eastern boundary.

Based on the timing of events, the most likely trigger of PSP inception is the Izanagi slab detachment beneath the East Asia margin. Such a mega-event has been dated 60–55 Ma, which gave enough time for the paleo-transform boundary between the Pacific plate and the proto-PSP to localize the deformation, evolve into a subduction zone around 52–50 Ma and produce the first boninitic lavas a few m.y. later around 49–48 Ma. At the same time, mantle flow reorganization triggered the formation of the Oki-Daito plume, which triggered, or enhanced, the splitting of the composite Mesozoic proto-PSP into several pieces around 54–50 Ma. The

plume also controlled the formation of the long-lived WPB behind the proto-Philippine Arc, especially in its northern sector, during the first 15 m.y. of opening. The IBM trench developed within the Pacific Plate, likely isolating a fore-arc sliver composed of a Pacific ophiolite bounded arcward by a paleo-transform zone. Later, tectonic erosion of the fore-arc removed most of the Pacific fore-arc material, except in a few areas, like off the Mariana Arc, where some remnants may outcrop, as well as serpentinite diapirs that might have developed along the altered paleo-transform relics. The ophiolitic suite presently sampled in the IBM fore-arc was not formed in a fore-arc position but at a significant distance ($>50 \text{ km}$) from the incipient trench, which would account for the considerable amount of material consumed at the front of that margin. Consequently, FABs and fortiori boninites are not markers s.s. of subduction initiation. The so-called “fore-arc basalts” were formed along short-lived spreading ridges, far and out of the influence of the Pacific subduction, until the slab's crust reached the depth of dehydration, favoring the eruption of the boninites. The intersection of numerous spreading centers and the newly forming arc above a shallow slab satisfied the necessary conditions for boninite formation. Use of the word “fore-arc” to characterize these basalts and boninites is genetically not appropriate.

Further investigations and physical experiments are needed to validate our model.

Abbreviations

ADO: Amami-Daito-Oki-Daito; CBF: Central Basin Fault; DSDP: Deep-Sea Drilling Project; FABs: fore-arc basalts; FZ: fracture zone; IBM: Izu-Bonin-Mariana; IODP: International Ocean Discovery Program; IP: Izanagi-Pacific; KPR: Kyushu-Palau Ridge; LOFZ: Luzon-Okinawa fracture zone; MORBs: mid-ocean ridge basalts; NTD: non-transform discontinuity; ODE: Oki-Daito Escarpment; OIB: ocean island basalt; PSP: Philippine Sea Plate; WPB: West Philippine Basin.

Competing interests

The author declares that he has no competing interests.

Authors' information

The author, after spending the first 2 years of his PhD studies at the Ocean Research Institute - Tokyo University in the early 1980s, has focused most of his research studies on understanding subduction processes, with special emphasis on the Japan and Taiwan areas. He collaborated with Anne Deschamps during her Master's and PhD studies as well as during her post-doctoral research at JAMSTEC between 1997 and 2003. Significant results on the geodynamics of the Philippine Sea Plate arose from that intense collaborative period. Anne should have co-authored this review paper if she had still been alive.

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I dedicate this review to Anne Deschamps who left us recently and who has provided crucial constraints on the Gagua Ridge, the Huatung Basin, the Mariana Trough and the Izu-Bonin-Mariana arc. Her major contribution, however, concerns the tectonic evolution of the West Philippine Basin. In memory of her new insights, especially on the "Central Basin Fault rift," Yasuhiko Ohara and Toshiya Fujiwara have suggested naming an oceanic high fringing the CBF rift in the center of the Philippine Sea Plate the "Deschamps seamount." Their suggestion was approved unanimously by the Intergovernmental Oceanographic Commission of UNESCO in mid-October 2015. This paper's cover image shows a detailed map of the Deschamps seamount (Ohara et al. 2015) and a picture of Anne Deschamps taken in 2010 before she explored the East Pacific Rise using the 6000 m deep-sea submersible Nautilie. Thanks to Jean-Yves Royer who provided me with this photograph.

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