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Archimer
<http://archimer.ifremer.fr>**Indentation of the Philippine Sea plate by the Eurasia plate in Taiwan:
Details from recent marine seismological experiments**Serge Lallemand^{a, b, *}, Thomas Theunissen^c, Philippe Schnürle^d, Chao-Shing Lee^{b, e},
Char-Shine Liu^{b, f}, Yvonne Font^g^a Géosciences Montpellier, Université Montpellier 2, CNRS, place E. Bataillon, 34095 Montpellier cedex 05, France^b LIA ADEPT, France, Taiwan^c IRAP, Observatoire Midi-Pyrénées, Toulouse, France^d Géosciences Marines, IFREMER, Plouzané, France^e Applied Geophysics, NTOU, Keelung, Taiwan^f Institute of Oceanography, NTU, Taipei, Taiwan^g Université de Nice Sophia-Antipolis, Institut de Recherche pour le Développement, Observatoire de la Côte d'Azur, Géoazur, France

*: Corresponding author : Serge Lallemand, tel.: + 33 4 67 14 33 01 ; fax: + 33 4 67 14 36 42 ;

email address : lallem@gm.univ-montp2.frthomas.theunissen@irap.omp.eu ; Philippe.Schnurle@ifremer.fr ; leecs@mail.ntou.edu.tw ; cslu@ntu.edu.tw ;font@geoazur.obs-vlfr.fr**Abstract:**

We analyze in this study a new set of marine data including 3D local tomography, 1992–2008 relocated earthquakes and two recent multichannel seismic lines to characterize the deformation style in the collision area offshore east Taiwan. We have mapped in detail the Mohos of the converging plates as well as the subduction interface with a resolution never reached before. We show that the sharp continental subduction of the Eurasia plate, beneath the middle part of the Central Range, indents the Philippine Sea plate (PSP) as attested by intra-oceanic slicing and incipient subduction of the PSP beneath the east coast of Taiwan. The westernmost part of the PSP slab is probably experiencing a beginning of break-off as attested by NW-trending en-échelon shear zones beneath the southern slope of the southern Ryukyu arc (SRA). These en-échelon shear zones have a sinistral component favored by the “collision-free” subduction of the PSP north of 24°30'N. The down-faulting of the subduction interface forms ramps along which earthquakes clusterize. Three M7 subduction earthquakes occurred offshore Suao city along these ramps with a recurrence interval of about 40 years: 1920 M_w 7.7, 1963 M_w 7.2 and 2002 M_w 7.1 events. The 1966 M_w 6.0–7.5 earthquakes sequence likely outlines a WNW-ESE left-lateral intra-slab shear zone. The SRA upper plate accommodates the complex geometry and deformation of the subducting PSP through seismic deformation. Shallow high velocities fringing the Luzon volcanic arc (LVA) beneath the Longitudinal Valley and north of the southernmost Ryukyu forearc basins are interpreted as relics of the LVA forearc basement squeezed in the collision zone. Based on the accommodation of a large part of the convergence through shortening within the PSP and the subsequent segmentation of the shallow subduction interface, we consider that the nucleation of a $M_w \geq 8$ earthquake along the southernmost Ryukyu megathrust is unlikely.

Highlights

► We have mapped the Moho and plates interface in NE Taiwan with a resolution never reached before ► Today, the Eurasia plate indents the Philippine Sea plate in central Taiwan ► The subducting Philippine Sea plate undergoes intra-oceanic slicing and tearing ► Subduction earthquakes clusterize on ramps of the subducting Philippine Sea plate ► The nucleation of a $M_w \geq 8$ earthquake along the southernmost Ryukyu megathrust is unlikely

Keywords: Arc–continent collision ; Local tomography ; Tear, subduction ; Intra-oceanic slicing ; Slab break-off

1. Introduction

Taiwan represents one of the youngest and most-studied collisional belt in the world (Angelier et al., 1986, Angelier et al., 1990, Biq, 1972, Chai, 1972, Chemenda et al., 1997, Ho, 1986, Huang et al., 2006, Lallemand et al., 2001, Lu and Hsu, 1992, Malavieille and Trullenque, 2009, Malavieille et al., 2002, Sibuet and Hsu, 2004, Suppe, 1981, Suppe, 1984, Teng, 1990, Teng, 1996, Teng et al., 2000, Tsai, 1986, Wu, 1970, Wu, 1978 and Wu et al., 1997). The orogen results from the oblique southeastward subduction of the South China Sea (SCS) rifted margin beneath the northern prolongation of the Luzon volcanic arc (LVA) carried by the Philippine Sea plate (PSP). This view reasonably accounts for the geological observations south of 23°N (i.e., southern third of the island), but north of 23°N , singular features add complexities to this simple setting. First of all, the eastward continental subduction of Eurasia plate (EP) beneath the PSP is orthogonal with the southern

38 Ryukyu subduction system where the PSP subducts northward beneath the EP, both plates
39 interacting at depth beneath the northeastern part of Taiwan island. Secondly, the southern
40 Ryukyu arc (SRA) is presently rifting, resulting in a rapid southward migration of the arc-trench
41 system with respect to the EP (Chinese platform). A large part of this collisional area north of
42 23°N is submerged below deep waters so that its investigation requires marine facilities.
43 Tremendous work has been done since the eighties but the intensity of the deformation
44 encountered east of Taiwan is so high that the geophysical images were unclear and left to
45 contrasting interpretations (e.g., Font et Lallemand, 2009).
46 To better illuminate the structure and the microseismicity of this complex region, we have
47 conducted a series of local seismological cruises in the frame of the “Ryukyu Arc Tectonics and
48 Seismology” (RATS) Project² conducted in parallel with the regional TAIGER Project³ (Wu et al,
49 2007b). Preliminary results were published by Klingelhoefer et al. (2012) and Theunissen et al.
50 (2012a). The RATS data allowed us to relocate past seismic events in the region within a new 3D
51 velocity model, giving way to a revised geodynamic interpretation of the region.
52 Based on the new 3D velocity model and earthquakes relocation, as well as two new
53 multichannel seismic lines, the remaining questions that we address in this study are the
54 following : how does the PSP accommodate the increasing shortening from south to north
55 offshore Taiwan ? how are the eastern part of Taiwan and the SRA affected by this deformation ?
56 and what is the potential for a great subduction earthquake along the southernmost Ryukyu
57 subduction interface ?
58

59 **2- Geodynamic context and previous work**

61 **2.1 - Kinematics**

62 The Taiwan mountain belt build up over the « quasi-stationary » (with respect to hotspots)
63 subducting continental lithosphere of the EP. The buttress, against which the orogen grows by
64 addition of crustal units in the foreland, consists of the deformed volcanic edge of the overriding
65 PSP. Near 23°N, two-third of the 8 cm/yr of shortening mainly localizes along the thrust
66 deformation front near the western coast of the island and the steeply dipping thrust of the
67 Longitudinal Valley that separates the indenting northern LVA from the growing orogen (Yu et
68 al., 1997)(Fig.1). The rest of the shortening must be accommodated offshore (Malavieille et al.,
69 2002; Simoes et Avouac, 2006; Huang et al., 2010). Near 24°N, one-third or less of the
70 convergence between the EP and the PSP is accommodated on the island mainly in the
71 Longitudinal Valley. North of 24°N, GPS measurements indicate block rotations and south-
72 eastward extrusion (Yu et al., 1997; Rau et al., 2008; Hou et al., 2009; Hsu et al., 2009; Lin et al.,
73 2010) that necessitates to account for more than 8 cm/yr of convergence offshore. The most
74 prominent offshore expression of convergence north of 24°N is undoubtedly the WNW-ESE
75 trending southern Ryukyu accretionary wedge, which strikes almost parallel to the direction of
76 EP-PSP convergence (N307°, Seno et al., 1993), and thus can not be the expression of NW-SE
77 compression. Recent GPS measurements indicate a southward drift of the SRA Japanese islands
78 reaching up to 7 cm/yr in Yonaguni (Nakamura, 2004; Nishimura et al., 2004) which makes the
79 convergence in the Ryukyu trench much less oblique and implies a rapid southward trench roll
80 back (Fig.1). Such motions are very recent since the amount of N-S extension in the southern
81 Okinawa Trough (SOT) has been estimated to 80 km in the last 2 Myr, i.e., about 4 cm/yr in
82 average (Kimura, 1985; Letouzey and Kimura, 1986; Sibuet et al., 1998). One would expect N-S
83 transform zones cutting through the arc to accommodate its southward drift, especially near
84 Taiwan as searched by Lallemand and Liu (1998) or later by Lallemand et al. (1999) but most
85 marine investigations in this sense were unsuccessful except some indications in the Hoping
86 Basin (Font et al., 2001). Kinematic studies have also shown that the motion of the PSP was
87 more northward or even northeastward before 5 or 8 Ma ago (Seno and Maruyama, 1984; Hall

² The RATS Project is part of the Active Tectonics and Seismic Hazard in Taiwan (ACTS) French ANR program

³ TAIGER : TAIwan Integrated GEodynamic Research project between US and Taiwanese teams.

88 et al., 1995; Faccenna et al., 2009). Lallemand et al. (2001) have proposed that this motion
89 change contributed in the westward migration of the Ryukyu trench across a propagating tear
90 within the EP. The corollary of this is that the present SRA constituted the northern passive
91 margin of the (now subducted) SCS prior to the westward migration of the Ryukyu trench
92 (Fig.2). The former northward motion of the PSP along its western convergent boundary is
93 probably responsible for the steepness of the subducting SCS as seen on global tomographic
94 sections (Lallemand et al., 2001) despite the rapid rollback that the Manila trench encounters
95 since the change of plate motion. Indeed, highly oblique subduction zones such as Andaman,
96 West Aleutian or South-Mariana are all associated with steep slabs.
97 The revised subduction obliquity along the southern Ryukyu trench, that accounts for the
98 southward drift of the SRA, is still of the order of 50° with respect to trench normal west of
99 $122^\circ 40'E$. The revised subduction rate in the direction of convergence presently reaches 14
100 cm/yr with a trench-normal component of 9.3 cm/yr and a trench-parallel component of 10.5
101 cm/yr. Previous studies have shown that the high subduction obliquity was responsible for
102 trench-parallel stretching within the forearc basins and strain partitioning at the scale of the
103 accretionary wedge (Lallemand et al., 1999). No strain partitioning at the scale of the SRA
104 basement was evidenced yet as previously suggested by Kao et al. (1998). Instead of that, WNW-
105 ESE sinistral transcurrent faulting was evidenced during the 1966 seismic crisis that included
106 one $M_w 7.5$ earthquake and three $6.0 \leq M_w \leq 6.2$ aftershocks occurring below the south-facing
107 Ryukyu arc slope southwest of Yonaguni island (Wu, 1978) whereas dextral strike-slip faulting
108 is expected in case of strain partitioning compatible with the present subduction obliquity. All
109 four events aligned along the NW-SE left-lateral nodal plane (see Fig.1).
110

111 2.2 - Seismicity

112 The seismicity rate is extremely high on, or east and south of, Taiwan even if no historical $M_w 8$
113 earthquake was recorded by the local seismic network (Theunissen et al., 2010). In 1998, Kao,
114 based on a study of historical earthquakes, said that it was inconclusive whether a $M > 8$ event
115 may occur or not in and around Taiwan. In terms of distribution, both Benioff zones are clearly
116 visible east of $121^\circ 30'E$ for the PSP slab and south of $23^\circ N$ for the SCS slab. Many authors
117 consider that the SCS slab, despite the absence of Benioff zones north of $23^\circ N$, still extends until
118 at least $24^\circ N$ (Fig.2) or even more northward, based on global or local tomography (Wu et al.,
119 1997; Wang et al., 2006; Lallemand et al., 2001 ; Ustaszewski et al., 2012). Seismicity
120 concentrates in some very active areas such as the Foothills in the western part of the orogen
121 (e.g., 1999 $M_w 7.6$ Chichi Earthquake), the Coastal Range which is an extinct segment of the LVA
122 colliding with the orogen (e.g., 1951 $M_w 7$ earthquakes triplet), the coastal region north of the
123 Coastal Range is also extremely seismic as well as the E-W-trending South Okinawa rift valley
124 near $24^\circ 45'N$. One puzzling feature is the clustering of earthquakes in the SRA forearc. Instead of
125 showing a classical pattern of seismogenic zone extending from ≈ 10 -20 km to ≈ 50 -60 km depths
126 (e.g., Peacock and Hyndman, 1999, Heuret et al., 2011), we observe a concentration of
127 earthquakes at shallow depths below the forearc basins. Font and Lallemand (2009) have
128 studied in detail this region by relocating events taking advantage of a new code and 3D velocity
129 model (MAXI; Font et al., 2004) and the joint use of Taiwanese and Japanese land stations. They
130 succeeded to better locate the epicenters of the events but were unsuccessful in determining
131 accurate depths using only land stations far from the sources. The largest magnitude earthquake
132 never recorded in Taiwan ($M_w 7.7$) occurred within this cluster (Theunissen et al., 2010) that
133 comprises mostly shallow northward dipping thrust faults (Kao et al., 1998). The stress pattern,
134 inversed from focal mechanisms of crustal earthquakes in the region northeast of Taiwan, is
135 complex (Wu et al., 2010a; Wu et al., 2010b; Huang et al., 2012). E-W intra-PSP compression is
136 observed in many places offshore the northern Coastal Range (e.g., 1967 $M_w 6.8$ or 1986 $M_w 7.3$
137 events) as well as strike-slip (see 1966 events described above and Fig.1) or normal faulting
138 compatible with E-W extension (e.g., 2001 $M_w 6.8$ event). The diversity of focal mechanisms
139 results from the complex 3D interaction of three plates (EP, PSP, SRA) including collision,
140 subduction and rifting (Fig.1). Deciphering between structural units involved first requires to
141 map the plates interfaces in depth.

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2.3 - Geology of PSP, SRA and Taiwan East coast

The study area includes three domains s.s. : the Huatung basin, the eastern coast of Taiwan and the southern Ryukyu forearc (Fig.1 & 3).

The Huatung basin is an oceanic domain, about 120 km wide, lying between the LVA to the west and the Gagua Ridge to the east. It subducts to the north beneath the SRA along the Ryukyu trench. The origin and age of this basin is controversial but samples dredged on basement highs suggest an Early Cretaceous age for the crust (Deschamps et al., 2000). Its mean depth is 400 m shallower than expected for an oceanic crust of that age (Deschamps et al., 1998). The crust is loaded by sediments, up to 2 km thick on edges, mostly supplied from the nearby growing orogen. Furthermore, refraction profiles indicated that the crustal thickness was about 12 km (McIntosh and Nakamura, 1998) revealing some possible excess of magmatism during its creation. Reflection seismics and gravity modeling both shows a flexure of the basin under the load of the LVA on one side and the Gagua Ridge on the other (Deschamps et al., 1998). The origin of this old piece of oceanic crust, separated from the Eocene West Philippine Basin by the N-S trending Gagua Ridge, remains unknown. The age contrast on both sides of the Gagua Ridge and its linearity argue in favor of an old transform zone uplifted during a plate reorganisation in the mid-Eocene (Deschamps et al., 1998). During the compressional event, part of the adjacent basin was probably subducted. The ridge now culminates at about 4 km above the surrounding seafloor.

The structure and nature of the eastern coast of Taiwan varies from south to north. South of 22°40'N, the growing crustal accretionary wedge is exposed along the coast. The extinct LVA parallels the coast on the east at a distance narrowing from 60 km in the south (Lanyu island) to 30 km at the latitude of Taitung (Lutao island). Between 22°40'N and 24°N, the LVA appears fully submerged except in some small places where it outcrops on the Coastal Range (e.g., Chimei edifice). The Coastal Range itself mainly consists of intra- or fore-arc volcano-sedimentary basins that have been shortened during the arc-continent collision (Huang et al., 1992). It overthrusts the Longitudinal Valley to the west that lies at the foot of the Central Range, i.e., the innermost unit of the orogen. North of Hualien (24°N), the metamorphic Central Range directly outcrops along the shore, where unstable flanks collapse into the 3000 m deep Hoping Basin. North of that, near 24°45'N, the triangular flat Ilan Plain corresponds to the westernmost termination of the SOT propagating rift, attesting for an abrupt change of the tectonic regime from compression south of 24°35'N to extension north of that (Lu et al., 1995). Kueishan Tao (Turtle island), located 10 km offshore Ilan (Fig.1), is an island arc volcano lying right in the E-W trending axis of the rift, with a chemistry which differs from subduction basalts (R. Shinjo, personal comm.).

The SRA elongates along an E-W direction, south of the SOT and east of the NE Taiwan coast. It consists of a presently non-volcanic arc on which a few islands rise above sea-level such as the small Yonaguni island and the largest Iri-Omote island (see Fig.1) where 13 Ma old OIB-type high-Mg andesites outcrop (R. Shinjo, personal comm.). Yonaguni island is mostly composed of tabular upper Pleistocene Ryukyu limestone and Holocene raised reef limestone lying on lower to middle Miocene sandstones and siltstones of the Yaeyama Group. Most of Iri-Omote island is covered by the Yaeyama Group but there, the Triassic and Jurassic metamorphic basement outcrops in the eastern part of the island as it also does in the other islands of the Ryukyus (Ishigaki, Miyako) (Fabbri and Fournier, 1999). South of the arc, a series of forearc basins align back of a well-developed sedimentary accretionary wedge (Lallemand et al., 1999). The basement below the forearc basins is characterized by lows (e.g., 10 km deep "hole" 50 km off Hualien) and highs (promontories) with an amplitude of at least 2 km (Font et al., 2001). One of these basement promontories : the Nanao rise, has been attributed to an uplift caused by the subducting Gagua Ridge whereas another one : the Hoping rise, closer to Taiwan, is not explained yet. The Ryukyu arc basement ends abruptly at the transition between the forearc basins and the accretionary wedge as first proposed by Font et al. (2001) and later supported by wide-angle refraction data (Klingelhoefer et al., 2012). According to geodynamic reconstructions,

195 the forearc basins and the accretionary wedge are probably no older than a few million years in
196 the study area (Lallemand et al., 2001).

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2.4 - Tectonic models

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Several models tempted to account for the main features in this region either based on surface

200 geology and geomorphology (Suppe, 1981, 1984; Teng, 1990; Lu et Hsu, 1992; Chang et al., 2000,

201 2003; Malavieille et al., 2002), on local seismicity and velocity data (Wu et al., 1997; Lin, 2000;

202 Wang et al., 2001; Carena et al., 2003; Kim et al., 2006; Ustaszewski et al., 2012), on reflection

203 and refraction seismic data (Lallemand et al., 1997, 1999; Font et al., 2001; McIntosh et al., 2005),

204 on analog modeling (Lu and Malavieille, 1994; Lu et al., 1995; Chemenda et al., 1997; Dominguez

205 et al., 1998; Malavieille et Trullenque, 2009;) or on geophysical data and global geodynamics

206 (Hsu and Sibuet, 1995; Teng, 1996; Teng et al., 2000; Lallemand et al., 2001; Ustaszewski et al.,

207 2012; Sibuet et Hsu, 2004). The models differ in the northern extent of the SCS slab, the

208 northeasternmost extent of the orogen deformation front, the existence of a pre-existing

209 volcanic arc fringing the Eurasia margin before the collision, the existence or the lack of

210 subduction polarity reversal, or the geometry of the plate boundaries offshore NE Taiwan. A few

211 of them accounts for the deformation below the junction between the SRA and the Taiwan

212 orogen (Lallemand et al., 1997, 2001 ; Malavieille et al., 2002 ; Ustaszewski et al., 2012). The

213 figure 2 shows a plausible geometry of the EP when removing the PSP modified after Lallemand

214 et al. (2001). The main discussion focusses on the plates existence and/or configuration beneath

215 northern Taiwan. In a recent paper, Ustaszewski et al. (2012) consider that the lower crust and

216 the mantle part of the EP subducts eastward up to 25°N (latitude of Taipei) and that all the SRA

217 has been delaminated. The plate configuration at depth together with the eastward component

218 of plates convergence explains why the deformation is paroxysmal offshore NE Taiwan. The

219 geometry of the subducting PSP is also controversial and will be discussed in this study.

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3- New results from RATS experiments

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3.1 - The RATS experiment

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225 A passive experiment was first conducted in 2008 using 22 ocean bottom seismometers (15

226 short-period and 7 broadband OBS) and 51 inland seismic stations which have recorded the

227 microseismicity during three months from July 19 to October 24. Then, an active experiment

228 (wide-angle and coincident reflection seismics) occurred in May 2009 using 24 short-period OBS

229 aligned across the SRA system, 6600 inch³ seismic sources and a 6 km long streamer of the US

230 R/V Marcus Langseth. Given the distribution of Taiwanese and Japanese land seismic stations

231 with respect to the active seismic zones, the passive OBS array was deployed in the forearc area

232 of the SRA on both sides of the active line (see Fig.3).

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3.2 - The active 2D line through the Ryukyu arc

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236 The active line has been set to cross the major seismic cluster observed in the forearc area of the

237 SRA, i.e., the Hopping cluster (located beneath the Hopping Rise - see Figs.1 & 4). The main

238 outcomes of the velocity structure along this line (Klingelhoefer et al., 2012) are (1) the presence

239 of a rather thin (5-6 km) subducting oceanic crust overlying a « low-velocity » mantle (7.8 km/s)

240 supposed to be hydrated by hydrothermalism, (2) the low velocities associated with the

241 accretionary wedge can be traced back to the Nanao forearc basin, resulting in an abrupt

242 termination of the SRA basement, (3) a low velocity zone has been detected between the base of

243 the SRA basement and the subducting PSP, and (4) the SRA Moho where the active line crosses

244 has been detected at a depth of about 25 km.

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3.3 - The passive experiment: a new 3D velocity model

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248 This velocity section together with other velocity data in the region were used to build a 3D a
249 priori P-wave velocity model, used itself as initial velocity model for the joint P and S-waves
250 velocities inversion with about 1000 micro-earthquakes hypocenters recorded during the
251 passive experiment (Theunissen et al., 2012a). Our 3D tomographic model results from an
252 inversion within a $10 \times 10 \times 6$ km³ grid further interpolated into a $1 \times 1 \times 1$ km³ grid in order to
253 calculate travel-times and to draw tomographic images. This last interpolation induces a
254 maximum uncertainty equal to the initial grid size, i.e., 6 km in vertical and 10 km in horizontal
255 at the most. Checkerboard tests revealed that the dataset, both with P- or S-wave, is able to
256 reconstruct in shape $20 \times 20 \times 12$ km³ and $30 \times 30 \times 12$ km³ pattern velocity anomalies in the target
257 area, below the OBS network and to the west at the transition with Taiwan, down to a depth of
258 about 40–50 km. Finally, using the new 3D tomography together with the OBS data greatly
259 improves the hypocenter determination, especially in depth.

261 **3.4 - Relocation of earthquakes (1992-2008 M_L≥3)**

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263 The new 3D velocity model has been tested without the use of the OBS for locating 77 reference
264 events well-located during the 3 months of deployment thanks to the new version of the MAXI-
265 code (Font et al., 2004 ; Theunissen et al., 2012b). The tests show that it greatly improves the
266 location of the hypocenters in the studied region (Theunissen et al., 2012a). In this study, we
267 have relocated all M_L≥3 events that were recorded both by the Taiwanese seismological network
268 (CWB for Central Weather Bureau) and the Japanese one (JMA for Japan Meteorological Agency)
269 during the period 1992-2008. We have applied the following procedure :

270 (1) In the area $121^{\circ}\text{E} - 124.4^{\circ}\text{E}$, $22^{\circ}\text{N} - 25.3^{\circ}\text{N}$, a total of 11,475 M_L≥3 events were
271 selected from picking catalogs of both JMA and CWB satisfying the criteria to have been detected
272 by at least one JMA and two CWB stations. For this initial catalog, the number of P-
273 measurements is, in average, 21 ± 6 of which 16 ± 6 from the CWB and 5 ± 1 from the JMA.

274 (2) In order to avoid uncertainties in the S-arrival-times and associated velocity model
275 inherent to the geometry of our seismic network (the distance between an event and the nearest
276 station may be large), we have only used P-arrival-times (Font et al., 2004; Font et al., 2013). P-
277 travel-times tables are calculated within the new 3D velocity model using the Shortest Path
278 Method (Moser, 1991). We have used a $10 \times 10 \times 3$ km³ cell size with a node distribution on each
279 facet every 1 km in depth and 2 km in horizontal.

280 (3) We then applied the MAXI algorithm (Theunissen et al., 2012b) to realize the
281 absolute earthquake location allowing the use of only P phases. Solutions are sorted according to
282 confidence factors based on MAXI algorithm (our estimated uncertainty is ± 8.0 km in depth and
283 ± 4 km in horizontal in average ($+1\sigma$)) rather than rms error (χ^2 uncertainty is ± 4.0 km in depth
284 and ± 1.8 km in horizontal in average ($+1\sigma$)) (see appendix for further details).

285 (4) Finally, 5996 earthquakes, recorded by 22 ± 6 phases in average, have been selected
286 and are used to build images (Fig.4) and make interpretations. Our relocated depths are, on
287 average, 4.4 km deeper than those of CWB but with large variations from place to place (see
288 appendix for further details). Two regions show opposite tendencies, i.e., shallower locations :
289 the Nanao cluster and the rift zone (see Fig.S1 in Appendix).

290
291 These results are in agreement with the shallow seismic activity in the upper crust at these
292 locations (Konstantinou et al., 2011; Lin et al., 2008) and the deeper one in the other areas and
293 especially in the southern Ryukyu forearc region. The comparison of our results and
294 hypocenters determined by the CWB for 27 events associated with well-determined focal
295 mechanisms (see Table 1 for sources), along three reference sections across the Southern
296 Ryukyu forearc system (Fig.5, see location on Fig.4), shows a seismicity pattern that is more
297 clustered and in better agreement with the expected geodynamics of this region. For example,
298 the Hoping seismic cluster (HC in Fig.4), that was initially located in the upper plate in the
299 previous studies (Kao et al., 1998; Font and Lallemand, 2009), now mainly concentrates in the
300 vicinity of the plates interface (Fig.5). Furthermore, we verify that focal mechanisms better

301 agree with the overall subduction context with most subduction-thrust events locating in the
302 upper part of the subducting crust or along the plate interface and the other events showing slab
303 lateral compression at depths larger than 40 km in the uppermost lithospheric mantle as
304 previously mentioned by Kao et al. (1998). Details on focal mechanisms are given in Table 1.
305 When different focal solutions were available for a single event, our selection priority order
306 was : (1) Kao et al. (1998), (2) CMT Harvard, (3) BATS (Kao and Jian, 1999) for most
307 earthquakes except three old events in 1951 (Cheng et al., 1996) and in 1963 (Chen et al., 2004).

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309 We will now discuss these new results and integrate them into a comprehensive geodynamic
310 model.

311 312 **4- Revised plates geometries based on the new 3D local tomography**

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314 In this study, we obtain a local tomography in the triangular area between Chengkung (50 km
315 north of Taitung), Ilan and Iri-Omote (see Fig.1) that undergoes the maximum strain caused by
316 arc-continent collision. The density of onland and offshore seismic stations during the RATS
317 experiments has allowed to achieve a much better resolution than previous studies for shallow
318 levels, decreasing in size in depth (see Fig.6).

319
320 The main challenge in this complex region is the discrimination between EP and PSP in the area
321 of interaction, i.e., the collision and subduction interfaces, which are partly located offshore
322 where deep seismic imagery was not reliable before the TAIGER and RATS experiments. We will
323 first point out the main differences of our new dataset with those already published (Rau and
324 Wu, 1995 ; Lin et al., 2004 ; Kim et al., 2005, 2006 ; Wu et al., 2007a ; Wang et al., 2006 ; Wu et al.,
325 2009a&b ; Kuo-Chen et al., 2012; Ustaszewski et al., 2012) and then examine the various
326 structural domains. All tomographic models have used local earthquakes recorded by CWB and
327 sometimes combined with JMA stations. A few of them have used teleseismic events in addition
328 to local events. Those who have used OBS data get better resolution in the areas where OBS are
329 deployed compared with others like Lin et al. (2004) for example in the SOT or Wu et al. (2009b)
330 in the area offshore SW Taiwan, but only Kuo-Chen et al. (2012) have used a large number of
331 OBS deployed all around the island during more than a year. Those OBS were scarcely spaced on
332 the seafloor whereas those of the RATS experiment were quite dense in the southern Ryukyu
333 forearc area.

334
335 Tomographic models generally show a thicker (up to 50 or even 60 km) crustal root beneath the
336 Central range near 24°N that extends from 23°20'N to 24°50'N (Rau and Wu, 1995 ; Wu et al.,
337 2007a; Kuo-Chen et al., 2012). The map of the « Moho » on Figure 7 has been drawn from our
338 new 3D tomographic model. The sudden increase in the velocity gradient, near 7.5 km/s beneath
339 the Huatung Basin and 7.75 km/s beneath the SRA, helped to constrain Moho's depth. We have
340 used 7.5 km/s for the PSP and 7.75 km/s for the EP and SRA as isovelocities to pick the
341 respective Mohos. One difficulty when drawing the Moho depth map is that the Mohos of the EP
342 and the PSP overlap in areas like beneath the Coastal Range or the SRA. We were thus very
343 cautious with the data using seriated vertical sections (like those shown in Figs.8&9) in order to
344 pick the right plate for a given « Moho » surface. We have accurately picked the seaward limit of
345 the continental « Moho » where it abuts against the top of the subducting PSP beneath the SRA
346 and the subvertical boundary between EP and PSP beneath E-Taiwan (see the black dashed line
347 on Fig.7). It was often difficult to trace the « Moho » of the subducting PSP beneath those of the
348 EP in the region of the SRA based on our 3D model because we lost resolution at such depths
349 and because eclogitisation occurred in the slab. Focussing on the area offshore where we have
350 improved the resolution compared with earlier models (see the thin line delimiting the region
351 where significant rays cross, i.e., >50 rays/cell), we are able to confirm some previous
352 observations and point new ones :

- 353 (1) The thickness of the Huatung Basin crust averages 9-10 km before its subduction but it
354 locally thickens up to 16 km beneath the Ryukyu forearc area as already noted by
355 McIntosh and Nakamura (1998) based on wide-angle seismics.
- 356 (2) The crustal root of the LVA (purple crosses on Fig.7) can be traced at a mean depth of
357 32 ± 3 km up to 24°N and it suddenly deepens to 42 ± 3 km north of Hualien up to $24^{\circ}20'\text{N}$.
- 358 (3) The PSP Moho deepens both northward and westward with a highest slope angle to the
359 west down to a maximum depth of 65 km between 24 and $24^{\circ}30'\text{N}$.
- 360 (4) The LVA is separated from the Central Range by a narrow elongated area of high
361 velocities (see small red dots on Fig.7), typically 6-7 km/s rising close to the surface (15-
362 20 km depth in general but locally up to 10 km immediately north of Hualien). Similar
363 feature has been observed by Kuo-Chen et al. (2012). Such narrow high V_p regions (HVZ)
364 have also been mapped in the SRA as far east as $122^{\circ}40'\text{E}$ beneath the southern slope of
365 the SRA (Fig.7).
- 366 (5) The PSP Moho appears slightly undulating west of $122^{\circ}50'\text{E}$ beneath the Ryukyu forearc
367 (right below our seismic network). At a large scale, the strike of the isodepths changes
368 from E-W (trench-parallel) east of $122^{\circ}50'\text{E}$ to NW-SE ($\approx\text{N}305^{\circ}$) west of it and then NNE-
369 SSW ($\approx\text{N}15^{\circ}$) beneath the east coast of Taiwan. The PSP Moho is almost flat and shallow
370 beneath the westernmost accretionary wedge near Taiwan and it is vertically offset
371 downward by several kilometers near $24^{\circ}15'\text{N}$ along a WNW-ESE-trending « step ». Less
372 pronounced vertical offsets are detected perpendicular to the previous one beneath the
373 Hoping Rise and the Nanao Basin. Such PSP Moho rise trending parallel to the E-Taiwan
374 coast was already observed by previous tomographic models (Wu et al., 2007a ;
375 Ustaszewski et al., 2012) but with a lower resolution in the absence of OBS data. The
376 highest amplitude of the rise is observed at the latitude of Hualien right below the
377 Hoping Rise. The Hoping Rise in the SRA basement toe is thus associated with a rise in
378 the PSP Moho.
- 379 (6) A wide zone located offshore East Taiwan between $23^{\circ}20'\text{N}$ and $24^{\circ}30'\text{N}$ is characterized
380 by anomalously low velocities (LVZ) between 7.5 and 8 km/s in the PSP mantle below
381 the Moho (see dashed brown line on Fig.7).
- 382 (7) The EP « Moho » rapidly deepens beneath the Central Range down to a depth of ~ 60 -70
383 km where it can not be traced deeper. Our results confirm those of Kuo-Chen et al.
384 (2012) and Ustaszewski et al. (2012) up to $24^{\circ}30'\text{N}$, but differ northward of this latitude.
385 We do not observe any sign of eastward subduction of the EP north of $24^{\circ}30'\text{N}$ whereas
386 Ustaszewski et al. extend the subduction zone at least up to 25°N . They also propose that
387 the continental Moho above the subducting PSP has been newly formed by delamination
388 and westward rollback of the lower part of the EP (and SRA) « slab ». Our data show that
389 the average depth of the SRA Moho is 30-35 km but it deepens from 35 km, at about 40
390 km east of the northeast coast of Taiwan, to about 55 km beneath the Ilan Plain.
- 391 (8) Our seismic network allows us to add new constraints at the junction between NE
392 Taiwan and the SRA. The crustal root outlined by the EP « Moho » appears quite thick
393 right beneath the Ilan plain coastal area as observed by Lin et al. (2004). Both the EP and
394 the PSP Mohos isocontours are torned by $\sim 90^{\circ}$ near $24^{\circ}30'\text{N}$ indicating an abrupt
395 change in the $\text{N}10^{\circ}$ global trend beneath east Taiwan.
- 396 (9) The thickened crust of the EP and SRA around $24^{\circ}35'\text{N} - 122^{\circ}05'\text{E}$ appears to be sharply
397 cut at the contact with the subducting PSP (SRA Moho isocontours normal to the
398 continental Moho edge), giving the impression that part of the orogen has been removed.

400 **5- Structure and deformation of the PSP**

401

402 The PSP mostly deforms offshore NE Taiwan where our OBS network has been deployed as
403 indicated by the high density of shallow earthquakes (Fig.4). We will further characterize the
404 structure and the deformation of the PSP on the basis of local tomographic sections (Figs.8&9),
405 relocated hypocenters and available focal solutions for main events (Figs.4&5).

406

5.1- The Huatung Basin

The Huatung Basin is an Early Cretaceous oceanic domain which carries the LVA on its western edge and is bounded by the Gagua Ridge on its eastern edge. The undulations observed on the PSP Moho (see Fig.7) are rather caused by a bending of the plate as a consequence of its incipient subduction beneath East Taiwan than a buckling at a larger scale. Indeed, (1) the maximum amplitude is observed about 30 km off the east coast of Taiwan, (2) the wavelength is of the order of 100 km whereas it should be closer than 500 km for a 110-140 Ma old lithosphere and (3) the Huatung crust is limited by the Gagua Ridge to the east. As said above, we observe a thicker oceanic crust than those estimated from a 2D recent wide-angle experiment : 5-6 km after Klingelhoefer et al. (2012) but ~12 km after McIntosh et al. (2005) compared with more than 9 km based on a 3D tomographic model. The discrepancy can be partly explained by the lower resolution of our tomography compared with refraction. Local thickening of the oceanic crust can also be caused by intra-oceanic slivering of the PSP as suggested by focal mechanisms of some recent $M \geq 6$ events (see events 24/4/72, 23/12/78, 14/11/86, 29/7/96, 14/4/00, 9/6/03, 1/6/05, 6/9/07 in Figs.4, 8&9). Such thrust mechanisms (see on the left side of Fig.4) were already described and attributed to « lateral compression » in the direction of oblique convergence by Kao et al. (1998). ESE-WNW shortening of the PSP is required as a consequence of its collision against the steeply east-dipping EP slab up to $24^{\circ}30'N$ (see sections 1 and 2 on Fig.8). The anomalous low velocities (LVZ areas on Figs.8&9) observed beneath the PSP crust can be attributed either to a hydrated mantle as proposed by Klingelhoefer et al. (2012) or to the presence of slices of oceanic crust (and uppermost mantle ?) at depth (see section 3 on Fig.8). Hydration of the oceanic mantle at those depths (30-70 km) can result from hydrothermal alteration from the seafloor through fault conduits or from below since this area is contiguous with the LVA above the EP descending slab. Along section 1 (Fig.8), most of the seismicity occurs at depths between 30 and 55 km at the tip of the westward incipient subduction of the PSP crust but also within the brittle part of the oceanic mantle which undergoes strong ESE-WNW compression. Sections 2 and 3 (Fig.8) image the subducting PSP buried beneath the SRA basement and accretionary wedge. Again, earthquakes cluster at the tip of the westward subducting PSP crust as well as within the PSP mantle.

5.2- The LVA and the suture zone in E-Taiwan

The LVA is identified on sections 1 and 2 (Fig.8) because of its associated low velocity root and because it is known to outcrop along and offshore the Coastal Range. The arc can be traced below the east coast of Taiwan north of Hualien up to $24^{\circ}20'N$ (Fig.7, section 2 on Fig.8). The tricky point concerns the presence of a narrow high velocity zone (HVZ) that fringes the LVA on its west side and locates below the « suture » between EP and PSP (heavy black dashed line on section 2, Fig.8). The surface projection of this narrow strip of high velocities is shown on figure 7 (elongated rise seen on the 7.5 km/s isocontour marked with red dots). This narrow strip appears indented by the LVA along section 2 with velocities as high as 6.5 km/s at only 8 km depth below the coastline just north of Hualien. Given to the structural position of the HVZ adjacent to the LVA and the velocities between 6.5 and 7.5 km/s, we suggest that this HVZ may correspond to pieces of oceanic crust (and uppermost mantle ?) that initially belonged to the LVA forearc basement and which are squeezed between the deformed LVA and the Taiwan orogeny in agreement with Kuo-Chen et al. (2012). Patches of HVZ are also observed below the southern slope of the SRA (Fig.7, section 3 on Fig.8 and sections 4, 5, 7 on Fig.9). These occurrences might indicate that the arc-continent collision migrated through there, east of its present location and that remnants of Luzon forearc (and arc ?) pieces were accreted at the front of the SRA basement. Focal mechanisms in the vicinity of the LVA are generally thrusts compatible with E-W shortening, sometimes with a significant left-lateral strike-slip component along a steep $N15^{\circ}$ -trending reverse fault (M7 1951 earthquake on section 2 for example).

5.3- The subducted part of the PSP

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462 Another type of deformation is observed in the subducted part of the PSP when looking in the
463 direction normal to the Ryukyu trench (Fig.9). One basic question is the nature of the two
464 seismic clusters that lie below the southern slope of the SRA, i.e., the Suao cluster near Taiwan
465 and the Hoping cluster further east (Fig.4, Font and Lallemand, 2009 ; Theunissen et al., 2012a).
466 Indeed, many historical earthquakes, including the largest one in 1920 (M7.7, Theunissen et al.,
467 2010), occur in this region. The major question concerned the depth constraints on these events
468 but this question has been likely solved using the new data acquired during the RATS
469 experiments that allow us to precisely map the subduction interface (Fig.10) and relocate past
470 seismic events in the new 3D velocity model.

471
472 The map of the subduction interface (Fig.10) reveals several features. Firstly, small scale
473 undulations especially close to Taiwan likely correspond to slivering of the oceanic crust (see
474 Fig.8) compatible with compression in the EP-PSP convergence direction. Such ESE-WNW
475 « lateral » compression, already described by Kao et al. (1998) is attested by many reverse focal
476 mechanisms (see section 5.1). Secondly, the PSP slab does not subduct smoothly beneath the
477 SRA but rather shows ramps as already observed by Theunissen et al. (2012a). The ramps along
478 the plates interface result from the northeastward downfaulting of the slab along en-échelon
479 steeply-dipping shear zones trending NW-SE (Figs.9&10). A close-view of the V_p model where
480 the velocity contours clearly mark a step is shown on section 5 (Fig.9). Spike tests were
481 successfully performed in that region in order to guarantee the reliability of the ~10 km vertical
482 offset. The Hoping seismic cluster is clearly associated with one of the ramps that accommodates
483 the slab offset (see sections 5, 6 and 7 on Fig.9). Major events occur either within the subducting
484 plate or along the subduction interface. Focal solutions of the largest events are very
485 homogeneous in this cluster (Font and Lallemand, 2009 and see focal mechanisms on top of
486 Fig.4 and on Fig.5) and the NW-SE trend of the shallow-dipping nodal planes argues for a thrust
487 mechanism along the ramps which results from the downfaulted slab. According to the global
488 geodynamic context, the motion along the subvertical shear zone cutting through the slab should
489 be normal with a sinistral component. It has been suggested by some authors (Lallemand et al.,
490 1997 ; Font et al., 1999; Theunissen et al., 2012a) that a tear develops within the PSP slab to
491 account for the collision along the Coastal Range and the subduction north of it. The en-échelon
492 ramps observed along the subduction interface and the change in slab dip across them argue for
493 a growing tear, non-mature yet, with en échelon segments slightly oblique to the convergence
494 direction. Normal fault mechanisms indicating NNE-SSW extension with sinistral component
495 near the subduction interface or within the subducting PSP were calculated from events
496 recorded during the 3-months RATS passive experiment (see events 7 and 13 in Theunissen et
497 al., 2012a). On March 12, 1966, a M7.5 WNW-ESE pure sinistral strike-slip earthquake occurred
498 between 28 km (Engdhal et al., 1998) and 22 km depth (Kao et al., 1998), i.e., close to the
499 subduction interface. The three M6 deeper (47 to 55 km) aftershocks that occurred in March and
500 May 1966 with similar mechanisms were likely associated with this intra-slab sub-vertical shear
501 zone. Accurate relocation of these events should be done to decipher whether they really occur
502 within the subducting slab or not.

503 Another characteristics may reveal the age contrast between the oceanic crust on each side of
504 the subducting PSP. The subducting crust of the Huatung Basin (HB) can be traced at least down
505 to ~50 km on sections 4 to 7 (Fig.9) whereas velocities up to 8 km/s or more are observed in the
506 crust deeper than 55 km beneath Yaeyama islands (Yonaguni, Iri-Omote, see Fig.10). We
507 interpret these high velocities as the result of the eclogitisation process occurring shallower
508 where the crust is younger. We also noted the presence of a small subducting ridge trending
509 NW-SE right in the axis of the paleo-West Philippine Basin (WPB) spreading center
510 reconstructed by Deschamps and Lallemand (2002). If the subducting elongated high
511 corresponds to the paleo-WPB extinct ridge, then the age of the subducting WPB below the
512 Yaeyama islands is about 35 Ma to be compared with 140 Ma on the west side of the subducting
513 Gagua Ridge (Fig.10). We also pointed a systematic deepening of the subduction interface at
514 depths larger than 30 km west of Yonaguni Island along the northern extent of Gagua Ridge. We

515 propose that this deepening together with the trend change of the isocontours of the subduction
516 interface can be caused by some thrusting of the young crust onto the old one.

517

518 **6- Structure and deformation of the southern Ryukyu forearc**

519

520 To characterize the structure and the deformation of the SRA, we use two new MCS lines 12 and
521 22 acquired during the TAIGER project in addition to local tomographic sections, relocated
522 earthquake hypocenters and available focal solutions for the main events.

523

524 **6.1- The SRA basement**

525

526 The mean thickness of the SRA crust is 30-35 km (Fig.7) which corresponds to the average depth
527 of the EP continental « Moho » far from the arc-continent collision: 30-35 km as depicted by
528 Nissen et al. (1995) in the northern margin of the South China Sea and 25-30 km in the Ryukyu
529 Arc near Okinawa for example (Iwasaki et al., 1990 ; Kodaira et al., 1996). Furthermore, the
530 thickest portion of the arc is clearly observed along its southern edge (south of Yonaguni Island
531 or east of Suao, Fig.7) suggesting a truncation of a thick crust from the south extending at least
532 as east as 123°30'E. These observations lead to two conclusions: (1) the minimum thickness of
533 the arc (25-30 km) is representative of the rest of the Chinese platform, so that it precludes the
534 arc from basal delamination processes as suggested by Ustaszewski et al. (2012) and (2) the
535 thickest crust truncated along its southern edge might indicate that the Taiwan orogen extended
536 there initially before its collapse and tectonic erosion associated with the ongoing subduction
537 and/or Luzon forearc or arc material has been underplated . Based on the 7.75 km/s V_p
538 isovelocity, we observe the same thickness for the EP crust below the Ilan Plain (55 km) than Lin
539 et al. (2004) but we are doubtful about its representativity as the EP Moho because Wang et al.
540 (2010) based on receiver functions obtained a very thin crust right below the Ilan Plain (18 km
541 at station SLBB using only Moho P_s arrivals). High V_p/V_s ratio suggests partial melting at those
542 depths (Lin et al., 2004) but at the same time, earthquakes occur in the same deep zone far
543 above the PSP Benioff zone. We thus can not conclude about the real EP Moho depth in this area.
544 Regarding the toe of the SRA basement, Font et al. (2001), based on a set of reflection seismic
545 lines, have shown that promontaries (or uplifted regions) existed below the Hoping Rise and
546 Nanao Rise (see Fig.1) and that the southern edge ends abruptly at the rear of the accretionary
547 wedge. The new 3D tomography indicates that high velocity zones (HVZ near 6.5 km/s) are
548 detected at shallow depths around 20 km within the SRA basement below the southern slope of
549 the Ryukyu Arc (see Figs.7, 8 and 9). We correlate these shallow HVZ with those observed west
550 of the LVA below the « suture » between the EP and the PSP. We propose that these HVZ might
551 represent pieces of LVA forearc basement offscraped from the colliding PSP when the LVA crept
552 along the southern Ryukyu margin around 1 or 2 m.y. ago.

553

554 The seismic image of the SRA basement is not clear as shown on the multichannel seismic line
555 12 acquired during the TAIGER-RATS project (Fig.11). Nevertheless, high-amplitude reflections
556 marking the top of the basement can be followed southward beneath the sediment of the Nanao
557 forearc basin. They stop right back of the crest made by the rear of the accretionary wedge (see
558 arrow on Fig.11, km62). The refraction line acquired simultaneously also shows an abrupt
559 termination of the SRA basement at the same location based on velocity contrast between the
560 trench sediment accreted in the frontal wedge and the SRA basement (Klingelhoefer et al., 2012).
561 Velocities inverted on section 6 (Fig.9) also point this abrupt change. Based on local tomography,
562 the subduction interface on that section exhibits two ramps, one below the accretionary wedge
563 where the reflections marking the top of the subducting oceanic crust disappear (see arrow on
564 Fig.11, ~km70) and another below the slope break between the forearc basin and the SRA slope
565 (~km115 on Fig.11). This last ramp is associated with the Hoping seismic cluster that mainly
566 concentrates near the ramp but also extends above within the SRA basement. The distribution of
567 relocated earthquakes in the vicinity of line 12 (same as those on section 6 of Figs.5&9) shows
568 two important things. First, the Hoping seismic cluster occurs in very narrow region that

569 coincides with the main ramp of the subduction interface. Second, most of the events occur
570 within the PSP crust or near the interface (see Fig.11) but still a significant number of them
571 apparently occur within the upper plate. The small white dots are microevents recorded during
572 the 3-months of passive recording using the OBS network, so that we are confident in their
573 location. All are located at depths larger than 16 km, i.e., close to the subduction interface,
574 whereas the older events relocated using the new 3D velocity model but without OBS stations
575 have a much larger extent in depth from the seafloor to about 32 km. This vertical spreading
576 may be caused either by the poor resolution in depth in the absence of near-field seismic
577 network and some seismic/tectonic accommodation of the SRA basement above the ramp. We
578 attribute the lost of deep reflections arcward of km62 (arrow on Fig.11) to strong internal
579 deformation in this sector.

580 **6.2- Forearc vertical movements and extension**

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582
583 A trench-sub-parallel line MCS-22 (see location on Fig.4 and section 3 of Fig.8) cuts through the
584 forearc basins (Hoping, Nanao and East Nanao) and well illustrates the structural context of the
585 SRA basement at the Hoping and Nanao rises (Fig.12). We see large variations in SRA basement
586 depths along this line. If we take a reference line at the depth of the Nanao Basin deepest point,
587 we can measure a vertical offset that exceeds 1 km for the Nanao Rise and 3 km for the Hoping
588 Rise. The line MCS-22 is not ideally oriented parallel to the trench. It obliquely cuts the
589 structures but still, we observe a large uplift of the SRA basement above the subducting northern
590 extent of the Gagua Ridge accompanied by normal faulting as attested by the intense seismicity
591 (see Nanao cluster on section 3 of Fig.8), numerous west-dipping normal faults visible on
592 seismic lines (see Fig.12) and detailed bathymetric map (Lallemand et al., 1999). A recent
593 expression of the E-W extension affecting the SRA basement is the occurrence of the December
594 18, 2001 M6.7 shallow earthquake that occurred within the Nanao Rise (Figs.4 and 8). The uplift
595 of the Hoping Rise appears much larger allowing the SRA basement to outcrop on the seafloor
596 with a scarp facing east (see Fig.12) and inducing a strong erosion of the folded Suao Basin
597 strata. Our interpretation of section 3 (Fig.8) is that slices of the shortened Huatung Basin
598 produce oceanic highs that deform the upper plate when subducting.

599 **6.3- Transcurrent faulting : dextral or sinistral ?**

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601
602 The section 4 (Fig.9) cuts across the Suao cluster near Taiwan. There, the seismicity still occurs
603 within the subducting slab but also extends throughout the upper plate from 35 km in depth up
604 to the seafloor. The cloud of events outlines a south-dipping high-angle (probably) listric ESE-
605 WNW-striking fault. This is in agreement with focal mechanisms which are mostly strike-slip
606 with nodal planes trending N-S (dextral motion) and E-W (sinistral motion). The aftershocks of
607 major events like the « Nanao earthquake » (June 5, 1994, $M_s6.1$) delineate a rupture along the
608 E-W nodal plane (Wu et al., 1997 ; Kao et al., 1998). As said above, the relocated cluster is clearly
609 elongated along an E-W direction from the surface down to 35 km. We thus favor an E-W
610 trending sinistral strike-slip mechanism for the « 1994 Nanao earthquake » as well as for the
611 June 14, 2001 M6.3 event reported on figure 6 and section 4 (Fig.9). We believe that the
612 presence of this cluster right above the incipient tear within the PSP slab is not fortuitous. We
613 thus propose that the motion along the incipient tear (presumably normal with a sinistral
614 component) transfers a similar stress within the overriding plate that deforms accordingly (see
615 the 1966 seismic crisis above). One may notice that such sinistral component along transcurrent
616 faults is just the opposite that one may expect from strain partitioning within a forearc in a
617 context of oblique convergence (McCaffrey, 1992 ; Chemenda et al., 2000). This observation
618 leads us to suggest that strain partitioning localizing on these en-échelon shear zones is not
619 caused by oblique subduction but rather by collision of the PSP with EP south of them and
620 « free » subduction north of them. Such sinistral motion along faults either trench-parallel or
621 convergence-parallel are likely caused by the present collision between the PSP and the EP
622 south of 24°30'N as discussed by Lallemand et al. (1997).

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7- Discussion and conclusions

In this study, we have analysed a new set of data on the area offshore NE Taiwan including a new 3D velocity model, relocated earthquakes and two new seismic lines. Based on this detailed structural analysis, we can answer several questions regarding the geometry of the boundary between converging plates (EP, SRA and PSP), the deformation style in the area of plates interaction and finally the reason for the specific distribution of seismicity.

7.1- Precise geometry of EP/PSP and SRA/PSP boundaries

The three plates mainly interact offshore East Taiwan so that large uncertainties characterized earlier models before getting OBS constraints.

We propose a new map for the PSP subduction interface that accounts for the new dataset and that largely differs from previous ones mainly based on poorly constrained Benioff zone at shallow depths (e.g., Kao et al., 1998; Font et al., 1999; Wu et al., 2009a; Ustaszewski et al., 2012). Part of the discrepancy comes from the fact that the PSP is highly deformed west of 122°45'E in the prolongation of the western foot of the Gagua Ridge. We have shown that slicing and tearing of the oceanic crust accommodate a NW-SE shortening of the PSP. The cartoon on Fig.13 illustrates the tectonic forces acting in the EP-PSP system. Following Ustaszewski et al. (2012), we consider that the sharp continental subduction of the EP beneath the Central Range of Taiwan acts as a vertical wall against which the PSP collides. This “wall”, sketched with three thick arrows on Fig.13, is the present indenter of the PSP between Taitung and Nanao. It ends for us near 24°30'N, so that north of this latitude, the deformed PSP is free to subduct in the northwest direction. NW-SE shortening together with incipient subduction of the PSP beneath the eastern coast of Taiwan is probably at the origin of the bulge visible on both the PSP Moho and top (Figs.7 & 10) parallel to the east coast. Because of the obliquity between the LVA and the continent-ocean transition within the EP, collision started in the north and propagated toward the south (Fig.1; Suppe, 1984). We thus infer that the NW-SE compression and shortening increases from south to north. The combination of the contrast between shortening south of 24°30'N and “free” subduction north of this latitude with the strong slab pull expected from an old (and dense) oceanic plate favours the inception of a slab break off of the NW corner of the PSP. The present geometry indicates en-échelon ramps offsetting the plates interface that probably reveal sub-vertical shear zones that might evolve soon into a continuous tear. The seismic activity concentrates near these ramps.

Many authors have attempted to characterize the tectonic regime east of Taiwan (e.g., Wu et al., 2009b; Wu et al., 2010a&b) but the poor vertical resolution of earthquakes hypocenters conjugated with uncertainties in the affinity of the concerned domains, weakened the previous conclusions. Understanding the complex tectonics in this region requires to precisely map the frontiers between each colliding plate and thus know which deformation style is attributed to which plate. We did that properly when mapping the interface between the SRA and the PSP but we still have uncertainties below the northeast coastal area between Hualien and Ilan as indicated in figure 10 and in a lesser extent in mapping the Mohos beneath Suao and Ilan (see Fig.7). Based on 10 years of onland seismic records, Lin et al. (2004) have identified a channel-like body starting from a depth of 120 km north of Ilan and reaching about 20 km north of Hualien, i.e., along the supposed PSP slab edge. The material within that “channel” was characterized by low V_s and high V_p/V_s ratios that pushed the authors to interpret it as melting features caused by dehydration processes from the subducted plate. This narrow area, difficult to map, is deforming extremely rapidly (see the rapid changes in GPS motions within a small sector; e.g., Rau et al., 2008). It is also the locus of atypical magmatism such those of Turtle Island (R. Shinjo, pers. comm.). Our new 3D velocity model is not good enough at those great depths to resolve those questions.

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7.2- Deformation style

Each converging plate undergoes deformation that depends on its rheology and inherited structure.

Starting from the PSP, and especially the old Huatung Basin lithosphere which lies between the east coast of Taiwan and the Gagua Ridge, we observe a range of facts (earthquakes, crustal shape and thickness, kinematics) that argue in favour of its NW-SE shortening through folding and slicing. Both types of deformation occur essentially at the crustal level with the help of a weak (hydrated) upper lithospheric mantle. We have imaged a wide zone offshore NE Taiwan (see LVZ in Fig.7) with low velocity anomalies at depths far below the Moho of the PSP (Figs.8 & 9) that we interpret as evidences for intra-oceanic thrusting. The largest intra-PSP earthquakes mostly occur along its western edge (Fig.4) where the LVA undergoes strong compression. The Taiwan orogen results primarily from the indentation of the EP by the LVA producing an accretionary wedge made of crustal slivers originating from the Chinese platform (e.g., Suppe, 1981). This process is still active in the southern part of the island but our observations show that, today, the EP indents the PSP in central Taiwan (Fig.13) as suggested by Wu et al. (1997). The next step is a flip of subduction polarity that is starting in the Hualien region (south of 24°20'N) where incipient subduction of the PSP beneath the Central Range has been evidenced (Chemenda et al., 1997).

There has been a long debate on the fate of the LVA forearc basement since it is apparently absent all along the Coastal Range (e.g., Malavieille et al., 2002). Based on the root of the LVA which is reported in Fig.7, we observe that the axis of the LVA is offset in three en-échelon segments mostly offshore the east coast except near 23°25'N where the Chimei volcanic edifice outcrops onshore. The LVA is thus almost fully submerged offshore the east coast. It is possible to trace it below the steep submarine slope north of Hualien at larger depths since it is buried and underthrust below the Central Range of Taiwan (Figs.7 and 8). The en-échelon pattern may be inherited from the original morphology of the volcanic edifices as it appears south of Taiwan (Fig.1) but also the result of progressive underthrusting of arc slivers as proposed by Malavieille et al. (2002). A singular feature has been evidenced from the new velocity model which is the presence of a narrow high velocity zone (HVZ) west of the LVA. The HVZ also trends en-échelon parallel to the LVA segments. The surface projection of the HVZ coincides with the northern part of the Coastal Range and Longitudinal Valley (we did not investigate the southern part because it is out of our resolved area) as well as the eastern part of the Central Range north of Hualien. We thus propose that the HVZ represent relics of the LVA forearc basement squeezed between the EP and the PSP.

We did not focussed much on the EP deformation since it has been extensively studied using landstations (Wu et al., 2007a, 2009a). The new insights provided by our dataset rather concerns the transition between the EP and the SRA and the characterization of the southern edge of the SRA in the studied area. We have seen that the thickness of the SRA near Yonaguni and Iri-Omote islands averages 32±4 km, the thickest portion being truncated along the southern edge. We use the term “truncation” since generally, the island arcs are thicker below the islands than along the contact with the subducting plate. We also use the term “truncation” because the upper plate crust thickens when approaching Taiwan between 122°25'E and 121°55'E with isocontours normal to the continental Moho southern edge (Fig.7). This domain corresponds to the transition between the EP, considered as fixed in Fig.1, and the SRA moving south at rates 5 to 7 cm/yr. We did not find localized deformation areas such as N-S trending shear zones that would accommodate the southward drifting of the SRA with respect to Eurasia. We thus conclude that the deformation is distributed over a wide zone between the east coast of Taiwan and Yonaguni Island. The only localized shear zone that we have evidenced trends ESE-WNW offshore at the latitude of Nanao city (see Fig.1 and section 4 in Fig.9). It probably

731 corresponds with a sinistral strike-slip fault compatible with several focal mechanisms of
732 earthquakes in this region (Fig.4). It has been interpreted by Huang et al. (2012) as the result of
733 the bending of the northern Central range in response to the SOT opening. We suggest that the
734 interaction of the Central Range in depth by the LVA also contributes to this “extrusion
735 tectonics”. Trench-parallel sinistral motion along SRA faults has also been revealed by a M7.5
736 earthquake in 1966 (Fig.4). We consider that such transcurrent motion along faults, visible in
737 the detailed bathymetry (Lallemand et al., 1999), might have also been triggered by the growing
738 tear within the subducting PSP (see section 7.1). Since the aftershocks of the 1966 earthquake
739 occurred within the subducting slab and delineated a WNW-ESE left-lateral shear zone, we
740 conclude that there is a strong coupling between the plates in this region. The last point
741 concerning the southern edge of the SRA basement is that HVZ have been found at shallow levels
742 north of the Hoping and Nanao basins. The occurrence of these HVZ together with the
743 “truncation” evidences of the southern edge of the SRA basement argue in favour of relics of the
744 Luzon volcanic forearc (and arc ?) incorporated at the toe of the SRA. We suggest that the former
745 northern passive margin of the SCS was strongly shortened/thickened and then eroded
746 tectonically when the Ryukyu trench propagated westward during the last 5-8 m.y. (Lallemand
747 et al., 2001) and that pieces of the LVA and forearc were accreted at the front of the truncated
748 newly active margin. The baionet shape of the SRA and forearc basins (Fig.1) is, at least partly,
749 the result of uplift of the SRA basement (Hoping and Nanao rises; see Fig.12) in response to the
750 shortening and thickening of the subducting PSP. Crustal accretion of the former LVA beneath
751 the toe of the SRA basement might also have contributed in the observed uplifted zone.

752 753 **7.3- Seismicity clustering and maximum potential earthquake magnitude**

754
755 One major question concerns the possible occurrence of great earthquakes ($M > 8.5$) in this
756 region characterized by a high seismic rate. We already know that the largest instrumentally
757 recorded earthquake in this region (including Taiwan) reached a moment magnitude 7.7
758 (Theunissen et al., 2010). This event in 1920 occurred at the junction between the Suao and
759 Hoping clusters (Fig.4). Two other large subduction earthquakes occurred since 1920 with a
760 ~ 40 yrs recurrence interval in the same area (Fig.4) : the 1963 $M_w 7.2$ and 2002 $M_w 7.1$ events.
761 Based on our study, we interpret the concentration of the seismicity within both clusters as the
762 result of the down-faulting of the PSP that creates ramps triggering the nucleation of
763 earthquakes along the subduction interface. The upper plate deforms accordingly to
764 accommodate the step-geometry of the subducting plate producing a seismic deformation right
765 above the ramps. Most of the seismic energy thus appear to dissipate along several crustal (or
766 lithospheric) faults cutting through the PSP (Fig.13) and the SRA (Fig.9). Based on interseismic
767 GPS data along the Coastal Range, Hsu et al. (2012) suggest that the aseismic shallow
768 southernmost Ryukyu subduction interface is fully locked. Considering a slip deficit accumulated
769 over 500 years, they predict a potential earthquake magnitude up to 8.7 if the rupture goes from
770 the east coast of Taiwan to the Gagua Ridge. To reach those conclusions, they made the
771 hypothesis that the accumulated strain on the east coast was fully accounted by fault coupling
772 along a unique surprisingly shallow (between 6 and 20 km, see section 2.2) megathrust. Our
773 results clearly indicate that a large part of the slip deficit is absorbed along intra-oceanic thrusts.
774 Furthermore, the shallow subduction interface is segmented into surfaces with trench-parallel
775 lengths smaller than 120 km (distance between east coast and Gagua), typically 60 km or less.
776 The GPS approach is limited by the constraints of using onland stations (same limitation for the
777 seismic approach before using OBS). To test the geodetic solutions would require the
778 deployment of seafloor crustal deformation observatories (Ando et al., 2009). To conclude, the
779 complex geometry of the subduction interface west of 123°E (northern prolongation of the
780 Gagua Ridge) does not militate in favour of large ruptures, confirming the conclusions of Kao et
781 al. (1998) and Theunissen et al. (2010).

782 **Figure captions**

783

784 Fig.1 : Geodynamic context of Taiwan with orthogonal opposite subductions between the
785 Philippine Sea Plate (PSP) and the Eurasia Plate (EP). East of Taiwan, the rift of the Southern
786 Okinawa Trough (SOT) separates the main EP from the southward drifting South Ryukyu Arc
787 (SRA). Arrows indicates the motion of some GPS stations with respect to the Penghu islands
788 after Huang et al. (2010) for the eastern coast of Taiwan and Nakamura (2004) for the SRA. V_s
789 are subduction relative velocities between converging plates (EP/PSP or PSP/SRA). H.B. =
790 Hoping Basin; H.R. = Hoping Rise; N.B. = Nanao Basin; N.R. = Nanao Rise; T.I. = Turtle Island;
791 C.O.T. = continent-ocean transition; SCS = South China Sea.

792

793 Fig.2 : 3D artist representation of the EP and SRA plates near Taiwan modified after Lallemand
794 et al. (2001). Note the very steep subducting EP beneath central Taiwan partly explained by a
795 tear propagating from north to south. Both cartoons are purely conceptual and the ideas behind
796 are discussed in this paper.

797

798 Fig.3 : 12 CNRS-INSU short period (SP) 4 components OBS (red dots) and 3 NTOU SP micrOBS
799 (yellow dots) passive array deployed between July 19 and October 24, 2008 together with 7 US
800 broadband (BB) OBS deployed during one year in the frame of the NSF TAIGER program (black
801 dots). 24 micrOBS linear array deployed during the active experiment in May 2009, 19 SP
802 micrOBS from IFREMER (white dots) and 5 from NTOU (blue dots).

803

804 Fig.4 : Relocated hypocenters of 5996 $M > 3$ events recorded by the CWB and the JMA during the
805 period 1992-2008 (east of 121°E) using the new 3D local tomography (Theunissen et al., 2012a)
806 and the new version of MAXI-code (Theunissen et al., 2012b). Three shallow seismic clusters
807 mentioned in figures 8 and 9 are outlined : SC = Suao cluster, HC = Hoping cluster, NC = Nanao
808 cluster. Multichannel seismic lines acquired during the active TAIGER-RATS experiments and
809 shown on figures 11 and 12 are drawn : MCS-12 and MCS-22. A selection of major historical
810 earthquakes is reported with their date of occurrence, revised moment magnitude (Theunissen
811 et al., 2010) except for the 24/02/06 event (CWB solution) and focal mechanisms estimated
812 from Harvard CMT in black, BATS in red and literature in grey (Kao et al., 1998; Chen et al.,
813 2004; Cheng et al., 1996; the focal mechanism of the 1920 earthquake does not exist, we only
814 suggest that it was close to the Harvard CMT of an analog event : 9/10/1994 $M_w 4.9$ based on
815 similar arrival times on common seismic stations, see Theunissen et al., 2010). This position is
816 very close (< 10 km in horizontal and same depth) of the 31/3/2002 $M 7.1$ thrust event.
817 Hypocenters of earthquakes younger than 1991 were relocated but not those older than 1992.
818 Location of the tomography lines shown on figures 8 and 9. See Table 1 for earthquakes details.

819

820 Fig.5: Hypocenter comparison between CWB and our earthquakes relocation for all earthquakes
821 having a well-determined focal mechanism along forearc sections 5, 6 and 7 (see location in
822 Fig.4). In gray, the final selection of the 5996 relocated earthquakes. Yellow dots are relocated
823 earthquakes having a focal mechanism. Blue squares are the same earthquakes located by CWB.
824 Focal mechanism: black (GCMT), gray (Kao et al., 1998) and red (BATS,
825 <http://bats.earth.sinica.edu.tw/>). The focal mechanism of the 1920 event with a question mark
826 is based on the nearest earthquake having a focal mechanism compatible with the best
827 candidate fault (see Theunissen et al., 2010). See Table 1 for earthquakes details.

828

829 Fig.6 : A: Seismic network used in this study including land and marine stations (in blue) as well
830 as earthquakes recorded during the 3-months deployment (in red). The rays between
831 earthquakes and stations indicate the spatial coverage of the passive experiment. B: 3D sketch of
832 local P-waves tomographic sections with earthquakes (grey dots) and the area crossed by more
833 than 50 rays per cell has been enlightened.

834

835 Fig.7 : Isodepths of the EP Moho (in black) and PSP Moho (in white) mostly estimated from
836 picking based on reference isocontours at $V_p = 7.5$ km/s for the PSP and at 7.75 km/s for the EP
837 and SRA into the new 3D V_p model. Seriated vertical sections within the 3D V_p model have been
838 used to discriminate between both plates in the area of plates interaction. The black dashed line
839 represents the southernmost extent of the continental Moho beneath the SRA as well as the
840 subvertical boundary between EP and PSP beneath the Taiwan orogen. Black crosses mark the
841 axis of the LVA, whereas small red dots delineate the regions of « high V_p rise » that may reflect
842 the suture in the collision zone. H.B. = Hoping Basin, H.R. = Hoping Rise, N.B. = Nanao Basin, N.R.
843 = Nanao Rise, E.N.B. = East Nanao Basin.
844

845 Fig.8 : Three sections within the new 3D tomography model in the azimuth of the convergence
846 between EP and PSP (see Fig.4 for location). The enlightened areas correspond to the inverted
847 volume based on a minimum of 50 rays crossing the cells, whereas the pale regions almost
848 correspond to the initial model (see the text for further details). EP = Eurasia Plate, PSP =
849 Philippine Sea Plate, SRA = South Ryukyu Arc, LVA = Luzon Volcanic Arc, HVZ = High Velocity
850 Zone, LVZ = Low Velocity Zone. The heavy dashed line in black represents the presumed
851 boundary between EP and PSP. The areas outlined by a thin dotted line in yellow represents low
852 velocity zones within the PSP mantle. Significant $M \geq 6$ earthquakes that occurred close to the
853 sections are plotted with their focal mechanisms (vertical projection in the section's plane), their
854 date of occurrence, their moment magnitude and their focal depth. The same events are plotted
855 in map view on Fig.4. The section 3 is not far from the MCS line 22 shot during the US-Taiwan
856 TAIGER survey in 2009. See Table 1 for earthquakes details.
857

858 Fig.9 : Four sections within the new 3D tomography model normal to the Ryukyu Trench. Same
859 legend as figure 8. Focal solutions in forearc sections 5, 6 and 7 appear in figure 5. The inset
860 within the section 5 represents the velocity model without interpretation in order to outline the
861 subvertical offset within the subducting plate. The section 6 coincides with the wide-angle
862 refraction ACTS line and the MCS line 12 shot during the US-Taiwan TAIGER survey in 2009.
863

864 Fig.10 : Isodepths of the top of the PSP. More than 50 seriated vertical sections within the 3D V_p
865 model have been used to constrain the subduction interface north of the Ryukyu Trench. The
866 blue line outline the Ryukyu Trench and the Longitudinal Valley that marks the limit between EP
867 and PSP onland. The faults in red were interpreted based on the seriated sections except the
868 dextral strike-slip in the Huatung Basin which comes from Schnürle et al. (1998). The paleo-
869 WPB ridge axis is extrapolated from magnetic lineations south of the trench based on
870 Deschamps and Lallemand (2002). As for the PSP Moho map on Fig.7, the change in trend of the
871 isocontours west of Yonaguni Island might represent the limit between the WPB and the older
872 Huatung Basin.
873

874 Fig.11 : Migrated depth section of multichannel seismic line 12 acquired during the TAIGER-
875 ACTS project. See location on figure 4. The relocated earthquakes and the interpretation of the
876 faulted subduction interface have been reported from section 6 of Fig.9 for comparison with the
877 velocity model. Grey dots : relocated 1992-2008 seismicity (see Fig.4). White dots : 2008
878 microevents recorded during the 3-months RATS experiment.
879

880 Fig.12 : Unmigrated TWT-time section of multichannel seismic line 22 acquired during the
881 TAIGER project. See location on figure 4 and section 3 of figure 8 for comparison with the
882 velocity model and seismicity distribution. Relocated earthquakes were not reported on this line
883 because they would mask the normal faults affecting the Nanao Rise.
884

885 Fig.13 : Conceptual sketch illustrating the deformation of the PSP when it interacts with
886 the Taiwan orogen.
887

888 Fig.S1 : Depth difference between CWB locations and our earthquakes relocations. This map was
889 built by using only the final selection of 5996 earthquakes for which a block median filter with
890 an increment of 5 km has been applied to the depth difference. SC = Suao cluster, HC = Hoping
891 cluster, NC = Nanao cluster.

892
893 Table 1 : List of parameters and source for focal mechanisms of earthquakes shown in figures 4,
894 5, 8 and 9. BATS means Broadband Array in Taiwan for Seismology, see Kao and Jian (2001) for
895 more details. CMT means Global Centroid Moment Tensor (<http://www.globalcmt.org/>). The
896 focal mechanism of the 1920 event is unknown but, based on Theunissen et al. (2010), the best
897 candidate for this event is the seismogenic subduction interface. We thus suggest that the
898 shallow-dipping thrust focal mechanism of the nearest earthquake after new 3D relocation, i.e.,
899 M7.1 31/03/02 event, is similar to those of the 1920 earthquake.

900
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914
915

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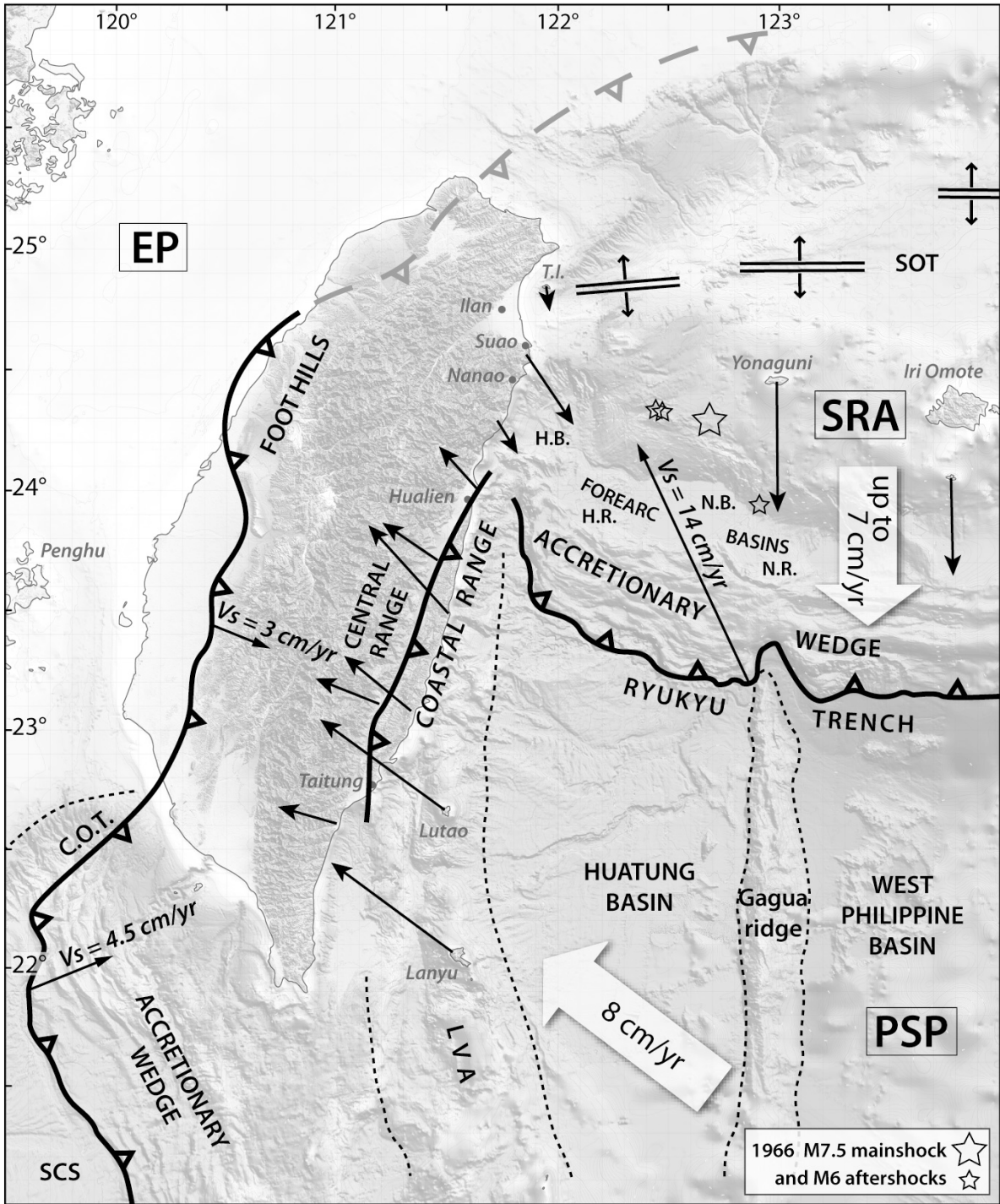
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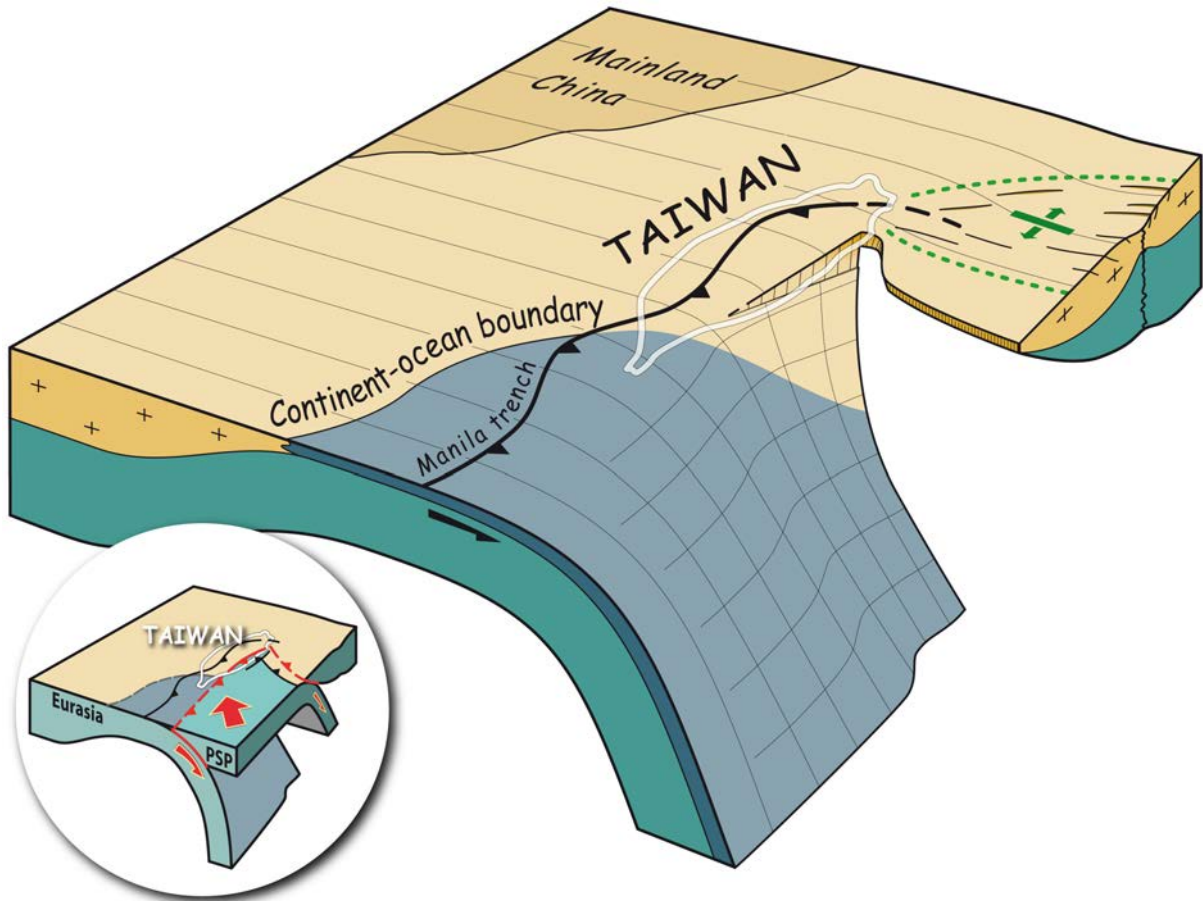
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1171 Appendix
1172 The procedure used to relocate the seismic events with MAXI-code is fully explained in Font et al.
1173 (2004) and Theunissen et al. (2012b). As said in the text, solutions are sorted according to
1174 confidence factors based on MAXI algorithm rather than rms error based. We chose equivalent
1175 criteria namely $Q_{EDT} \geq 0.6$ (ratio between 0 and 1 describing quality of Equal Differential Time
1176 volumes intersections); inter-barycenter distance (between volumes V1 and V3) ≤ 5 km; V1 size
1177 $\leq 400 \text{ km}^3$; and V2 size $\leq 200 \text{ km}^3$. After selection, average rms was $0.20 \pm 0.03(1\sigma)$ and average
1178 uncertainties (χ^2) dx, dy and dz were 0.94 ± 0.51 km, 1.25 ± 0.84 km and 2.25 ± 1.69 km (1σ)
1179 respectively. The final search volume, V4, may indicate also a kind of uncertainties: in particular
1180 V4 diameter spreads in depth from 5.1 ± 2.7 km (up) to 5.4 ± 2.7 km (down). In the final selection
1181 (5996 earthquakes), primary azimuthal gap ranges between 38° and 315° , with an average value
1182 of about $133 \pm 29^\circ$ (1σ). Distance to the first station is in average of 36 ± 23 km. There are $14 \pm 7\%$
1183 of outliers. Compared with the CWB catalog, relocations are distant (in 3D), on average, by 15.2
1184 ± 14.6 km (1σ) (median about 12 km). Solutions are sorted according to confidence factors based
1185 on MAXI algorithm rather than rms error. We observe that our depths of relocation are, on
1186 average, 6.3 ± 16.0 km (1σ) (median of 4.4km) deeper than those of CWB. The median of depth
1187 difference between CWB hypocenters and our relocations is illustrated on figure S1.
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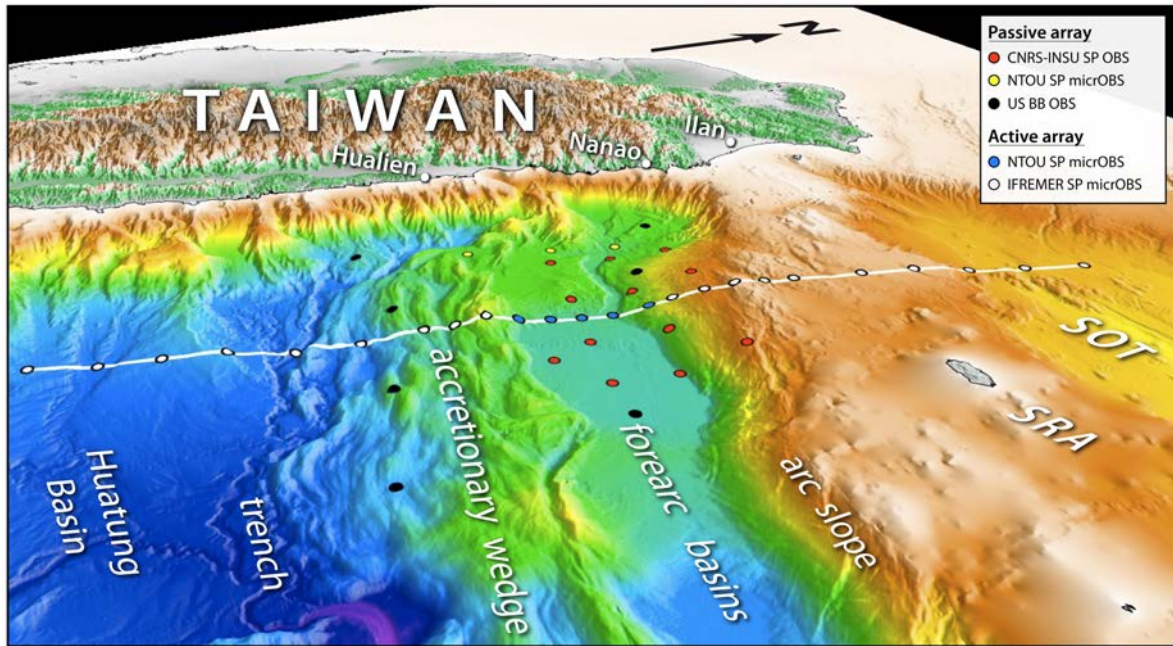


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Fig.1

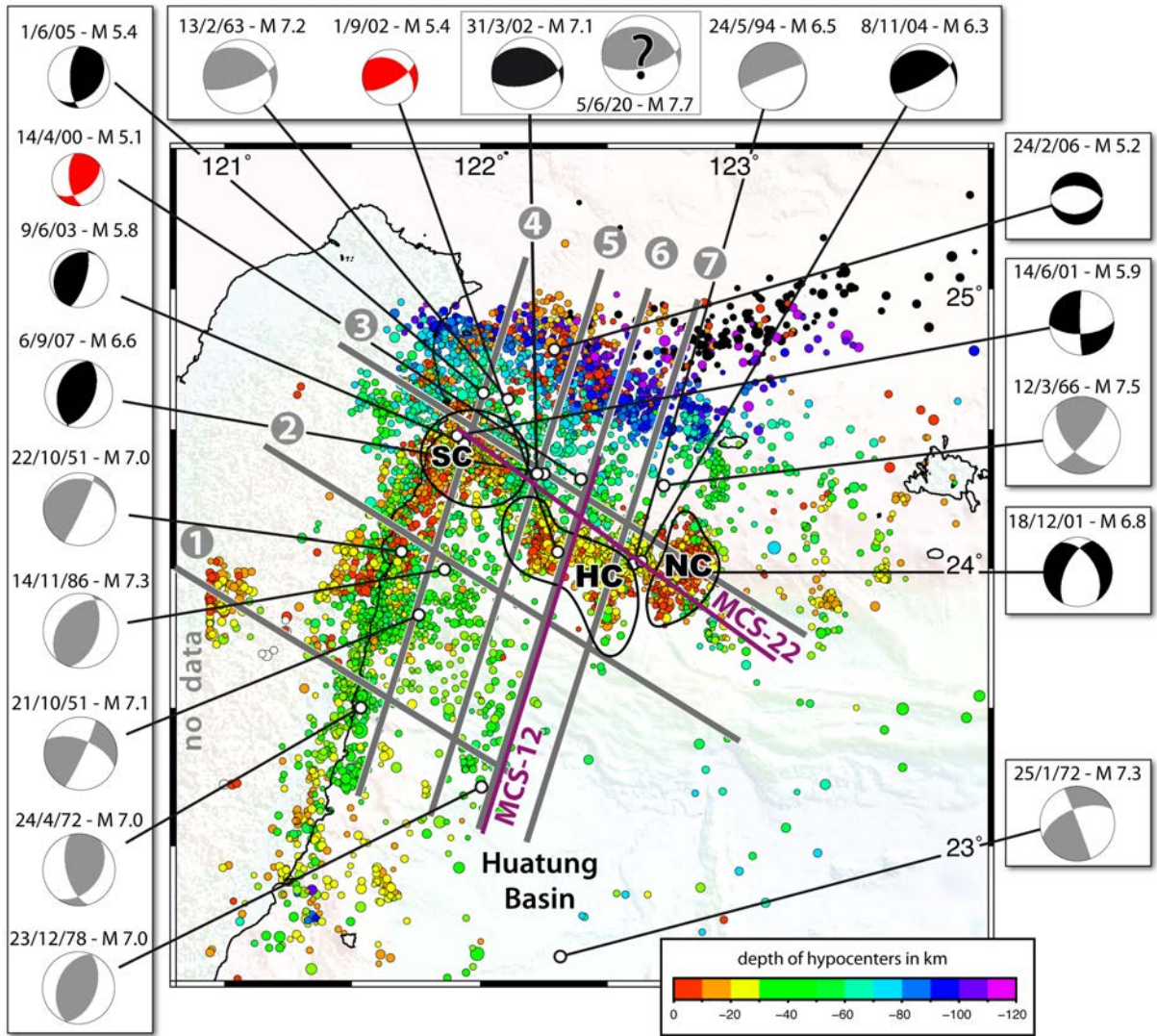


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1194 Fig.2
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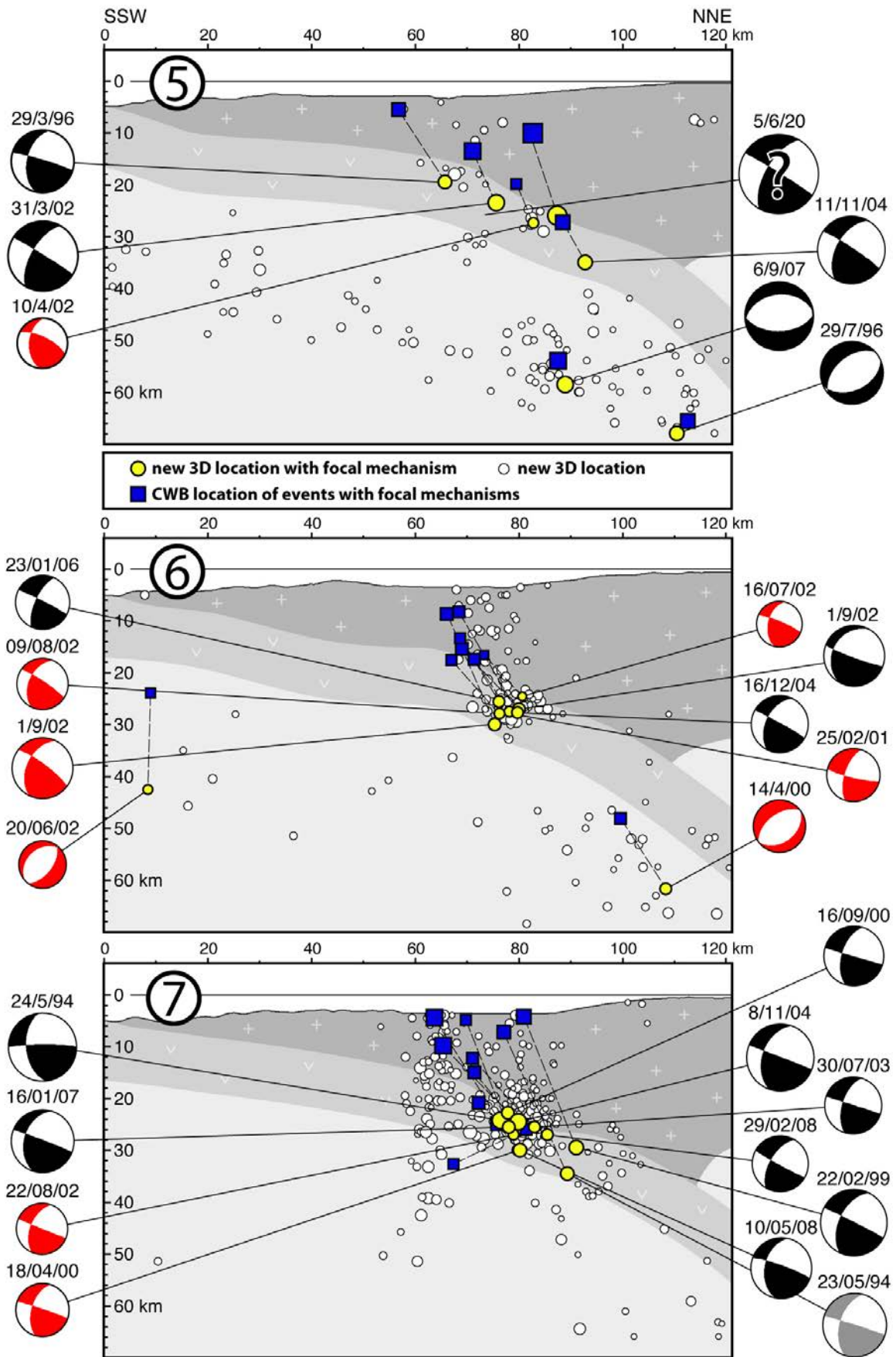
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Fig.3



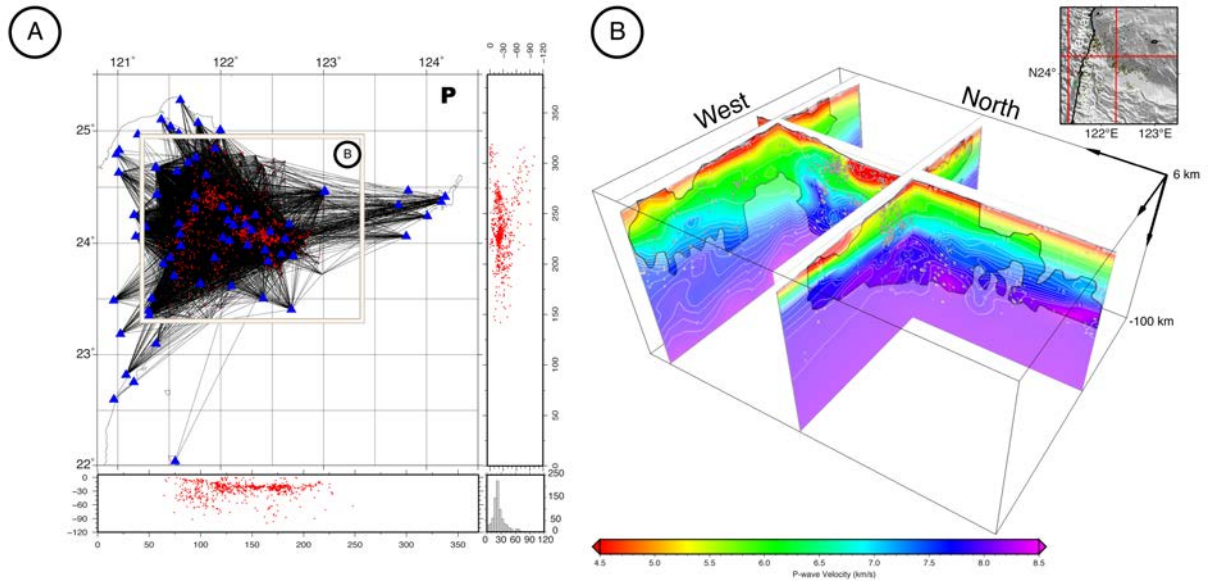
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Fig.4



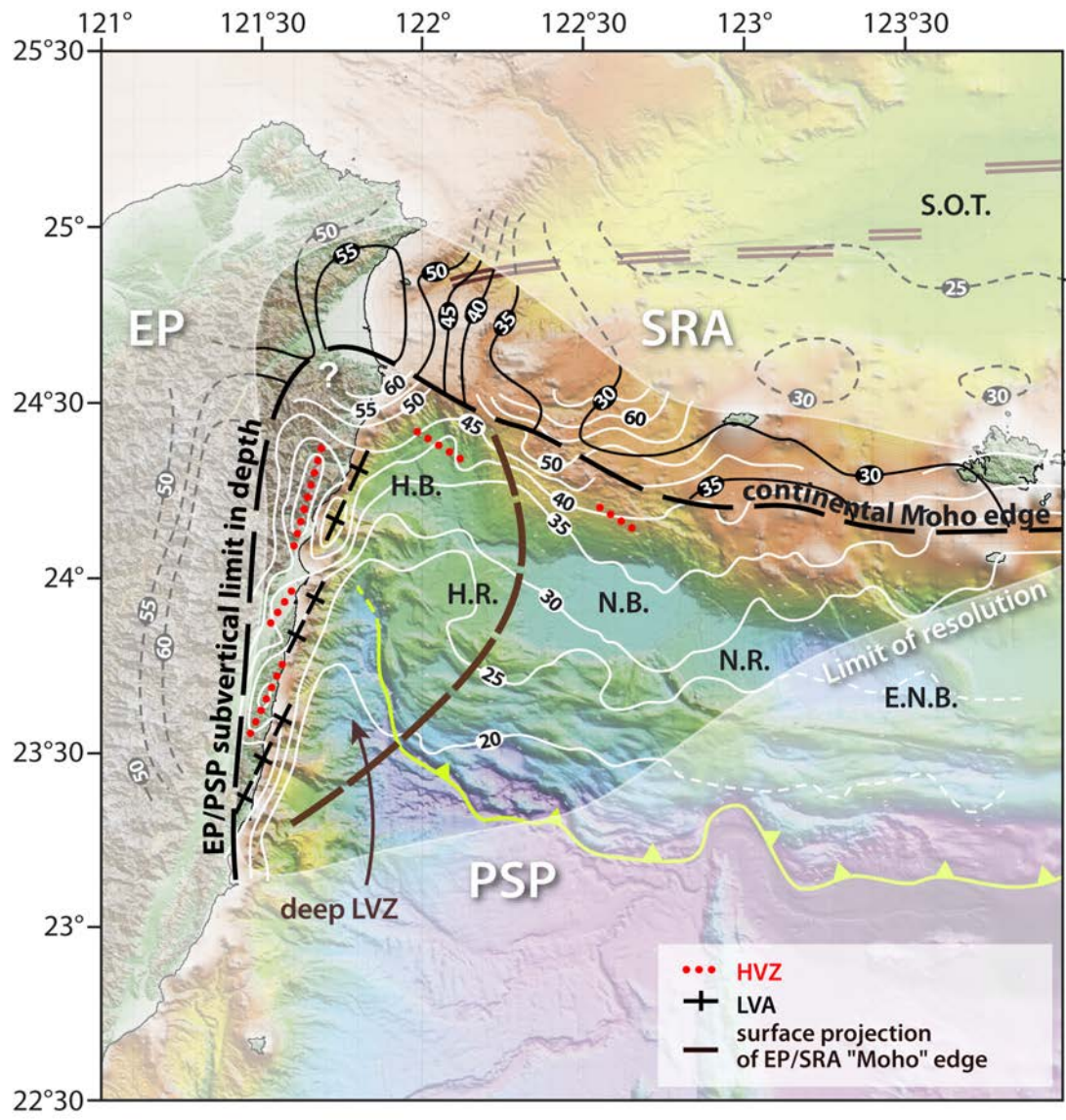
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Fig.5

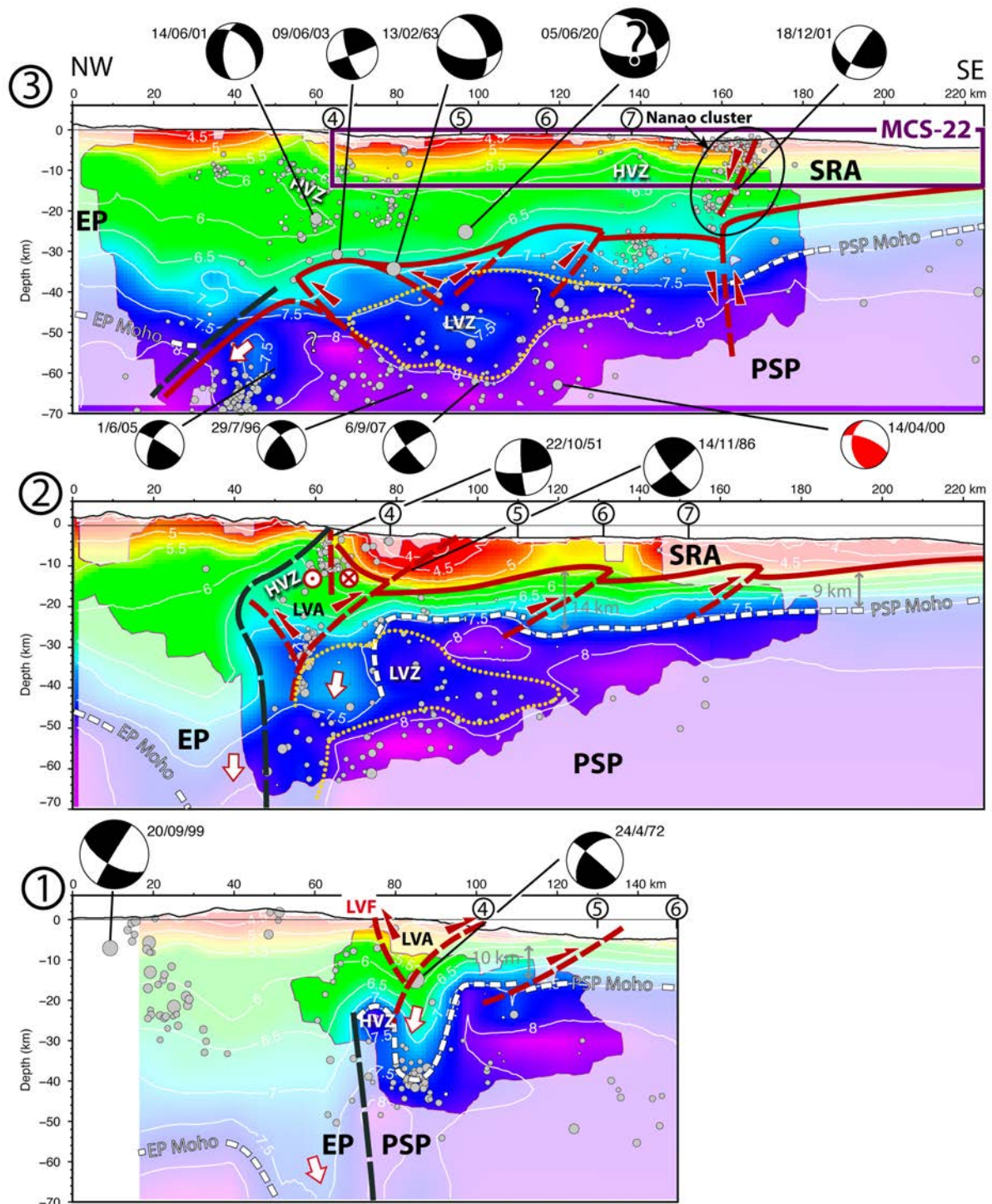


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Fig.6

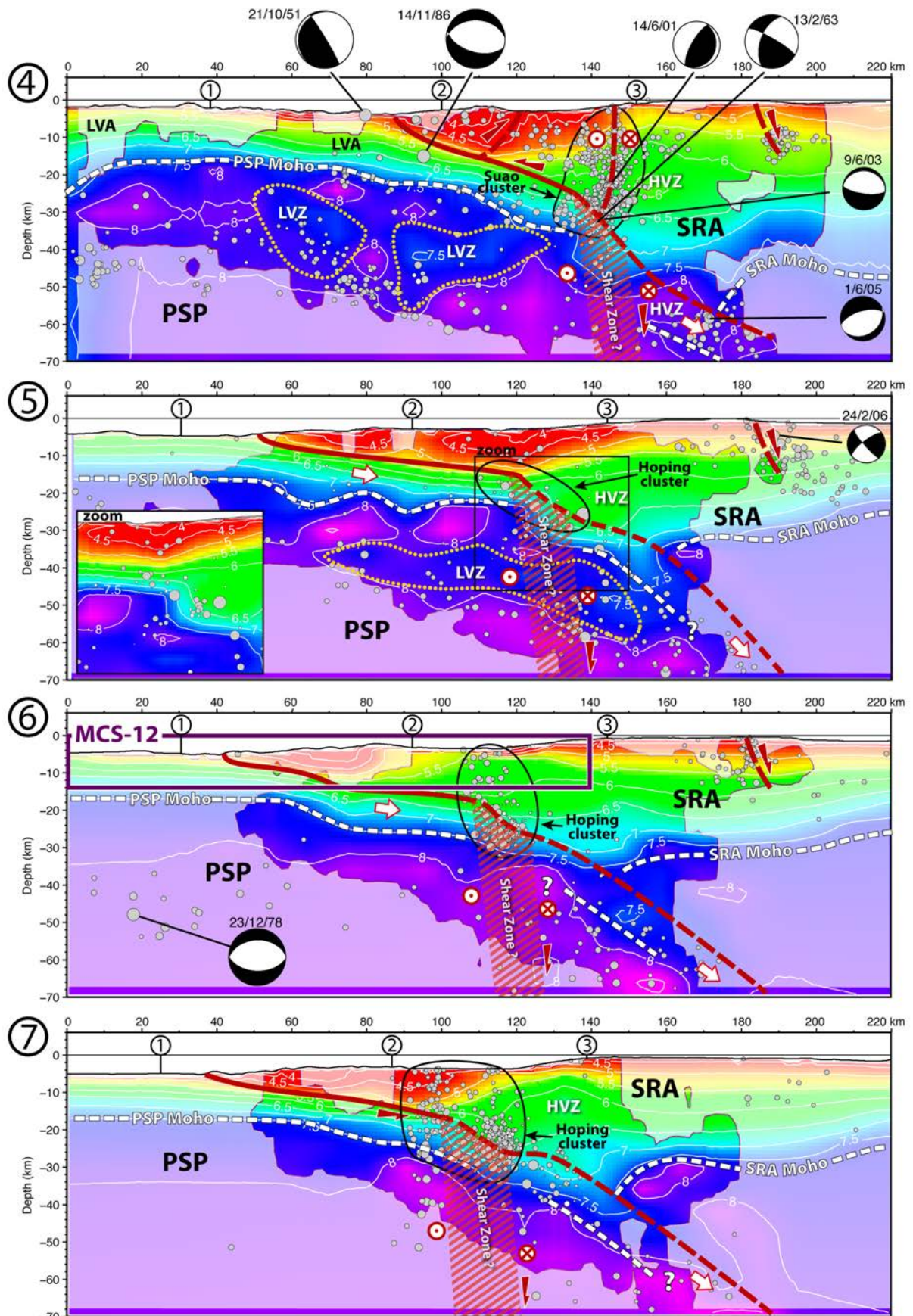


1208 Fig.7
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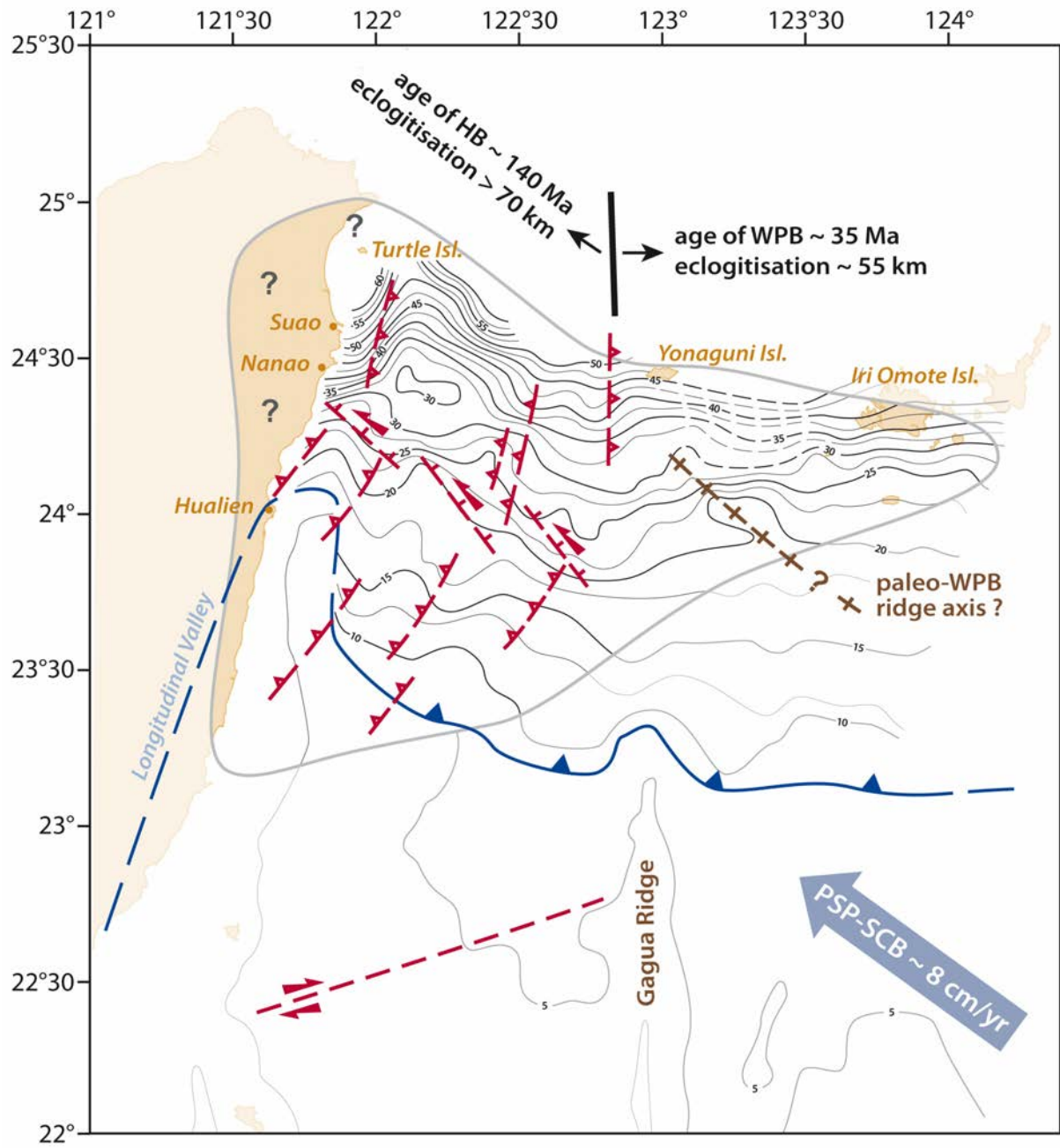
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Fig.8



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Fig.9



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 1218 Fig.10
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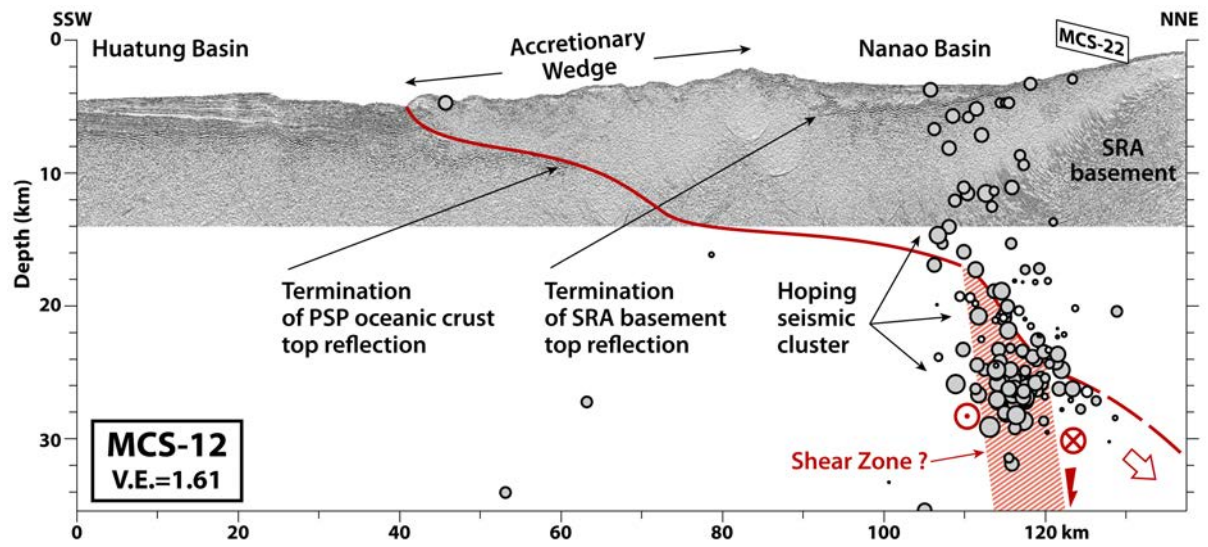
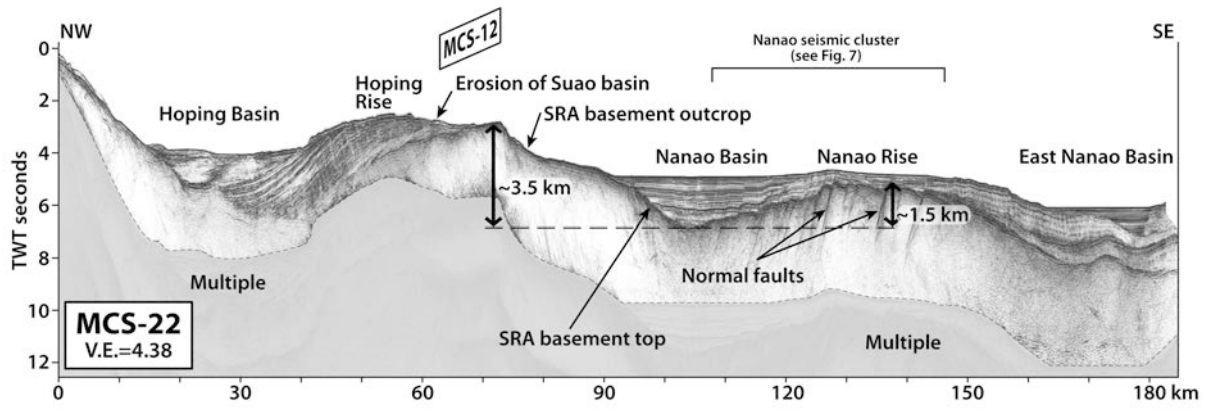


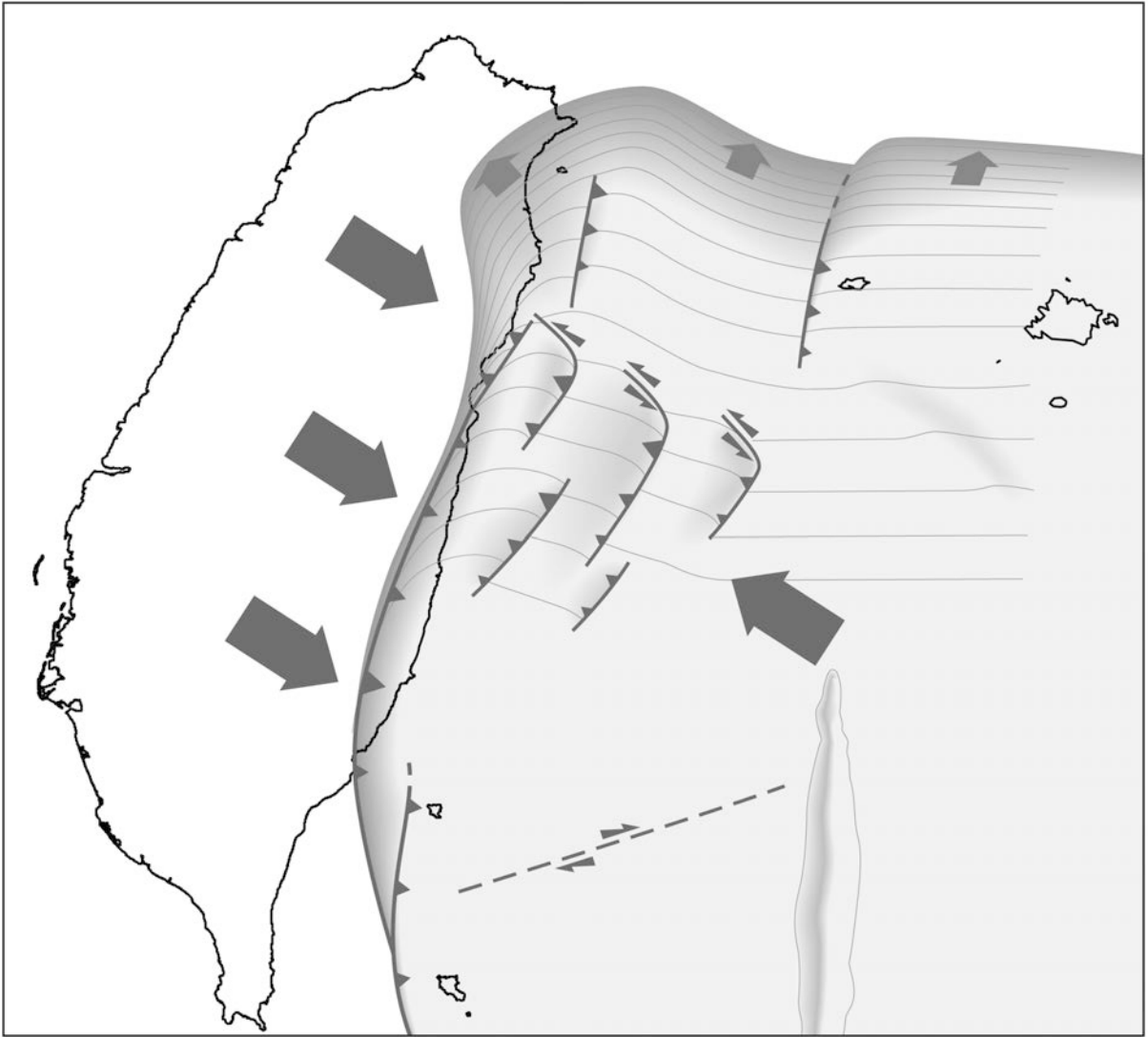
Fig.11

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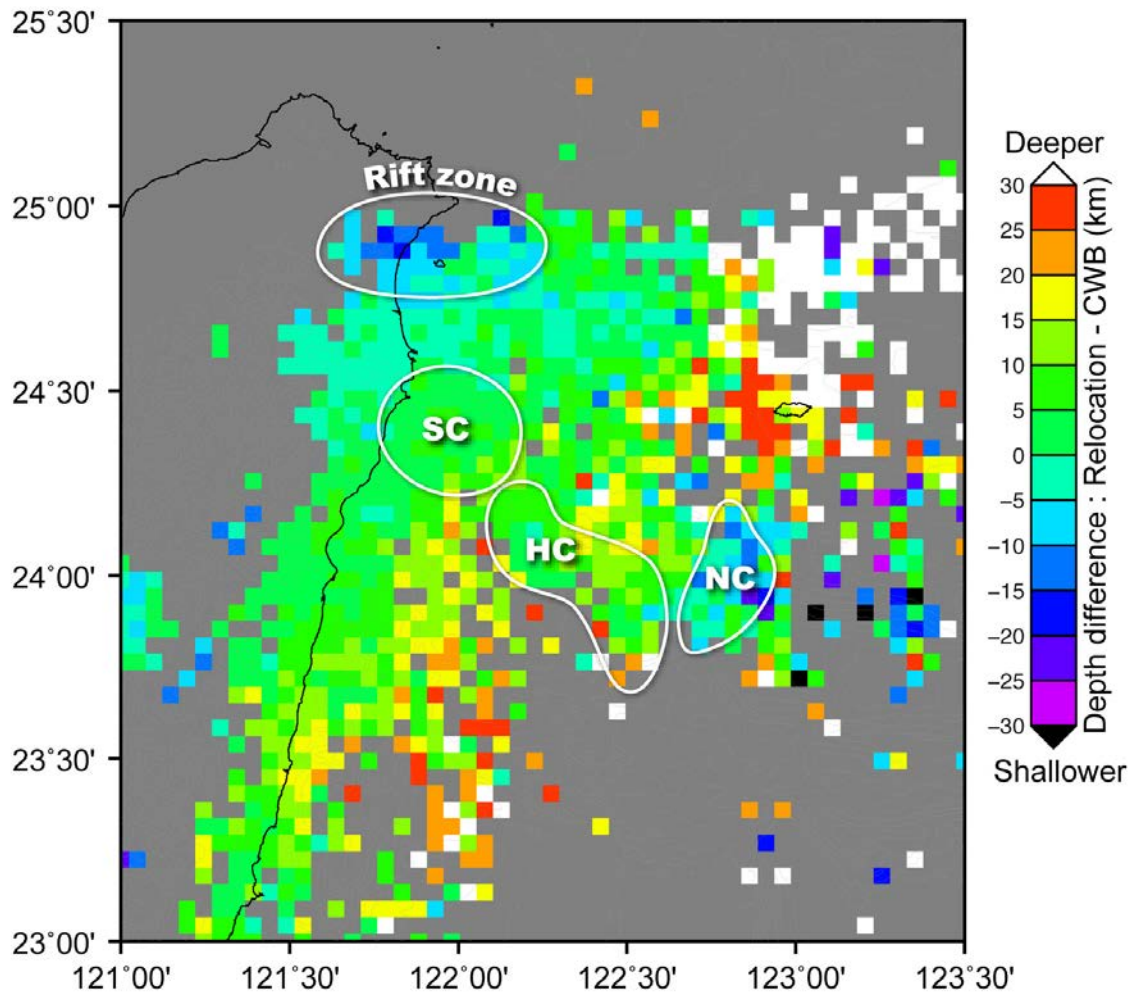
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Fig.12



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Fig.13



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Fig.S1

date of event day/month/year/ (time)	magnitude		estimated depth (km)		preferred source for focal mechanism
	revised moment magnitude M'_w (Theunissen et al., 2010)	local magnitude M_L (Central Weather Bureau)	from Central Weather Bureau	this study otherwise the origin is mentioned	
10/05/2008		5.6	25	30	CMT
29/02/2008		5.1	26	27	CMT
06/09/2007	6.2	6.6	54	58	CMT
16/01/2007	5.2	5.4	21	25	CMT
24/02/2006		5.2	13	7	CMT
23/01/2006		4.9	18	26	CMT
01/06/2005	5.4	6.0	65	58	CMT
16/12/2004		5.1	9	28	CMT
11/11/2004	5.6	6.1	27	35	CMT
08/11/2004	6.3	6.6	10	24	CMT
30/07/2003	5.2	5.1	12	25	CMT
09/06/2003	5.8	5.7	32	31	CMT
01/09/2002 – 7h07	5.3	5.5	16	27	CMT
01/09/2002 – 5h56	5.4	5.5	9	30	BATS
22/08/2002		4.7	33	27	BATS
09/08/2002		4.6	18	28	BATS
16/07/2002		4.1	17	24	BATS
20/06/2002		4.2	3	42	BATS
10/04/2002		4.5	20	27	BATS
31/03/2002	7.1	6.8	14	23	CMT
18/12/2001	6.8	6.7	12	15	CMT
14/06/2001	5.9	6.3	17	22	CMT
25/02/2001		4.7	14	27	BATS
16/09/2000	5.2	5.3	15	23	CMT
18/04/2000		4.6	5	30	BATS
14/04/2000	5.1	5.2	48	62	BATS
20/09/1999	7.6	7.3	8		CMT
22/02/1999	5.9	5.9	4	29	CMT
29/07/1996	5.4	6.1	66	68	CMT
29/03/1996	5.7	5.6	6	19	CMT
24/05/1994	6.5	6.6	5	24	Kao et al., 1998
23/05/1994	6.1	5.8	7	34	Kao et al., 1998
14/11/1986	7.3	7.8	15		Kao et al., 1998
23/12/1978	7.0	7.0	4	48 – EHB	Kao et al., 1998
24/04/1972	7.0	6.7	15		Kao et al., 1998
25/01/1972	7.3	7.0	33	10 – EHB	Kao et al., 1998
12/03/1966	7.5	7.8	42	29 – EHB	Kao et al., 1998
13/02/1963	7.2	7.4	47	35 – EHB	Chen et al., 2004
22/10/1951	7.0	7.2	20	1 – Cheng et al., 1996	Cheng et al., 1996
21/10/1951	7.1	7.4	9	4 – Cheng et al., 1996	Cheng et al., 1996
05/06/1920	7.7	8.0	20	25	see table legend

Table 1