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Complex slab structure and arc magmatism beneath the Japanese Islands

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ABSTRACT

A dense nationwide seismic network recently constructed in Japan has resulted in the production of a large amount of high-quality data that have enabled the high-resolution imaging of deep seismic structures in the Japanese subduction zone. Seismic tomography, precise locations of earthquakes, and focal mechanism research have allowed the identification of the complex structure of subducting slabs beneath Japan, revealing that the subducting Philippine Sea slab underneath southwestern Japan has an undulatory configuration down to a depth of 60–200 km, and is continuous from Kanto to Kyushu without disruption or splitting, even within areas north of the Izu Peninsula. Analysis of the geometry of the Pacific and Philippine Sea slabs identified a broad contact zone beneath the Kanto Plain that causes anomalously deep interplate and intraslab earthquake activity. Seismic tomographic inversions using both teleseismic and local events provide a clear image of the deep aseismic portion of the Philippine Sea slab beneath the Japan Sea north of Chugoku and Kyushu, and beneath the East China Sea west of Kyushu down to a depth of ~450 km. Seismic tomography also allowed the identification of an inclined sheet-like seismic low-velocity zone in the mantle wedge beneath Tohoku. A recent seismic tomography work further revealed clear images of similar inclined low-velocity zones in the mantle wedge for almost all other areas of Japan. The presence of the inclined low-velocity zones in the mantle wedge across the entirety of Japan suggests that it is a common feature to all subduction zones. These low-velocity zones may correspond to the upwelling flow portion of subduction-induced convection systems. These upwelling flows reach the Moho directly beneath active volcanic areas, suggesting a link between volcanism and upwelling.

1. Introduction

The Japanese Islands are located in a region of subduction zones, where earthquakes occur frequently including many destructive earthquakes that have caused extensive damage to inhabited areas. The M9.0 11 March 2011 great Tohoku-oki earthquake, which occurred along the plate interface east of Tohoku, northeastern Japan, is an example of a significantly destructive earthquake in the region. This was the largest-magnitude event in the modern history of Japan and caused severe damage to eastern Japan, resulting in ~20,000 dead and missing.

Four tectonic plates converge beneath the Japanese Islands (Fig. 1), comprising two oceanic plates subducting beneath two continental plates. Northeastern Japan is located on the southernmost portion of the North American Plate, with southwestern Japan forming part of the eastern edge of the Eurasian Plate. These two plates converge along the Itoigawa-Shizuoka Tectonic Line in the south and along the eastern margin of the Japan Sea in the north. Beneath southwestern Japan, the Philippine Sea Plate subducts northwestward at a rate of 3–5 cm/yr along the Sagami trough in the east, along the Nankai trough in the west, and along the Ryukyu trench in the southwest (Heki and Miyazaki, 2001; Miyazaki and Heki, 2001; Seno et al., 1993, 1996; Wei and Seno, 1998). Beneath northeastern Japan, the Pacific Plate subducts west-northwestward at a rate of 8–9 cm/yr along the Kuril trench in the north and along the Japan trench in the middle, and at a rate of ~6 cm/yr along the Izu–Bonin trench in the south (DeMets et al., 1994).

This complex tectonic convergence generates not only significant and frequent seismic activity, but also leads to a complex structure of subducting slabs beneath the Japanese Islands, in addition to volcanism. A recently constructed dense (station separation of ~20 km) nationwide seismic network, called "Kiban seismic network", covers all of the Japanese Islands, enabling the imaging and determination of deep structures beneath the region. Data from this network have been used in a number of seismic structural studies, the results of which are summarized here. The

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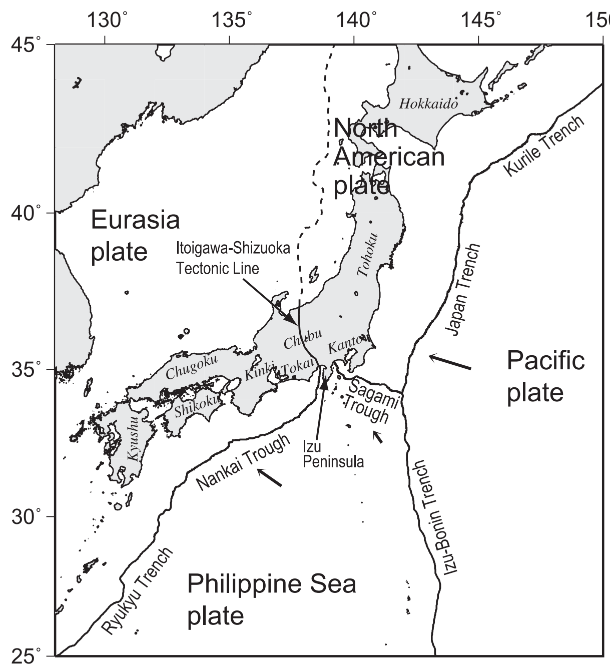

Fig. 1. Tectonic setting of Japan. The plate boundaries such as the trenches, troughs and tectonic lines are shown by thick solid or broken lines. Thick arrows indicate directions of the plate motion relative to the NA plate.

implications of these data for arc magmatism and the form of subduction beneath Japan are discussed in detail.

2. Subduction of the Pacific and Philippine Sea Plates beneath the Japanese Islands

2.1. Configuration of the Pacific Plate

Intermediate-depth intraslab earthquakes within the Pacific slab form a deep double-planed deep seismic zone beneath Hokkaido, Tohoku, and Kanto in northeastern Japan (Tsumura, 1973; Umino and Hasegawa, 1975; Hasegawa et al., 1978a; Suzuki et al., 1983). Hasegawa et al. (1978b) undertook the first investigation of the geometric relationship between this double-planed seismic zone and the subducting Pacific slab using converted ScS-to-P waves (ScSp phase) at the plate interface beneath Tohoku, determining that the upper interface of the subducting plate lies immediately above the upper seismic plane of the double-planed seismic zone. This observation was subsequently confirmed by Matsuzawa et al. (1986, 1990), who detected converted S-to-P and P-to-S waves (SP and PS phases) at the plate interface in seismograms from intermediate-depth intraslab events, using the arrival times of these waves to locate the upper interface of the subducting plate. Later, Zhao et al. (1997b) estimated the depth distribution of the upper interface of the subducting Pacific slab beneath Tohoku by inverting converted wave arrival-time data. These studies indicate that, at depths greater than ~50 km, the upper envelope of intermediate-depth intraslab seismicity approximately coincides with the upper surface of the subducting Pacific Plate.

Nakajima et al. (2009a) and Kita et al. (2008) determined the configuration of the upper surface of the subducting Pacific Plate down to a depth of ~500 km beneath all of Japan by relocating earthquake hypocenters observed by the nationwide Kiban seismic network and by using focal mechanism information for those earthquakes. They located plate interfaces at depths greater than ~50 km by adopting the upper envelope of relocated intraslab earthquakes as the upper plate interface, in accordance with the previous research. The plate interface at depths shallower than ~50 km was located by using the relocated hypocenters of interplate earthquakes, such as those with low-angle thrust-type focal mechanisms and small repeating earthquakes (Uchida et al., 2003)—seismic events that are thought to occur at the plate interface. The geometry of the subducting Pacific Plate beneath the Japanese Islands is shown as contours in Fig. 2.

2.2. Configuration of the Philippine Sea Plate

Until recently, the configuration of the Philippine Sea slab subducting beneath southwestern Japan was estimated by assuming the upper envelope of the distribution of intraslab earthquake hypocenters represented the upper slab interface (e.g., Mizoue et al., 1983; Yamazaki and Ooida, 1985; Ishida, 1992; Noguchi, 1996; Miyoshi and Ishibashi, 2004), as observed for the subducting Pacific slab. This is roughly supported by the following observations. ScSp waves were also detected and were used to estimate the upper surface of the Philippine Sea Plate subducting beneath southwestern Japan (Nakanishi, 1980; Nakanishi et al., 1981; Iidaka et al., 1990). Fukao et al. (1983), and later Hori et al. (1985), Hori (1990) and Ohkura (2000), detected seismic guided waves propagating within the oceanic crust of the Philippine Sea slab, which indicates that the oceanic crust remains untransformed to depths of at least ~60 km. These studies suggest that intraslab earthquakes in the Philippine Sea slab occur in the upper portion near the upper surface of the slab.

However, the configuration of the subducting Philippine Sea Plate beneath southwestern Japan remains poorly understood, primarily because only limited seismic activity is associated with subduction of this plate. Interplate earthquakes on the upper surface of the subducting plate are almost completely absent, meaning that reliable determination of the interface between the Philippine Sea Plate and the overlying continental plate is difficult. Moreover, simply regarding the upper envelope of intraslab seismicity as the upper plate interface is not a reliable approach, as indicated by seismic wide-angle reflection and refraction surveys undertaken along lines crossing the arc in the Tokai, Kinki, and Shikoku areas. These surveys suggested that the plate interface is several kilometers shallower than the upper limit of intraslab seismicity (Kodaira et al., 2000, 2002, 2004; Kurashimo et al., 2002; Ito et al., 2005).

Hirose et al. (2008a) used a double-difference tomography method (Zhang and Thurber, 2003) with arrival time data for earthquakes recorded by the Kiban seismic network in Japan to detect a layer with low S-wave velocity and high Vp/Vs, and located immediately above a region of intraslab seismicity in areas from Tokai to Kyushu in southwestern Japan. When this layer is compared with the upper interface of the Philippine Sea Plate estimated from the seismic reflection and refraction surveys, it becomes clear that this low-Vs and high-Vp/Vs layer represents oceanic crust of the Philippine Sea Plate (Fig. 3). Therefore, this layer can be used to reliably estimate the distribution of the upper interface of the subducting Philippine Sea Plate at depth; indeed, Hirose et al. (2008a) used this layer to determine the location of the upper interface of the Philippine Sea Plate at depths of 20–60 km across the entire area of southwestern Japan from Tokai to Kyushu, as shown by isodepth contours in Fig. 2. Here, the shallower portion of the subducting plate beneath the Pacific Ocean (isodepth contour of 10 km) was determined by Baba et al. (2002), who estimated the position of the upper interface of the Philippine Sea Plate using offshore seismic reflection and refraction surveys. In Fig. 2, the deeper portions (60–200 km) of the plate beneath Chubu and Chugoku-Kyushu represent the estimates of Nakajima and Hasegawa (2007), who determined the upper interface in the same area as the seismic tomography-derived upper envelope of the high-seismic-velocity slab.

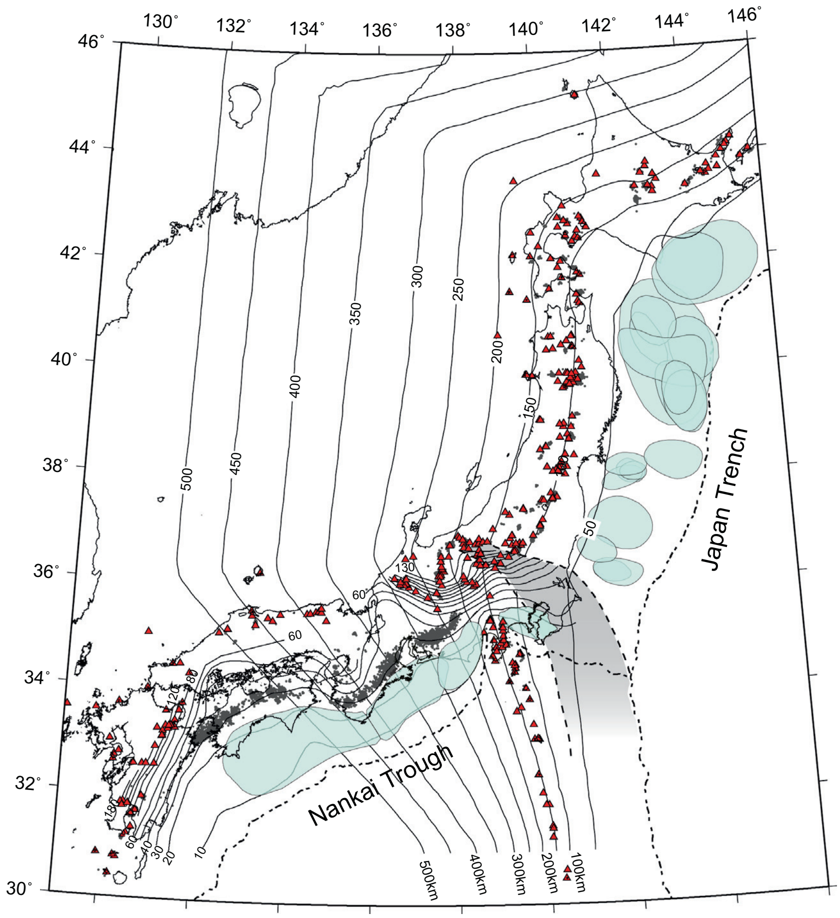

Fig. 2. Map showing isodepth contours of the upper surfaces of the Pacific and Philippine Sea Plates (Baba et al., 2002; Nakajima and Hasegawa, 2007; Hirose et al., 2008a,b; Nakajima et al., 2009a,b; Kita et al., 2010). The contact zone between the Philippine Sea and overlying Pacific Plates is shaded in gray and enclosed by two broken curves (Nakajima et al., 2009a; Uchida et al., 2009). Source area of the 1923 Kanto earthquake (Wald and Somerville, 1995) and those of predicted Tokai, Tonankai and Nankai earthquakes on the upper interface of the Philippine Sea Plate (HERP, MEXT, http://www.jishin.go.jp/main/index.html) are shown as light blue ellipses. Source areas of $M > 7$ interplate earthquakes on the upper interface of the Pacific Plate in the past 80 years (Umino et al., 1990; Uchida et al., 2009) are also shown as light blue ellipses. Red triangles denote Quaternary volcanoes. Deep low-frequency tremors/earthquakes are shown as dots.

A number of differing models for the configuration of the subducting Philippine Sea Plate at depth below Kanto have been proposed, primarily using hypocenter distributions, focal mechanisms, later phase analyses, three-dimensional seismic velocity structures, and seismic reflection and refraction surveys (e.g., Iidaka et al., 1990; Ishida, 1992; Sekiguchi, 2001; Kodaira et al., 2004; Sato et al., 2005; Matsubara et al., 2005; Kimura et al., 2006; Hori, 2006; Wu et al., 2007; Nakajima and Hasegawa, 2007). Recently, Hirose et al. (2008b) used double-difference tomography with earthquake arrival time data for the Kanto area obtained by the Kiban seismic network to clearly delineate a several-kilometer-thick low-Vs and high Vp/Vs layer that is shallowly inclined along the subduction direction of the Philippine Sea slab. Comparison of the location of this low-Vs and high Vp/Vs layer with the delineated upper surface of the Philippine Sea slab, as estimated from seismic refraction surveys, hypocenter distributions, and focal mechanisms, indicates that this layer is most probably oceanic crust within the subducting slab, as previously observed in southwestern Japan. Based on this observation, Hirose et al. (2008b) delineated the upper surface of the subducting Philippine Sea Plate beneath Kanto at depths down to ~90 km; this delineation is shown as 10–90 km isodepth contours in the Kanto area of Fig. 2.

The region north of the Izu Peninsula between Kanto and Tokai is characterized by a large fan-shaped seismicity gap in intraslab events (e.g., Ishida, 1992) and by a sparse distribution of Quaternary volcanoes (see Fig. 2), with some researchers proposing that the Philippine Sea slab tears in this area, splitting into the eastern Kanto slab and the western Tokai slab (e.g., Ishida, 1992; Mazzotti et al., 1999). In comparison, other researchers have suggested that an aseismic portion of the Philippine Sea slab exists in this region (e.g., Iidaka et al., 1990; Sekiguchi, 2001; Nakamichi et al., 2007; Matsubara et al., 2008), although the nature of this aseismic portion was not known until recently. A recent seismic tomography study using arrival time data recorded by the Kiban seismic network (Nakajima et al., 2009a,b) revealed that a seismic high-velocity zone extends continuously down to ~140 km depth to the north of the Izu Peninsula (Fig. 4), suggesting that the Philippine Sea Plate is present as a single slab from Kanto to Tokai, with subduction as an aseismic slab. Nakajima et al. (2009a,b) also identified aseismic subduction of the Philippine Sea slab down to ~140 km depth beneath Kanto, connecting this slab to the north of the Izu Peninsula with the slab geometry in Kanto and Tokai using seismic tomography; these results are shown as 10–130 km isodepth contours in the region between Kanto and Tokai in Fig. 2.

2.3. Slab–slab contact zone beneath Kanto

To the south of the Sagami Trough, the Pacific Plate subducts beneath the Philippine Sea Plate west-northwestward along the

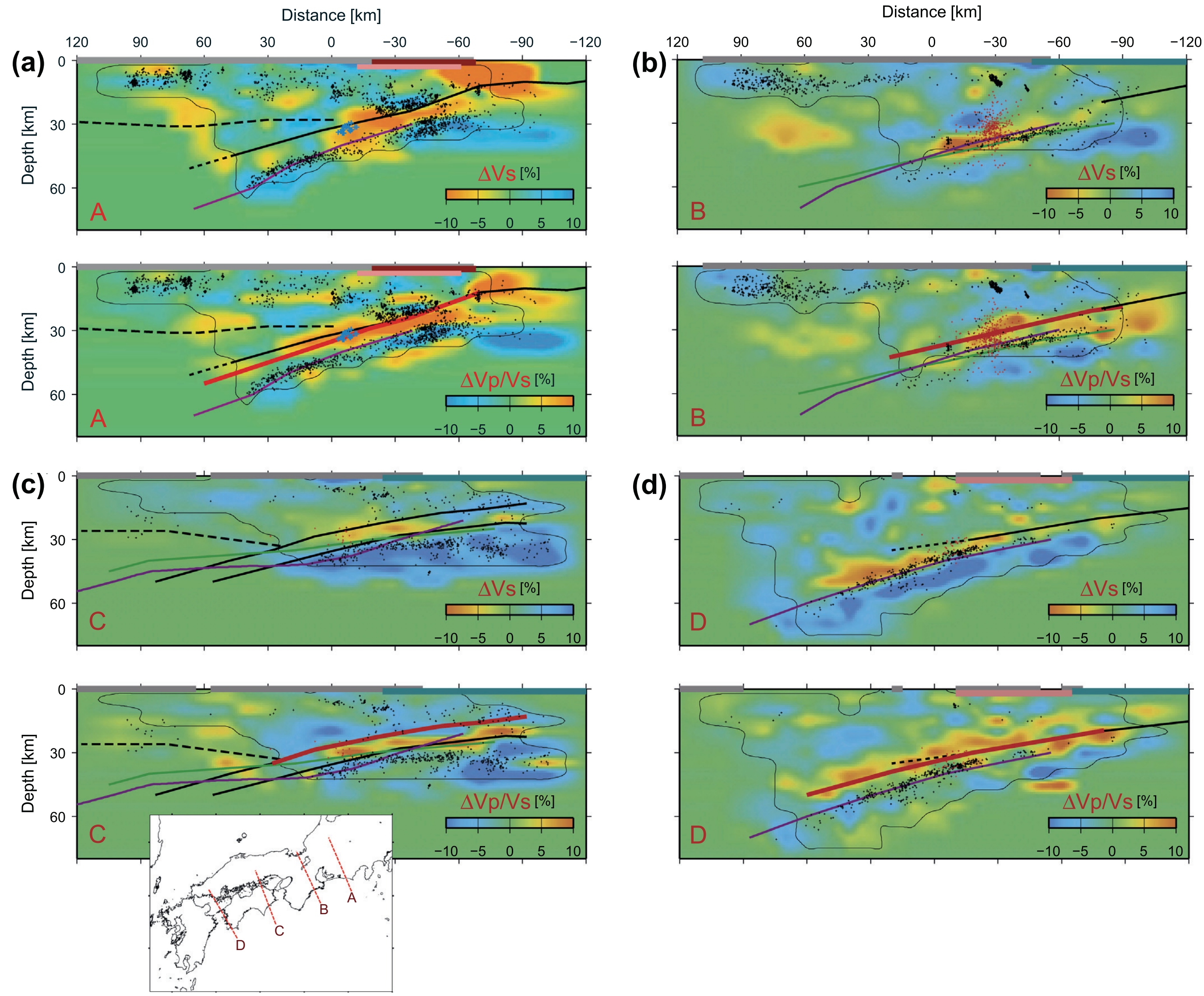

Fig. 3. Across-arc vertical cross-sections showing (upper) S-wave velocity perturbations and (lower) Vp/Vs ratio perturbations along profiles: (a) A, (b) B, (c) C, and (d) D, with locations shown in the inset map (Hirose et al., 2008b). S-wave velocity and Vp/Vs ratio perturbations are shown as color scales at the bottom right of each figure. Thin black lines represent a DWS (derivative-weighted summation) of 3000 (Thurber and Eberhart-Phillips, 1999). Black and blue crosses denote relocated earthquakes and relocated deep, nonvolcanic, low-frequency earthquakes, respectively. Red crosses are nonvolcanic, low-frequency earthquakes located by the JMA. Red lines in the lower figures indicate the estimated location of the upper plate interface, and solid black lines denote the plate interface estimated by seismic reflection and refraction surveys (Kodaira et al., 2002, 2004; Nakanishi et al., 2002; Takahashi et al., 2002). Purple and green lines denote the slab Moho imaged by receiver function analyses (purple lines: Shiomi et al., 2006; green lines: Yamauchi et al., 2003), and dashed lines indicate the island arc Moho (Kodaira et al., 2004).

Izu–Bonin trench, the two plates being in contact with each other at shallow depths. To the north of the Sagami Trough, the Philippine Sea Plate in turn subducts beneath the North American Plate. After the subduction of the Philippine Sea Plate beneath Kanto, this contact zone is expected to deepen and move northwestward following the westward dip of the directly underlying Pacific Plate. Recent investigations using data obtained by the dense Kiban seismic network indicate that this slab–slab contact zone between the base of the Philippine Sea slab and the top of the Pacific slab underlies a wide area beneath the Kanto Plain (Nakajima et al., 2009a,b; Uchida et al., 2009).

The northeastern limit (updip limit) of the slab contact zone corresponds to the boundary between the overlying Philippine Sea and North American Plates, and can be estimated using the spatial distribution of interplate earthquake slip vectors (Seno and Takano, 1989; Noguchi, 2007). Recently, Uchida et al. (2009) investigated interplate earthquake focal mechanism solutions along the interface between the Pacific and Philippine Sea Plates and the interface between the Pacific and North American Plates. They identified the northeastern limit by using a large number of focal mechanism solutions for interplate events from the catalogue of the F-net moment tensor solution database (Broadband Seismic Network Laboratory, NIED, 2008) with a higher resolution than the previous estimations.

The southwestern edge, or downdip limit, of the slab contact zone can be determined if the precise location and configuration of the upper surface of the Pacific slab and the lower surface of the Philippine Sea slab are known. Nakajima et al. (2009a,b) accurately estimated the location of the upper surface of the subducting Pacific slab using relocated hypocenter distributions, low-angle thrust-type focal mechanisms, and a high-velocity anomaly imaged by seismic tomography. The location of the bottom of the Philippine Sea slab was estimated using the thickness of the high-velocity anomaly and the location of intraslab events.

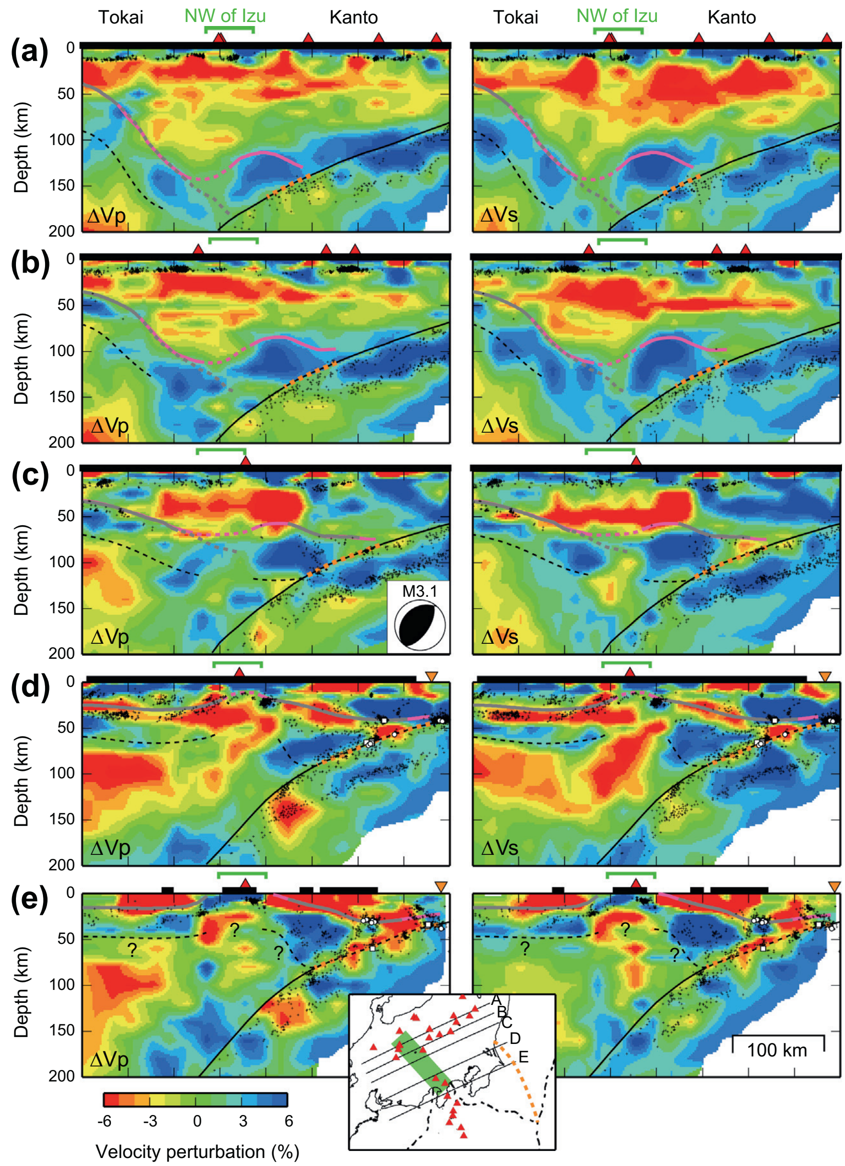

Fig. 4. Along-arc vertical cross-sections of P (left) and S (right) wave velocity perturbations along profiles: (a) A, (b) B, (c) C, (d) D, and (e) E, as located in the inset map. Velocity perturbations are shown by the color scale at the bottom of the figure, pink lines indicate the estimated upper surface of the Philippine Sea slab, and dashed pink lines denote where the slab surface is not constrained by a high-velocity anomaly. Black dashed lines show the approximate location of the bottom of the Philippine Sea slab, and gray solid and dashed lines denote the upper surface of the Philippine Sea slab, as estimated by Hirose et al. (2007, 2008a,b) and Nakajima and Hasegawa (2007), respectively. The estimated upper surface of the Pacific slab is delineated by black lines, and orange dashed lines indicate the contact zone between the Philippine Sea and Pacific slabs. Crosses denote earthquakes relocated with 3D seismic velocity structures, and white circles and squares denote interplate and small repeating earthquakes along the upper interfaces of the Pacific and Philippine Sea slabs (Uchida et al., 2009), respectively.

The configuration of the Philippine Sea slab in the contact zone thus estimated has a wedge shape, and its thickness reaches ∼60 km at the maximum beneath Kanto, which is about twice the estimated thickness in the area from Tokai to Shikoku in southwestern Japan (e.g., Ishida, 1992). This estimate of the slab thickness is consistent with the age of the Philippine Sea slab, which is >48 Ma in the Kanto area and 15–28 Ma in the area from Tokai to Shikoku (Seno and Maruyama, 1984). Moreover, Uchida et al. (2010) confirmed this wedge-shaped configuration of the Philippine Sea slab in the contact zone based on analyses of converted waves at the plate interfaces and small repeating earthquakes.

The slab contact zone thus estimated is shown as two broken curves in Fig. 2, indicating that the contact zone is present across a wide area beneath Kanto, with a correlation between the lateral extent of the slab–slab contact zone and the location of the Kanto Plain, the largest plain in the Japanese Islands. This observation suggests that the slab-slab contact plays a part in the subsidence of the Kanto Plain, although further discussion on a geological timescale is required to relate the slab-slab contact zone to the formation of the Kanto Plain (Nakajima et al., 2009a,b). Isodepth contours delineating the upper surface of the Philippine Sea slab vary significantly in the north–south direction in a broad area extending from Kanto to Shikoku (Fig. 2), including the area north of the Izu collision zone. This undulatory configuration of the subducting Philippine Sea slab may be related to along-arc contraction and buckling deformation caused by subduction of this slab into the mantle, where the downgoing slab becomes spatially constrained with increasing depth, primarily due to the westward dip of the immediately underlying Pacific Plate.

The location of this broad slab–slab contact zone at depths of 0–140 km beneath Kanto hinders heating of both the Pacific and Philippine Sea slabs by the hot mantle wedge, causing an

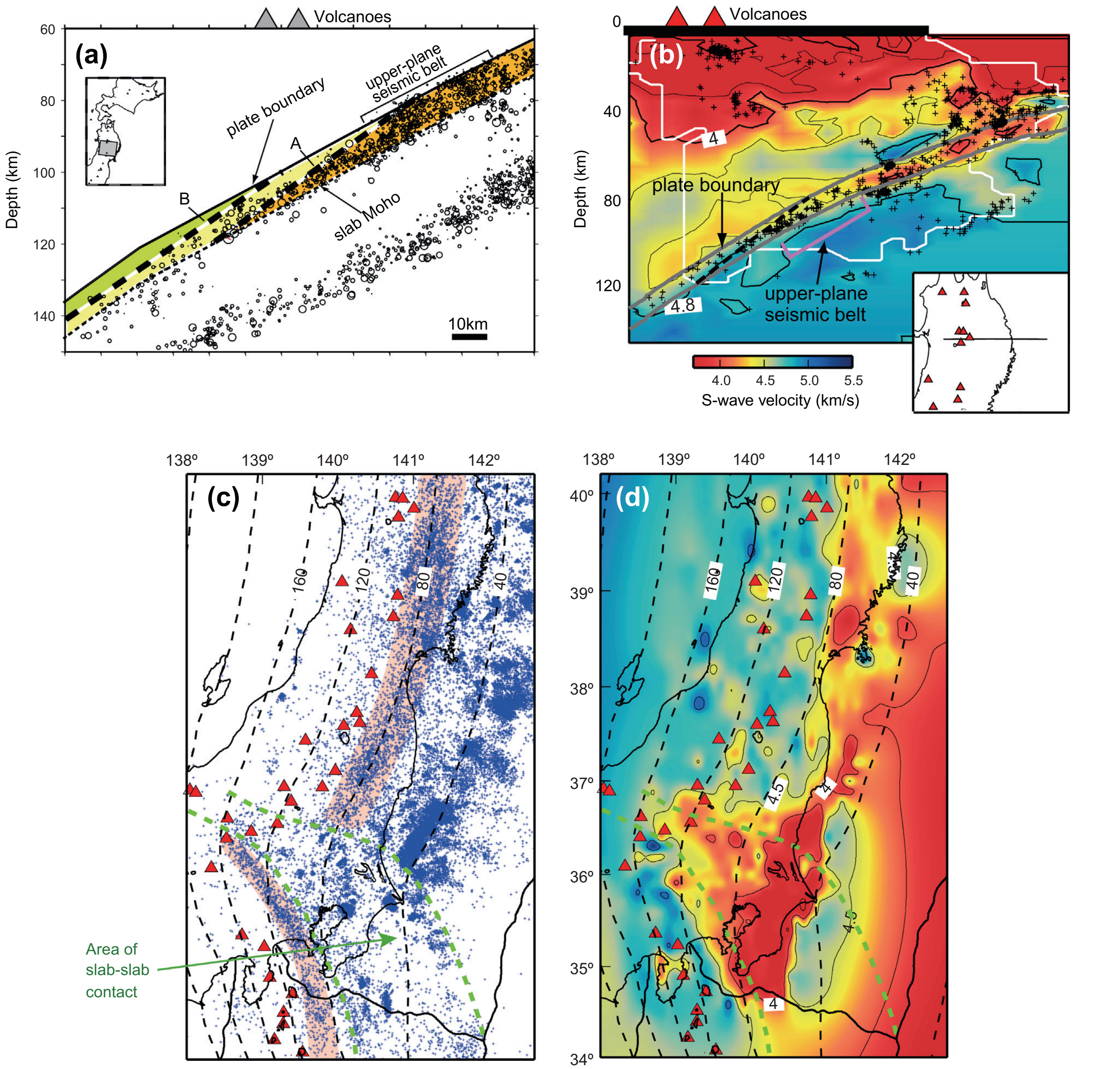

Fig. 5. (a) Across-arc vertical cross-section showing the location of intraslab earthquakes (open circles) in central Tohoku (Kita et al., 2006), with A denoting the location of the upper-plane seismic belt, and B and C indicating facies boundaries. (b) Across-arc vertical cross-section showing variations in S-wave velocity (color scale at bottom of figure) in central Tohoku (Tsuji et al., 2008), with the upper interface of the Pacific Plate shown as a solid line, and A denoting the upper-plane seismic belt. The B facies boundary shown in part (a) is here shown as a broken line. (c) Distribution of earthquake epicenters within the crust of the Pacific slab; earthquakes at 0–10 km below the upper plate surface are shown as blue dots, and the upper plane seismic belt is pink. The contact zone between the Philippine Sea and Pacific slabs beneath Kanto is delineated by two broken green curves, with solid curves and red triangles showing isodepth contours of the upper surface of the Pacific slab and the location of active volcanoes, respectively. (d) S-wave velocity distribution in the crust of the Pacific slab on a curved plane 5 km below the upper plate surface (Nakajima et al., 2009b); broken black and green curves denote isodepth contours of the upper plate surface and the slab–slab contact zone, respectively.

anomalously deep downdip limit of interplate, intraslab, and shallow inland earthquake activity beneath Kanto. The downdip limit of interplate events estimated from the hypocenter distribution of the low-angle thrust-type earthquakes and from the distribution of fault planes of large interplate events is subparallel to an isodepth contour of ∼50 km for the Pacific Plate and that of ∼25 km for the Philippine Sea Plate. This is also seen from the distribution of the source areas of large interplate earthquakes shown by ellipses with light blue color in Fig. 2. The fact that this downdip limit is subparallel to the isodepth contours of the upper surface of the subducting slab indicates that this limit is temperature controlled. However, the downdip limit of the interplate events locally deepens beneath Kanto, with interplate events along the upper surface of the subducting Philippine Sea slab occurring down to ∼55 km depth immediately above the slab–slab contact zone, whereas interplate earthquakes along the upper surface of the Pacific slab occur down to ∼80 km depth within the slab–slab contact zone (Nakajima et al., 2009a,b; Uchida et al., 2009). These depths are much greater than the ∼25 and ∼50 km depths reported for the Philippine Sea and Pacific slabs, respectively, in other regions of the Japanese Islands, indicating that the anomalously low-temperature conditions beneath Kanto are related to the slab–slab contact in this region.

The cutoff depth of shallow inland earthquakes within the continental North American Plate is also deeper beneath Kanto than

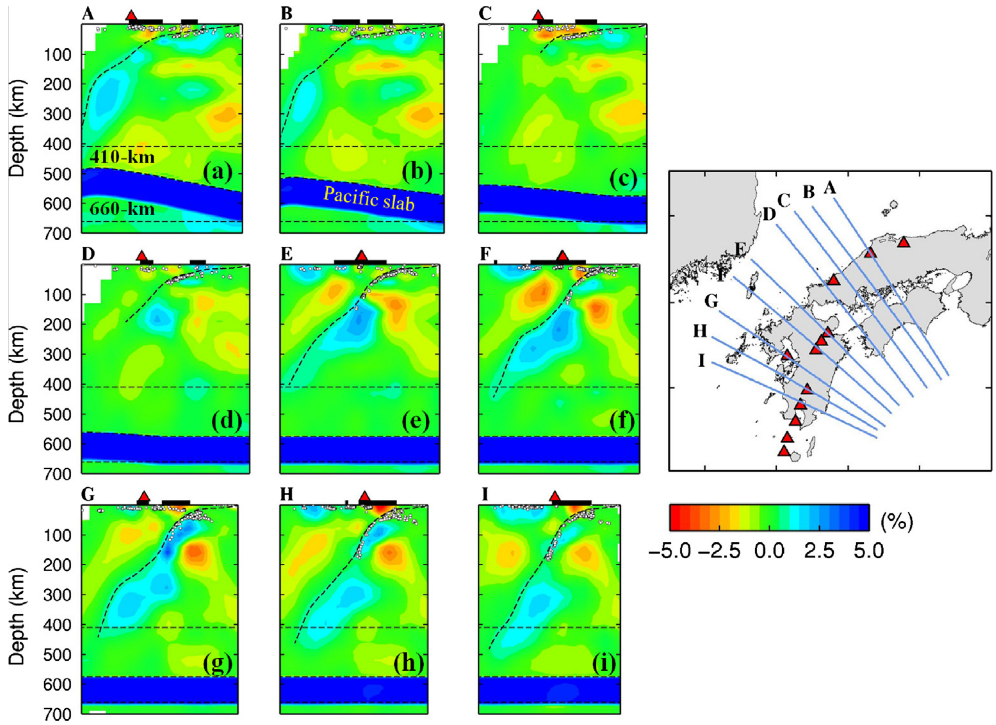

Fig. 6. Across-arc vertical cross-sections showing P-wave velocity perturbations beneath western Japan; cross-sections shown are along profiles A through I in the inset map (Zhao et al., 2012), and velocity perturbations are indicated by the color scale at bottom right. Red triangles and thick horizontal lines above each cross-section denote the location of active volcanoes and the land area, respectively. Two dashed lines in each cross-section represent 410 and 660 km discontinuities, and white dots indicate earthquakes that occurred within a 20 km width along strike of each profile. The estimated upper interface of the Philippine Sea slab is shown as a dashed line in each cross-section. Nearly horizontal blue zones indicate the Pacific slab stagnated in the mantle transition zone.

observed elsewhere, with cutoff depths of up to 20–30 km immediately above the slab contact zone beneath Kanto, contrasting with depths of 10–15 km in other inland areas of the Japanese Islands (e.g., Omuralieva et al., 2012). Heat flow is also low immediately above the slab contact zone beneath Kanto (Furukawa, 1993; Tanaka et al., 2004), supporting an interpretation where a local low geothermal gradient is related to the proximity to the slab–slab contact, with subduction of the two cold oceanic slabs creating a dual cooling effect, in addition to the broad contact zone hindering heating of the two oceanic slabs by hot mantle wedge material.

The distribution of intermediate-depth intraslab earthquakes and low-seismic-velocity oceanic crust correlate well with the slab–slab contact zone beneath Kanto (Hasegawa et al., 2007; Nakajima et al., 2009a,b). Kita et al. (2006) detected the existence of a pronounced seismic belt in the upper plane of the double seismic zone, nearly parallel to the ~80 km isodepth contour of the upper surface of the Pacific slab beneath Hokkaido and Tohoku (Fig. 5a). This pronounced seismic belt, hereby termed an ‘upper-plane seismic belt’, is located within the oceanic crust at a depth near a metamorphic facies boundary in the crustal material (Fig. 5a), suggesting that this belt was associated with intraslab earthquakes generated by dehydration-related embrittlement (Kita et al., 2006). Seismic tomography imaging indicates that low-seismic-velocity slab crust persists down to the depth of this upper-plane seismic belt, but not below (Tsuji et al., 2008), suggesting that a phase transformation and eclogite formation occur in the slab crust at this depth (Fig. 5b).

The depth at which this phase transformation takes place is thought to be dependent on the temperature within the slab (e.g., Hacker et al., 2003a,b); if this is the case, the local low-temperature conditions in the Pacific slab immediately beneath the slab contact zone beneath Kanto should cause a delay in phase transformation. As expected, both the upper-plane seismic belt and the low-seismic-velocity oceanic crust deepen beneath the slab contact zone in the Kanto area. Hasegawa et al. (2007) showed that the upper-plane seismic belt beneath the Kanto area is oblique to the ~80 km isodepth contour, deepening toward the north from a depth of ~100 km to 140 km along a trend nearly parallel to the downdip edge of the slab contact zone (Fig. 5c). A seismic tomography study by Nakajima et al. (2009a,b) indicated that the low-seismic-velocity region in the slab crust extends to the uppermost depths where the obliquely trending upper-plane seismic belt is observed (Fig. 5d). These observations strongly support an interpretation where low-temperature conditions associated with the slab–slab contact zone beneath the Kanto area cause a delayed onset of eclogite-forming phase transformations. These observations also indicate that the estimated location and areal extent of the slab–slab contact zone, shown as broken lines in Fig. 5c and d, is accurately located.

2.4. Deep aseismic subduction of the Philippine Sea Plate

Recent seismic tomography studies using both local earthquakes and teleseismic events have enabled the high-resolution determination of the structure of crust and mantle down to a depth of 700 km beneath the Japanese Islands. Zhao et al. (2012) combined a large amount of high-quality local and teleseismic data recorded by the dense Japan-wide Kiban seismic network to locate deep seismic structures, resulting in the identification of aseismic subduction of the Philippine Sea slab at depth under southwestern Japan. The deeper portion of the subducting Philippine Sea slab is clearly imaged as a zone with high P-wave velocity down to a depth of ~370 km under the Japan Sea to the north of Chugoku, and to ~430 km depth under western Kyushu (Fig. 6), although intraslab seismicity ceases at depths of 40–60 km beneath Chugoku and 140–180 km under Kyushu.

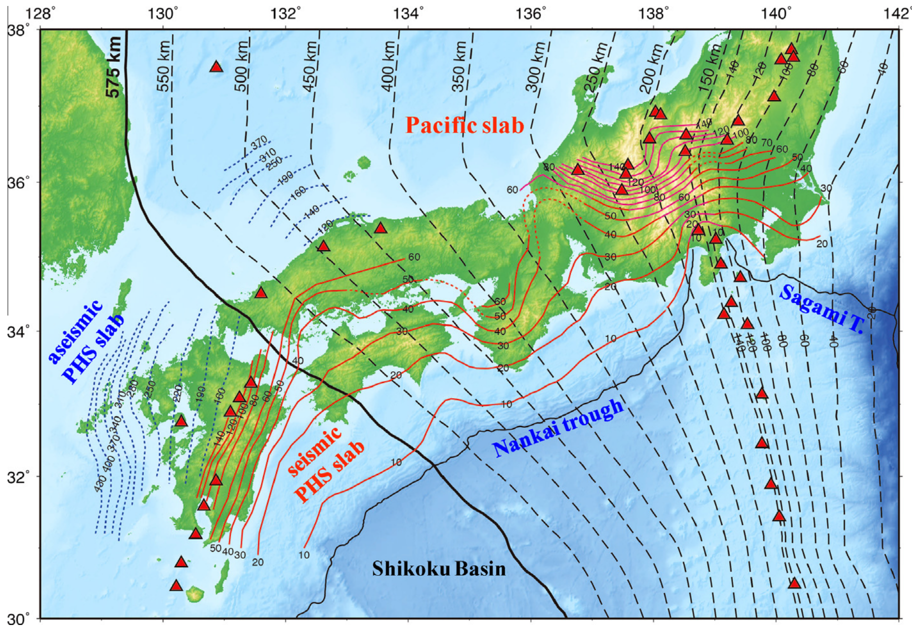

Fig. 7. Map showing isodepth contours of the upper interface of the Pacific (black dashed lines) and Philippine Sea (red and blue dashed lines; see below) slabs (Zhao et al., 2012). The Pacific slab stagnates in the mantle transition zone to the west of the 575-km-depth contour, shown as a thick black line. Red isodepth contours delineate shallow portions of the Philippine Sea slab estimated from earthquake hypocenter distributions and local-earthquake tomography (Fig. 2), and blue dotted lines in the Japan Sea and western Kyushu area delineate the upper interface of the aseismic Philippine Sea slab, as estimated from teleseismic tomography.

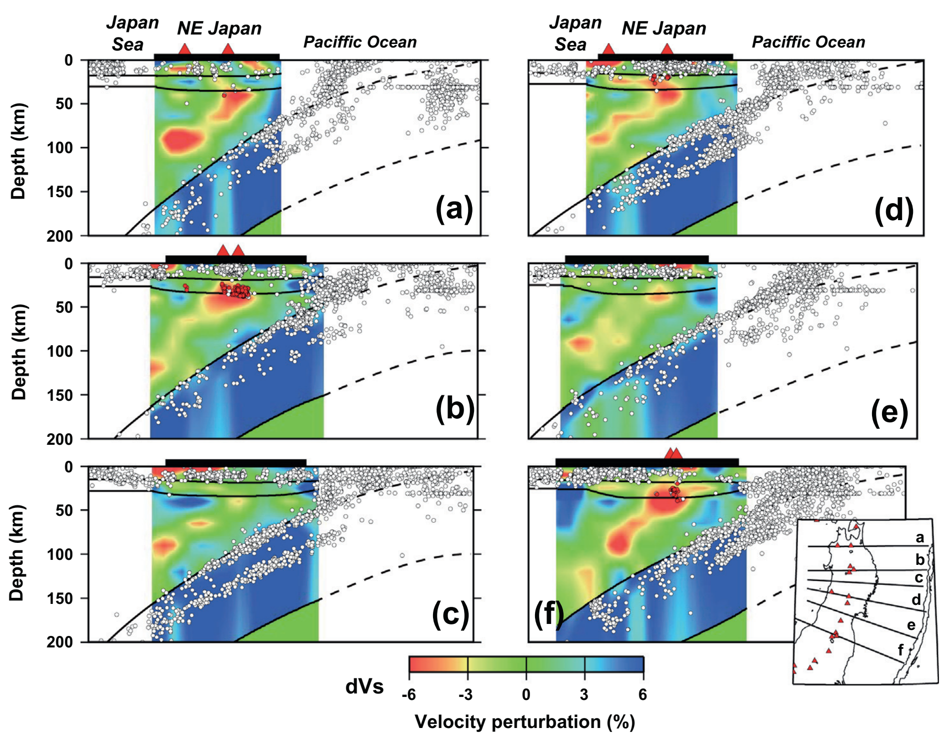

Fig. 8. Across-arc vertical cross-sections showing S-wave velocity perturbations in Tohoku; cross-sections are along profiles: (a) a, (b) b, (c) c, (d) d, (e) e, and (f) f shown in the inset map (Nakajima et al., 2001). S-wave velocity perturbations are indicated by the color scale at the bottom of the figure, with red triangles and the solid black bar at the top of each figure denoting the location of active volcanoes and the land area, respectively. Open circles indicate earthquake locations, with red circles denoting deep low-frequency earthquakes, and solid lines denoting the Conrad and Moho discontinuities within the upper plate and upper and lower surfaces of the Pacific slab.

The configuration of the upper surface of the aseismic portion of the subducting Philippine Sea slab, as estimated from the location of tomographically imaged high-velocity zones, is shown in Fig. 7. The deep aseismic portion of the Philippine Sea slab appears to be discontinuous and is visible under the Japan Sea to the north of Chugoku and under the East China Sea to the west of Kyushu. This aseismic portion of the slab is not detectable between these two areas, suggesting that the aseismic portion of this slab may split into two portions at depth. Another possibility is that the tomographic imaging did not capture all of the aseismic portion of the

Philippine Sea slab, possibly due to insufficient ray-path coverage relating to a lack of seismic stations immediately above the slab window. This problem was addressed in a recent seismic tomography study by Huang et al. (submitted for publication), which combined local and teleseismic data recorded by dense seismic networks in Japan and South Korea, with a much improved ray-path coverage provided by data obtained from stations in South Korea. The tomographic images obtained during this study confirm the existence of this slab window, supporting the findings of Zhao et al. (2012), although the processes that formed this window are not clear.

3. Arc magmatism

3.1. Magma genesis in the northeastern Japan arc

Seismic tomography studies in the northeastern Japan arc have detected inclined low-velocity and high-attenuation zones oriented sub-parallel to the subducting Pacific Plate at depths of 30–150 km within the mantle wedge (Zhao et al., 1992, 1994; Tsumura et al., 2000; Nakajima et al., 2001; Hasegawa and Nakajima, 2004; Hasegawa et al., 2009). Fig. 8 shows across-arc vertical S-wave velocity cross-sections (Nakajima et al. (2001)) that demonstrate the continuation of this inclined low-velocity zone underneath the entire Tohoku area. This zone extends from a depth of 100 and 130 km under the Japan Sea coast to the Moho immediately under the volcanic front in the middle of the arc that runs nearly parallel to the trench axis. Results of both the checkerboard resolution test and the restoring resolution test showed that major features such as the high-velocity subducted slab and this inclined low-velocity zone in the mantle wedge could be well resolved with a resolution scale of 25 km in the horizontal direction and 10–30 km in the depth direction (see Fig. 4 of Nakajima et al. (2001)).

This inclined, sheet-like, low-velocity, high-attenuation zone is thought to represent the upwelling portion of a secondary convection cell that was mechanically induced by slab subduction (McKenzie, 1969). A hydrated layer directly above the slab has been imaged as a thin seismic low-velocity layer at depths of ~70–120 km by seismic receiver function analysis (Kawakatsu and Watada, 2007) and by double-difference tomography (Tsuji et al., 2008), and probably corresponds to the downflow portion of this secondary convection system (e.g., Iwamori, 1998). The temperature of this inclined low-velocity zone was estimated by comparing seismic attenuation structures and experimental data, and is thought to be higher than the wet solidus of peridotite, suggesting partial melting occur within this upwelling flow (Nakajima and Hasegawa, 2003). Actually, Nakajima et al. (2005) showed that melt-filled pores with aspect ratios of 0.10–0.01 and with melt fraction volumes of 0.1 to several percent exist within this upwelling flow, by comparing the observed fall-off rates of Vp and Vs with a diagram of Takei (2002) which shows the relative role of liquid compressibility and pore geometry in determination of Vp/Vs ratio.

H_2O transportation paths in the northeastern Japan arc can be inferred from seismic observations (Hasegawa and Nakajima, 2004; Hasegawa et al., 2005; Fig. 9a), with H_2O liberated from the hydrated slab crust migrating upward into the overlying mantle wedge, where it reacts with mantle materials to form a layer containing hydrated minerals such as serpentine, chlorite, and amphibole (e.g., Davies and Stevenson, 1992; Iwamori, 1998). This hydrated layer lies immediately above the subducting slab and is thought to be dragged downward by the slab, moving to depths where further dehydration-related decomposition may occur (Iwamori, 1998; Schmidt and Poli, 1998; Maruyama and Okamoto, 2007). The H_2O thus liberated from the hydrated layer at depth migrates upward to meet the immediately overlying upwelling flow zone. This addition of H_2O to hot mantle material within the upwelling flow causes partial melting, producing melt with volume fractions of 0.1 to several percent. Melt is formed both by decompression melting and by melting due to the addition of water. The slab-derived water is thus eventually incorporated into melt within the upwelling flow, with melts accumulating immediately below the Moho along the volcanic front. These melts then rise, penetrating the arc crust and ultimately reaching the surface to form volcanoes, with the volcanic front thought to form at locations where inclined sheet-like upwelling flows reach the arc Moho (Fig. 9a).

Quaternary volcanoes in the northeastern Japan Arc are grouped into a number of clusters in a long band oriented nearly perpendicular to the arc (Kondo et al., 1998; Tamura et al., 2002; Fig. 10). Tamura et al. (2002) determined that this volcanic clustering is spatially correlated with back-arc topographic highs, low-velocity zones beneath the Moho along the volcanic front, and negative Bouguer gravity anomalies along the Japan Sea coast, suggesting that clustering is caused by the development of localized inclined fingers (~50 km wide) of hot mantle wedge material. Repeated supply of magma from these hot fingers into the overlying arc crust causes uplift and volcano formation.

More distinct low-velocity zones in the mantle wedge that are spatially correlated with volcanic clusters were imaged by a seismic tomographic study of Hasegawa and Nakajima (2004; Fig. 10a). Distinct along-arc variations are evident in S-wave velocity images along the inclined sheet-like low-velocity zone in the mantle wedge, with areas of particularly low seismic velocity

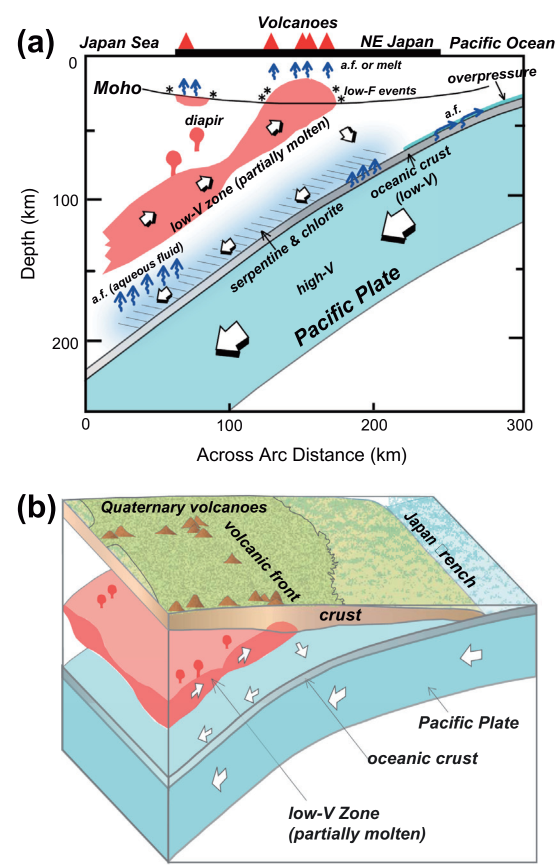

Fig. 9. Schematic figure showing an across-arc vertical cross-section of the crust and upper mantle structure of Tohoku, illustrating the inferred pathway of H_2O transportation (a) and 3D expression of the crust and upper mantle structure showing upwelling flow with varying thicknesses in the mantle wedge (b; Hasegawa and Nakajima, 2004).

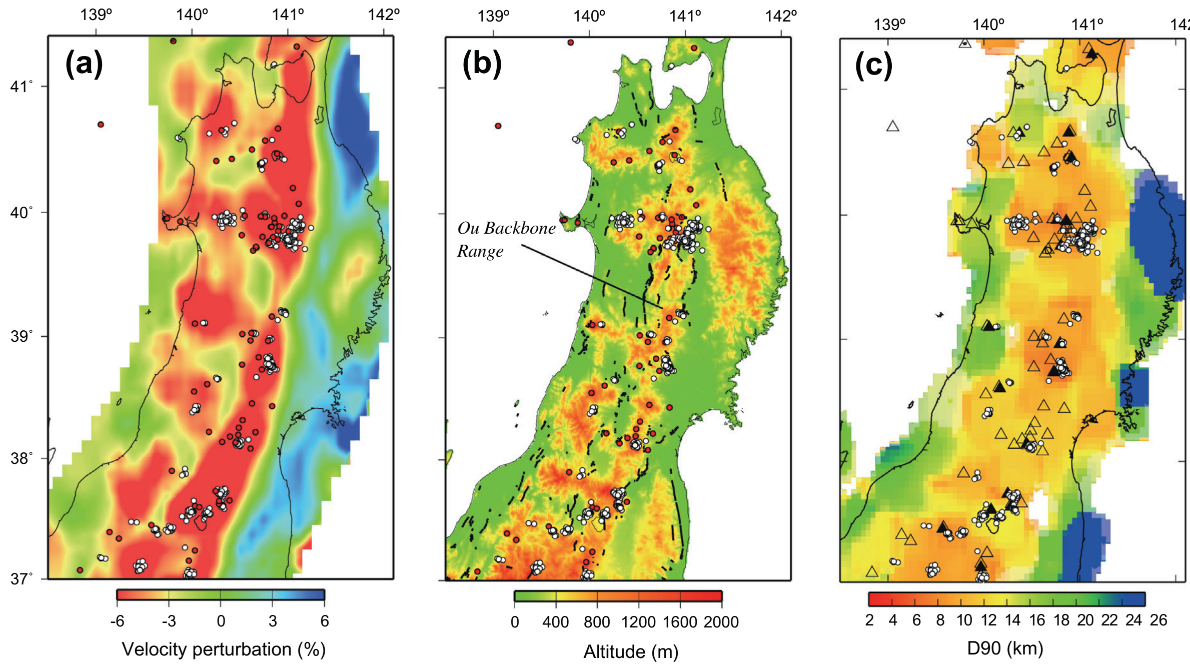

Fig. 10. (a) S-wave velocity perturbations along the inclined low-velocity zone within the mantle wedge (Hasegawa and Nakajima, 2004), with (b) topography and (c) shallow seismicity cut-off depth in Tohoku (Omuralieva et al., 2012). S-wave velocity perturbations in (a), altitudes in (b), and cut-off depths of shallow seismicity (D90) in (c) are indicated by color scales at the bottom of each figure. Red circles in (a) and (b), and open and solid triangles in (c) indicate the location of Quaternary volcanoes. Open circles in (a–c) indicate the location of deep low-frequency earthquakes. S-wave velocity perturbations in (a) are those taken along the core of the inclined sheet-like low-velocity zone at depths of 30–150 km within the mantle wedge beneath Tohoku, which was imaged by a high-resolution seismic tomography (Hasegawa and Nakajima, 2004).

occurring periodically every ~80 km along the strike of the arc (Fig. 10a). These very-low-velocity areas are spatially related to topographic highs between the volcanic front and back arc areas that host volcanic clusters (Fig. 10b) and local shallow seismicity-cutoff depths (Fig. 10c), suggesting that the repeated supply of magma from these hot regions in the mantle wedge to the arc crust has led to volcano formation and a local increase in the geothermal gradient. Using these observations, Hasegawa and Nakajima (2004) proposed a structure for the crust, the upper mantle, and the upwelling flow zone in the mantle wedge within the northeastern Japan arc (Fig. 9b), extending the two-dimensional structure shown in Fig. 9a to three dimensions.

3.2. Mantle upwelling flow beneath the Japanese Islands

As described above, the inclined zone of upwelling flow in the mantle wedge, which was mechanically induced by slab subduction, is clearly visible as seismic low-velocity and high-attenuation zones in the Tohoku area. This upwelling flow is considered to exist in common to all the subduction zones. Inclined low-velocity zones nearly parallel to the subducting slab, probably corresponding to the upwelling flow similarly to the case of Tohoku, have been detected in the mantle wedge of several other subduction zones, including those in Alaska (Zhao et al., 1995), eastern Aleutians (Abers, 1994), Kamchatka (Gorbatov et al., 1999), Hokkaido (Wang and Zhao, 2005), Kyushu (Abdelwahed and Zhao, 2007), and Tonga (Zhao et al., 1997a,b), although the low-velocity zones in these subduction zones have not been as clearly imaged as that in Tohoku (Fig. 8). The clear images of the inclined low-velocity zone in the Tohoku area were probably enabled by the high density of the seismic stations within the back arc region. If this is the case, the dense nationwide Kiban seismic network should enable tomographic inversions for all of the Japanese Islands that should clearly image these inclined low-velocity zones in the mantle wedge. Recent seismic tomography studies using data from both local and teleseismic events recorded by this dense nationwide seismic network have clearly shown the presence of inclined low-velocity zones in the mantle wedge beneath the entirety of the Japanese Islands (Yanada et al., 2010; Zhao et al., 2012; Huang et al., submitted for publication).

An example of the clear imaging of this inclined low-velocity zone, probably corresponding to the upwelling flow zone in the mantle wedge, is provided in Fig. 11, showing across-arc vertical cross-sections of P-wave velocity structure for an area between the southwestern portion of the Kuril Arc (Hokkaido) and the northern portion of the Izu–Bonin Arc in eastern Japan. Inclined low-velocity zones are visible in the mantle wedge for all cross-sections between Hokkaido (sections AA′ and BB′) to the Izu arc (sections LL′ and MM′), barring those covering arc–arc junctions in the Hokkaido corner (section DD′) and Kanto (section KK′) areas. As discussed in Section 2.3, two subducting oceanic plates are in direct contact within a broad zone beneath the Kanto area, a process that probably hinders secondary convection in the mantle wedge in this area. In comparison, within the Hokkaido corner, the Kuril forearc sliver is colliding with and subducting beneath the forearc crust of the northeastern Japan arc (e.g., Kimura, 1986; Kita et al., 2010), a differing process that also hinders return flow within secondary convection zones. These two differing processes can explain why inclined low-velocity zones are not visible in cross-sections in these areas (Fig. 11). Without the return flow part of secondary convection, volcanoes cannot form, as indicated by the volcanic gap present within these arc–arc junction areas.

Fig. 11 further shows that inclined low-velocity zones in the Tohoku mantle wedge are present not only beneath land areas but extend to a depth of ~200 km beneath the Japan Sea (sections GG′, HH′ and II′). Imaging of this zone was made possible by improving the

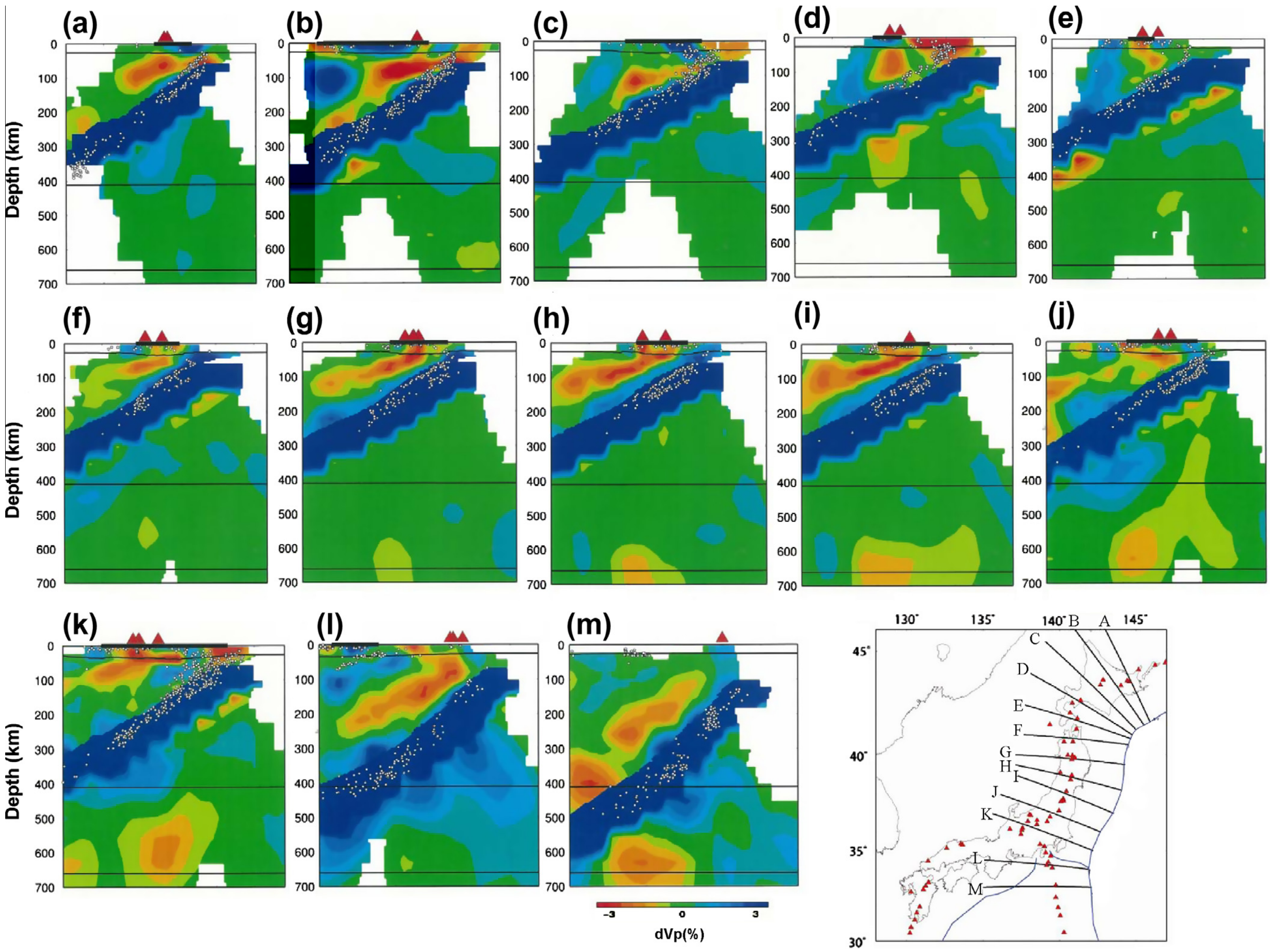

Fig. 11. Across-arc vertical cross-sections showing P-wave velocity perturbations in eastern Japan along profiles shown in the inset map (Yanada et al., 2010), with P-wave velocity perturbations indicated by the color scale at the bottom of the figure. Red triangles and thick horizontal lines at the top of each figure denote active volcanoes and the land area, respectively, and earthquake locations are shown as open circles.

seismic tomography resolution of the area beneath the Japan Sea by adding a number of teleseismic events. In the Izu arc, inclined low-velocity zones also extend to depth, down to ~250–300 km (sections LL′ and MM′), suggesting that these low-velocity zones have deep roots.

Inclined low-velocity zones, which probably represent the upwelling portion of a mechanically induced convection zone, are also distinctly visible in the southwestern Japan subduction zone, associated with subduction of the Philippine Sea slab. As shown in Fig. 6, across-arc vertical P-wave velocity cross-sections beneath Kyushu (sections E to H), obtained by seismic tomography using both local and teleseismic events recorded by the dense Japanese Kiban seismic network, clearly show the presence of inclined low-velocity zones in the mantle wedge at depths of ~30–200 km. In comparison, these low-velocity zones are not visible beneath Chugoku (sections A–D in Fig. 6); here, we suggest that the lack of a visible inclined low-velocity zone is an artifact caused by the lower spatial resolution of seismic tomography in this area, due to poor ray-path coverage.

The presence of inclined low-velocity zones in the mantle wedge across the entirety of Japan, barring arc–arc junctions associated with volcanic gaps and the Chugoku region where seismic tomographic imaging resolution is poor, suggests that upwelling flow in the mantle wedge probably occurs in all subduction zones. In addition, we suggest that volcanoes form at locations where this inclined upwelling flow in the mantle wedge reaches the arc Moho.

4. Conclusions

A recently constructed dense nationwide seismic network in Japan has provided a large volume of high-quality data that has enabled seismic analysis at unprecedented resolution over large areas of the Japanese Islands. Seismic tomography research, precise earthquake hypocenter relocation, and focal mechanism studies using data acquired using this network have shown that the Philippine Sea Plate subducts continuously beneath the area from Kanto to Kyushu in southwestern Japan without splitting, even in areas to the north of the Izu Peninsula. During subduction, this downgoing plate has an undulatory geometry, potentially caused by contractive deformation in an along-arc direction due to restricted space within the mantle with increasing depth. The downdip limit of the plate is ~200 km in the north Chubu and Kyushu areas, whereas the plate reaches only about 60 km beneath a region between north Kinki and north Chugoku. Beneath Kanto, the Philippine Sea slab comes into contact with the underlying subducting Pacific slab within a broad contact zone comparable to the areal extent of the Kanto Plain. This slab–slab contact zone causes a lowering of temperature within the mantle, resulting in anomalously deep downdip interplate, intraslab, and shallow inland earthquake limits beneath Kanto, including the Tokyo Metropolitan Area.

Seismic tomography studies using both local and teleseismic events obtained by the new nationwide seismic network in Japan also enabled the identification of deep aseismic subduction of the Philippine Sea slab beneath the westernmost portion of the southwestern Japan arc. The front of the aseismic subducting Philippine Sea slab is distinctly visible as a zone of high P-wave velocity down to depths of ~370 km beneath the Japan Sea north of Chugoku, and down to depths of ~430 km beneath the East China Sea west of Kyushu, indicating more accurately the geometry of the deeper portion of the Philippine Sea slab in these areas.

Transportation of slab-derived water to the arc crust beneath the northeastern Japan arc has also been estimated using high-quality seismic observations. The ascending part of the mantle wedge secondary convection system, mechanically induced by slab subduction, has been imaged as a series of inclined seismic low-velocity and high-attenuation zones. This mantle upwelling reaches the arc Moho immediately beneath the volcanic front, indicating that volcanism in Tohoku is directly associated with this upwelling flow. This type of ascending portion of a mechanically induced mantle wedge secondary convection system has been identified not only for the Tohoku volcanic front, but also for all active volcanic chains from Hokkaido to Kyushu, suggesting that these Japanese volcano chains are directly related to upwelling flow in the mantlewedge.

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