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GR Focus

Plate subduction, and generation of earthquakes and magmas in Japan as inferred from seismic observations: An overview

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A B S T R A C T

A dense nationwide seismic network recently constructed in Japan has been yielding large volumes of high-quality data that have made it possible to investigate the seismic structure in the Japanese subduction zone with unprecedented resolution. In this article, recent studies on the subduction of the Philippine Sea and Pacific plates beneath the Japanese Islands and the mechanism of earthquake and magma generation associated with plate subduction are reviewed. Seismic tomographic studies have shown that the Philippine Sea plate subducting beneath southwest Japan is continuous throughout the entire region, from Kanto to Kyushu, without disruption or splitting even beneath the Izu Peninsula as suggested in the past. The contact of the Philippine Sea plate with the Pacific plate subducting below has been found to cause anomalously deep interplate and intraslab earthquake activity in Kanto. Detailed waveform inversion studies have revealed that the asperity model is applicable to interplate earthquakes. Analyses of dense seismic and GPS network data have confirmed the existence of episodic slow slip accompanied in many instances by low-frequency tremors/earthquakes on the plate interface, which are inferred to play an important role in stress loading at asperities. High-resolution studies of the spatial variation of intraslab seismicity and the seismic velocity structure of the slab crust strongly support the dehydration embrittlement hypothesis for the generation of intraslab earthquakes. Seismic tomography studies have shown that water released by dehydration of the slab and secondary convection in the mantle wedge, mechanically induced by slab subduction, are responsible for magma generation in the Japanese islands. Water of slab origin is also inferred to be responsible for large anelastic local deformation of the arc crust leading to inland crustal earthquakes that return the arc crust to a state of spatially uniform deformation.

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1. Introduction

Earthquakes occur frequently in the Japan region, including many major earthquakes that have caused extensive damage to inhabited areas of the Japan Islands. The high seismic activity of this region is associated with the convergence of four tectonic plates, (Fig. 1). Northeast Japan lies on the southernmost portion of the North American plate (or the Okhotsk Plate), and southwest Japan lies on the eastern edge of the Eurasia plate (or Amuria plate). These plates converge along the Itoigawa–Shizuoka Tectonic Line in the south, and along the eastern margin of the Japan Sea in the north. In southwest Japan, the Philippine Sea plate subducts northwestward at a rate of 3–5 cm/yr along the Sagami Trough in the east and the Nankai Trough in the west (Seno et al., 1993, 1996; Wei and Seno, 1998; Heki and Miyazaki, 2001; Miyazaki and Heki, 2001), although a new global model for recent plate velocities based on space geodesy shows a slightly higher subduction rate (Sella et al., 2002). In northeast Japan, the Pacific plate subducts beneath the Philippine Sea plate west–northwestward at a rate of 8–9 cm/yr along the Kuril and Japan trenches, and at ca. 6 cm/yr along the Izu–Bonin Trench (DeMets et al., 1994). This complex circumstance of plate convergence causes extremely high seismic activity in Japan and its surrounding areas mainly along the plate boundaries (interplate earthquakes) but also within the subducted plates (intraslab earthquakes) and in the crust of the overriding plates (inland crustal earthquakes). The plate subduction also causes volcanic activity in this country mainly along the volcanic front running the middle of the land area nearly parallel to the trench axis.

A dense nationwide seismic network has been recently constructed in Japan (Okada et al., 2004). This spatially dense seismic network covering the whole country and high seismic activity there provide large volume of high-quality seismic data, rapidly increasing the spatial resolution of data that can be used to estimate the seismic structure of the subduction zone. Many studies on seismic activity, earthquake source processes, and seismic structure have been conducted in this region using data acquired by this dense seismic network. These studies have revealed the detailed structure of the subducting plates and the associated mantle wedges and overlying crust, and have contributed to the collective understanding of the mechanism of earthquake generation in the subduction zones,

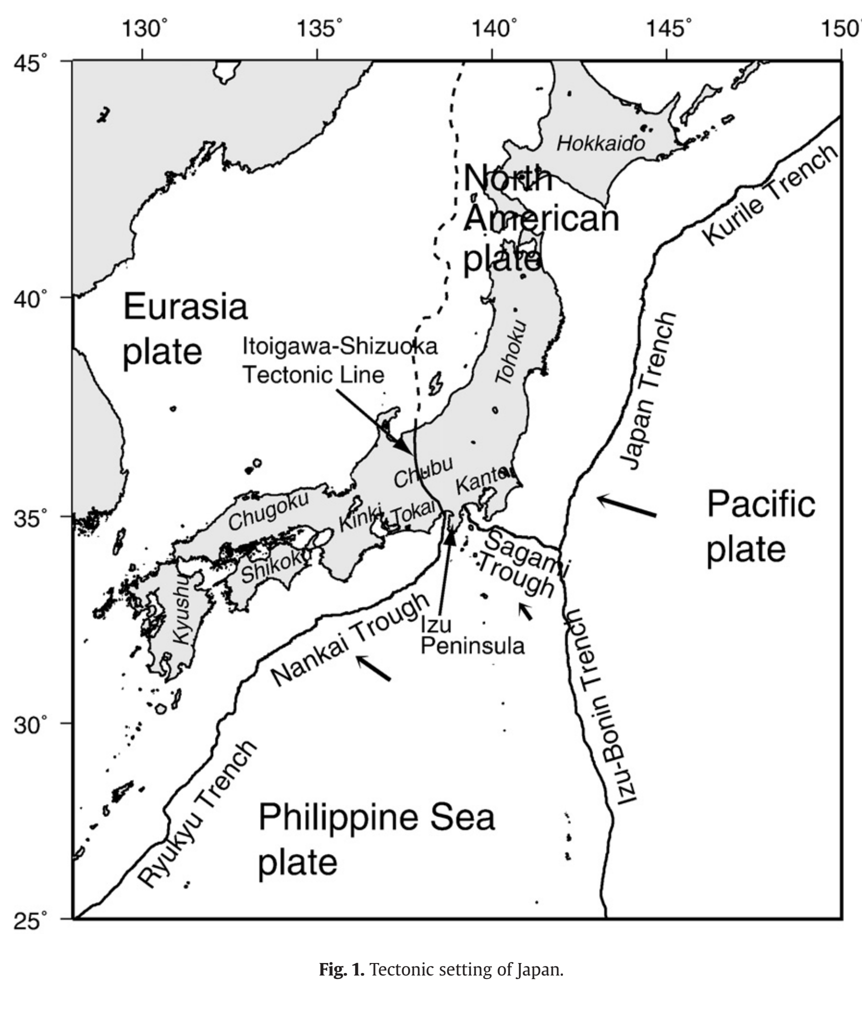

Fig. 1. Tectonic setting of Japan.

especially for its shallower portion down to 150–200 km. In the present article, these studies are summarized, and the inferred generation mechanisms for interplate, intraslab, and inland crustal earthquakes and arc magmatism are discussed.

2. Plate subduction

2.1. Configuration of subducting Pacific and Philippine Sea plates

Fig. 2 shows the epicenter distribution of earthquakes of magnitude 4 or greater (M>4) for the period from 1995 to 2005 in the Japan region. The subduction of the Pacific (PAC) plate along the Japan Trench in the Tohoku region forms a deep planar seismic zone known as the Wadati–Benioff zone, which extends to a depth of approximately 600 km. To the north, the deep seismic zone undergoes a distinct bend at Hokkaido and continues along the Kuril arc. To the south, the deep seismic zone undergoes a distinct bend near the Kanto region at the triple junction of the PAC, Philippine Sea (PHS), and North American (NAM) plates, and continues along the Izu–Bonin arc. The geometry of the deep seismic zone in the Japan region was presented in a review by Utsu (1971). That study also reported evidence that the deep seismic zone is located within a seismic high-Q, high-velocity zone underlying a low-Q mantle, a zone that was inferred to correspond to the subducting PAC slab.

Intermediate-depth earthquakes occurring within the subducting PAC slab in the Hokkaido, Tohoku, and Kanto regions form a double-planed deep seismic zone that lies at depths of 70–180 km (Tsumura, 1973; Umino and Hasegawa, 1975; Hasegawa et al., 1978a; Suzuki et al., 1983). Hasegawa et al. (1978b) first investigated the geometric relationship between this double-planed deep seismic zone and the subducting PAC plate using ScSp waves (ScS-to-P converted waves at the plate boundary) originally reported by Okada (1971). The upper boundary of the subducting plate thus estimated was found to lie immediately above the upper seismic plane of the double seismic zone. This observation was subsequently confirmed by Matsuzawa et al. (1986, 1990), who located the upper surface of the subducting plate based on the detection of converted waves originating from intermediate-depth earthquakes (S-to-P and P-to-S conversion at the plate boundary). It was further shown that events in the upper plane of the double seismic zone occur down to a depth of approximately 150 km within the oceanic slab crust, which has a lower seismic velocity than the mantle wedge above (Matsuzawa et al., 1986). Zhao et al. (1997a,b) later estimated the configuration of the upper surface of the subducting PAC plate in Tohoku by inverting the arrival times of these converted waves. These studies indicated that the upper plate interface is approximately coincident with the upper envelope of intraslab earthquakes at depths deeper than ca. 50 km.

Adopting the upper envelope of intraslab seismicity as the upper plate interface, as indicated by the above observations, Nakajima et al. (2008) and Kita et al. (2008) estimated the configuration of the upper surface of the subducting PAC slab for the entire region of Japan (Fig. 3). In their estimation, the plate interface at shallower depths (<ca. 50 km) was located using the hypocenters of low-angle thrust-type earthquakes and small repeating earthquakes (Uchida et al., 2003) which are considered to be interplate events. At deeper depths (>ca. 150 km), intraslab earthquakes tend to occur in the slab mantle as will be described later (see also Fig. 14), which indicates that their upper envelope does not absolutely show the plate interface but

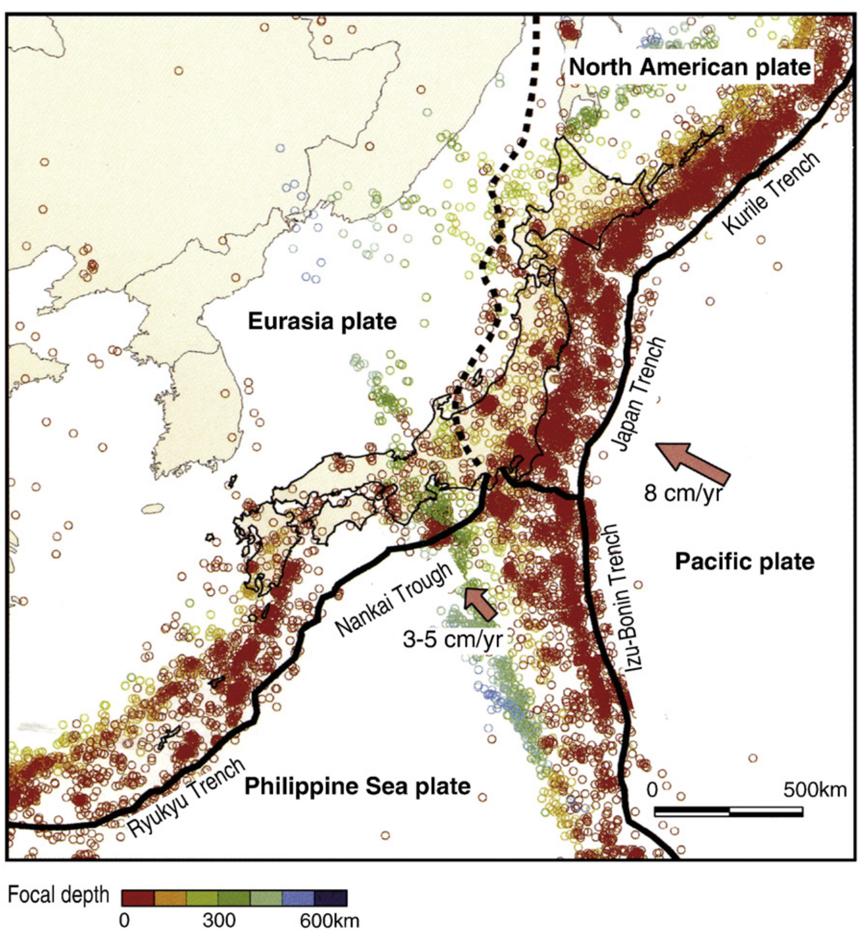

Fig. 2. Epicenter distribution of M >4 earthquakes located by USGS for the period 1995–2005 (Headquarters for Earthquake Research Promotion, Ministry of Education, Culture, Sports, Science and Technology, http:// www.jishin.go.jp/main/index.html). Focal depths are indicated by the color scale.

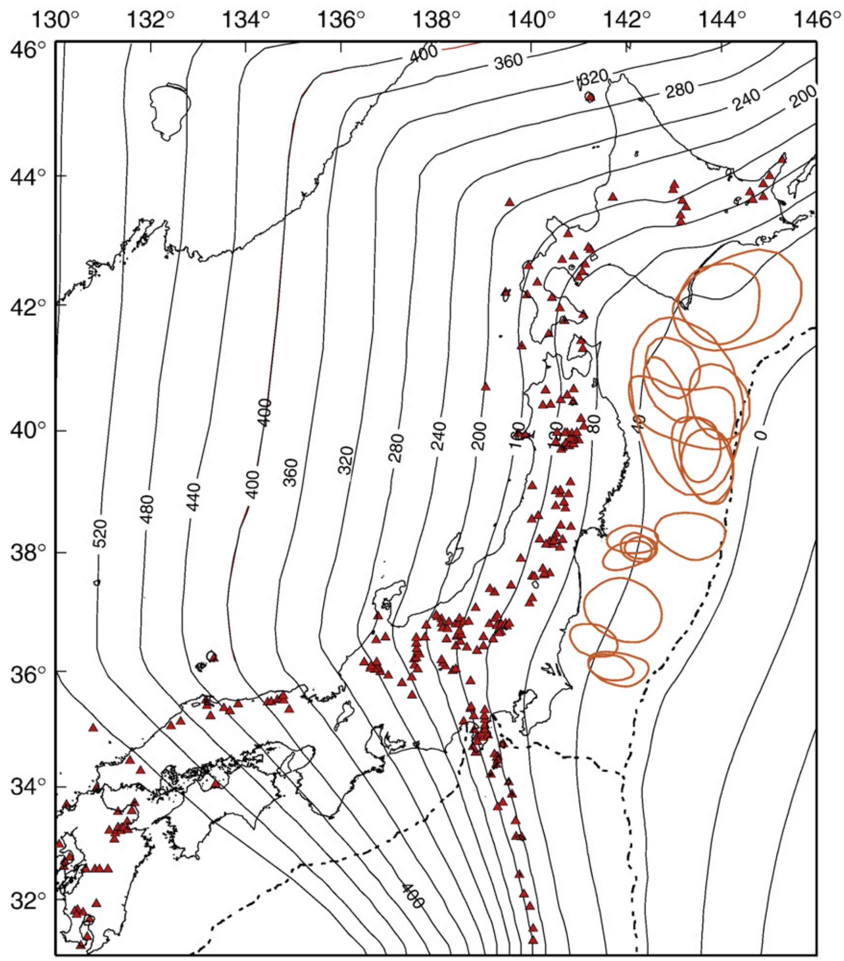

Fig. 3. Map showing iso-depth contours of the upper surface of the PAC slab (Nakajima et al., 2008). Source areas of $M \geq 7$ interplate earthquakes on the upper surface of the PAC plate in the past 80 years (Umino et al., 1990) are enclosed by orange curves. Red triangles denote Quaternary volcanoes.

perhaps slightly below it. However, the amount of this discrepancy is not very large, and does not affect the conclusions of the present paper, since the discussion is mainly focused on the shallower portion shallower than ca. 150 km. The hypocenter data were obtained from the Japan Meteorological Agency (JMA) unified catalogue based on the dense nationwide seismic network and from relocated hypocenters determined by the double-difference hypocenter location method (Waldhauser and Ellsworth, 2000).

ScSp waves have also been used to estimate the location of the upper surface of the PHS plate in southwest Japan (Nakanishi, 1980; Nakanishi et al., 1981; Iidaka et al., 1990). Through analysis of the seismograms of intraslab earthquakes, Fukao et al. (1983) detected seismic guided waves propagating in the crust of the subducting PHS plate, an observation that implies that the subducting crust remains untransformed to depths of at least 60 km. Later, Hori et al. (1985), Hori (1990) and Ohkura (2000) showed that such guided waves propagating in the untransformed oceanic slab crust are observed for earthquakes in a wider area including Kinki, Chugoku, Shikoku and Kanto. The results of these studies suggest that many earthquakes in the PHS slab occur in the uppermost portion, that is, in the oceanic slab crust. Until recently, the configuration of the subducting PHS plate has been estimated mainly on the basis of such seismicity data by setting the plate boundary so as to coincide with the upper envelope of intraslab earthquakes (e.g., Mizoue et al., 1983; Kasahara, 1985; Yamazaki and Ooida, 1985; Ishida, 1992; Noguchi, 1996, 1998; Miyoshi and Ishibashi, 2004; Hori, 2006).

The results of recent seismic reflection and refraction surveys carried out in southwest Japan, however, suggest that the plate boundary is located several kilometers shallower than the upper limit of intraslab seismicity (Kodaira et al., 2000, 2002; Kurashimo et al., 2002; Nakanishi et al., 2002; Kodaira et al., 2004; Ito et al., 2005), implying that the upper envelope of intraslab seismicity does not necessarily correspond to the plate boundary. Seismic tomography studies have provided important clues for understanding subduction zone structure in this area (e.g., Hirahara, 1980; Hirahara and Mikumo, 1980; Horie and Aki, 1982; Ohmi and Hurukawa, 1996; Kamiya and Kobayashi, 2000; Matsubara et al., 2005, 2008). Recent studies by Hirose et al. (2008a,b) reported detailed structures for the crust and upper mantle beneath southwest Japan based on the double-difference seismic tomography of Zhang and Thurber (2003, 2006), revealing an anomalous zone of several kilometers in thickness dipping shallowly in the plate subduction direction at depths near the top of the PHS plate over a wide area of southwest Japan. This anomalous zone exhibits characteristically low S-wave velocity and high ratio of P-wave to S-wave velocity (V_p/V_s). Comparison of this layer with the results of seismic reflection and refraction surveys (Kodaira et al., 2000, 2002, 2004; Kurashimo et al., 2002; Ito et al., 2005) and with the results of receiver function analyses (Yamauchi et al., 2003; Shiomi et al., 2004, 2006) suggests that this layer corresponds to the crust of the subducting PHS plate. Hirose et al. (2008a,b) thus estimated the configuration of the upper surface of the PHS plate in the Kanto region and in the region extending from Tokai to Kyushu, regarding the upper surface of the dipping anomalous layer as the plate boundary. The hypocenters of low-angle thrust-type events relocated by the double-difference method (Waldhauser and Ellsworth, 2000) were also employed in that study to delineate the plate boundary at shallower depths. The configuration of the subducting PHS plate thus estimated is shown in Fig. 4. The plate configuration model of Baba et al. (2002), obtained by compiling airgun seismic survey results, is adopted as a basis for estimating the plate boundary at shallow depths off the coast, and the estimation at greater depths is based on the seismic tomography study of Nakajima and Hasegawa (2007a). The configuration of the subducting PHS plate can thus be obtained for the entire region of southwest Japan, from Kanto to Kyushu, except for the Izu collision zone where the arc crust of the Izu Peninsula on the PHS plate collides with the NAM plate (e.g., Hyodo and Niitsuma, 1986; Soh et al., 1998).

The region north of the Izu Peninsula is characterized by a large fan-shaped gap in intraslab seismicity (e.g., Ishida, 1992), and by a sparse distribution of Quaternary volcanoes. Some researchers have suggested that the slab tears in this region, splitting the PHS slab into two parts referred to as the Kanto and Tokai slabs (e.g., Ishida, 1992; Mazzotti et al., 1999). Other researchers, however, have proposed the existence of an aseismic portion of the PHS slab to the north of the Izu Peninsula (e.g., Iidaka et al., 1990; Sekiguchi, 2001; Nakamichi et al., 2007; Matsubara et al., 2008). A recent seismic tomographic study of central Japan by Nakajima et al. (2009) has shown that a seismic high-velocity zone extends continuously down to a depth of approximately 140 km in the region north of the Izu Peninsula, suggesting that the PHS slab continues to subduct in this region without disruption. Using the tomography results, the PHS slab north of the Izu Peninsula was successfully connected with the slab geometry in Kanto and Tokai (Fig. 4).

Nakamura et al. (2008) compared geochemical data for rocks from 28 Quaternary volcanoes in central Japan with the configuration of the PHS and PAC slabs subducted below. They found a distinct correspondence of regional variations in fluid flux and composition to the configuration of the two slabs. Although the configuration of the PHS slab which they adopted is slightly different in Kanto from that shown in Fig. 4, comparison of geochemical data (Nakamura et al., 2008) with the updated slab configuration by Nakajima et al. (2009) (Fig. 4) gives a similar correspondence. Volcanoes in the Nasu area located right above the northern edge of the PHS slab or north of it have a significant amount of PAC fluid and a smaller amount of PHS fluid, whereas those in the Ryohaku area which are located above the steeply dipping PHS slab have a large amount of PHS fluid. In the Fuji–Myoko area below which the former Izu arc is subducting, a small amount of PHS fluid is included probably due to the relatively anhydrous PHS slab there (Nakamura et al., 2008). This dry condition of the subducting PHS slab also causes its aseismic nature in the north of the Izu Peninsula (Seno and Yamazaki, 2003).

Yamamoto et al. (2009-this issue) discussed subduction of the Izu arc beneath Honshu based on geology of the collision zone of the Izu arc and seismic tomography images of the subducted PHS slab by Nakajima et al. (2009). They pointed out that paleogeographic reconstruction suggests that the Proto-Izu arc must have subducted northward back to 17 Ma, and later the Izu arc collided at 5 Ma. If all of those arcs with ca. 20–30 km thick crust were accreted under the Honshu arc, the thickness of the crust there must be about 100 km. There is no evidence for such an unusually thick crust in central Honshu, suggesting most portion of the arc crust is subducted for the last 17 Ma. Seismic tomography study by Nakajima et al. (2009) clearly imaged this subducting PHS slab beneath the Honshu arc north of the Izu Peninsula as a high-velocity zone. According to the recent studies, it seems that such an arc subduction is common in the western Pacific region (Maruyama et al., 2009; Komabayashi et al., 2009; Santosh et al., 2009; Yamamoto et al., 2009; Senshu et al., 2009).

2.2. Slab–slab contact zone in Kanto

To the south of the Sagami Trough, the PAC plate subducts below the PHS plate west–northwestward along the Izu–Bonin Trench, bringing the plates into contact at shallow depths. This contact zone is

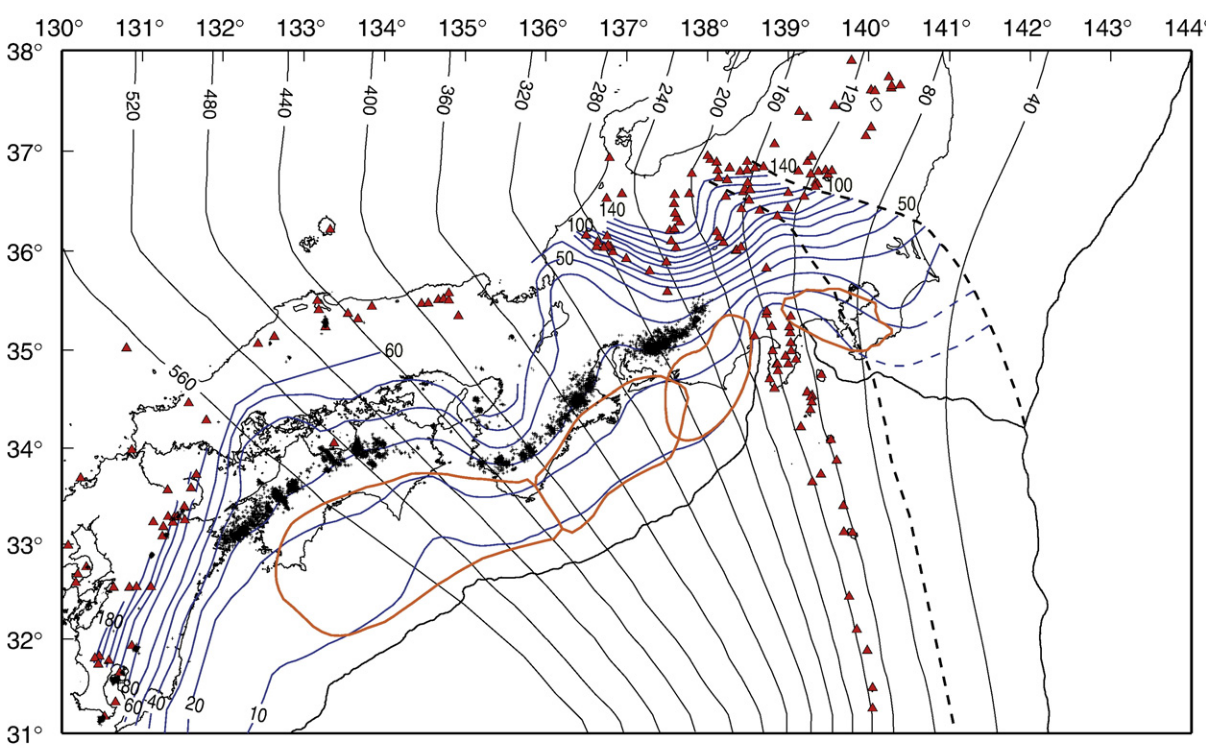

Fig. 4. Map showing iso-depth contours of the upper surface of the PHS and PAC slabs (Nakajima et al., 2008). The contact zone between the PHS and PAC slabs is enclosed by thick broken curves. Source areas of the 1923 Kanto earthquake estimated by Wald and Somerville (1995), and anticipated source areas of forthcoming Tokai, Tonankai, and Nankai earthquakes (Headquarters for Earthquake Research Promotion, Ministry of Education, Culture, Sports, Science and Technology, http:// www.jishin.go.jp/main/index.html.) are enclosed by orange curves.

expected to deepen and migrate northwestward after subduction of the PHS plate in Kanto, following the westward dip of the immediately underlying PAC plate. The location of this contact zone in Kanto has recently been estimated from the spatial distribution of slip vectors of interplate earthquakes and from seismic tomographic images. It should be possible to estimate the northern edge of this slab contact zone at shallow depths from the slip vectors of interplate earthquakes, since the direction of relative plate motion between the PAC and PHS

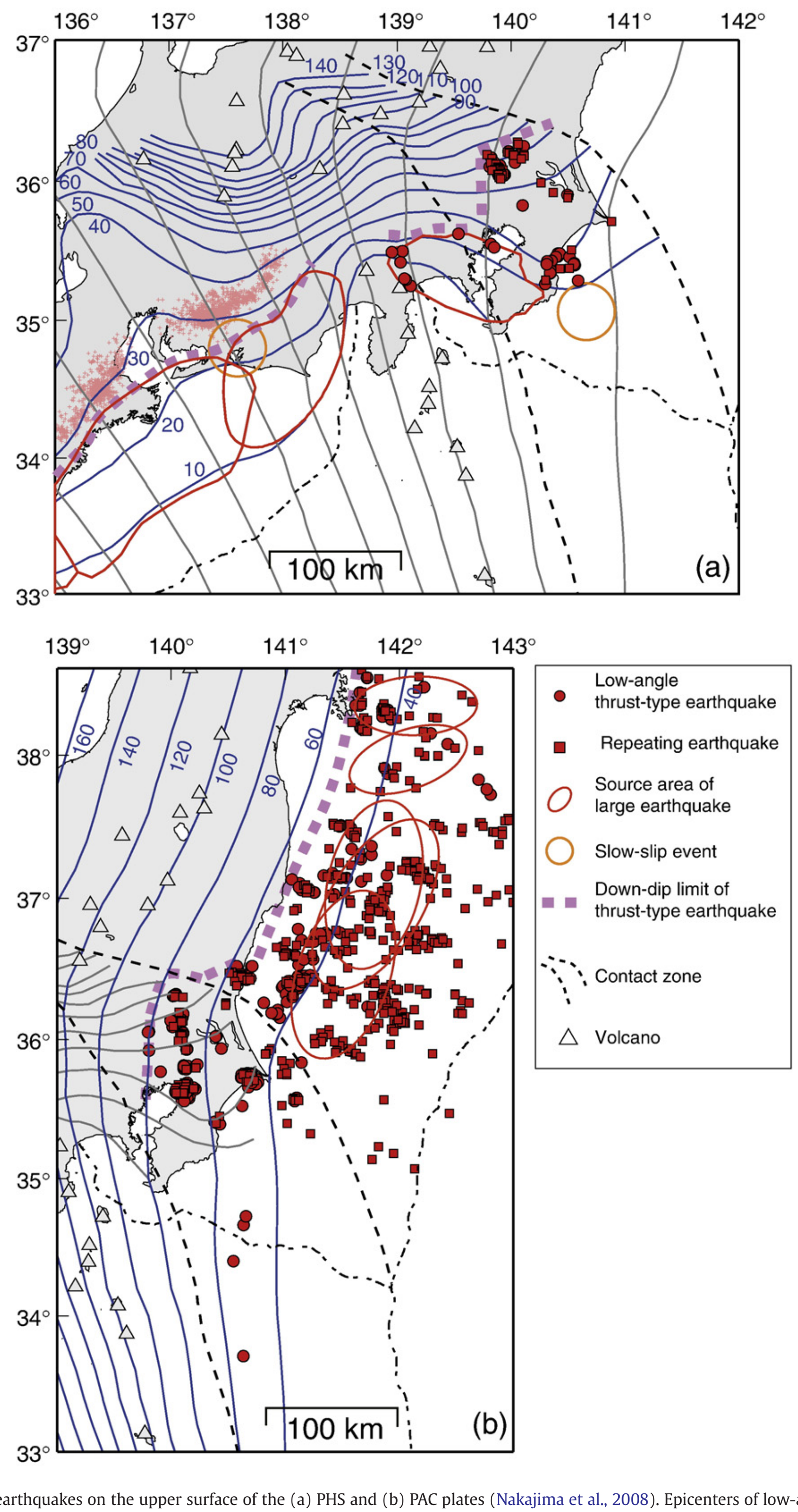

Fig. 5. Distribution of interplate earthquakes on the upper surface of the (a) PHS and (b) PAC plates (Nakajima et al., 2008). Epicenters of low-angle thrust-type events and small repeating events (Uchida et al., 2007) are shown by red circles and red squares, respectively. Downdip limit of interplate earthquake is shown by pink broken lines. Epicenter of non-volcanic deep low-frequency tremors/earthquakes are from Obara (2002).

plates differs slightly from that between the PAC and NAM plates (Seno and Takano, 1989). In a recent study by Uchida et al. (2009), a large number of focal mechanism solutions for interplate events from the catalogue of the Full Range Seismograph Network of Japan (F-net) focal mechanism database (Broadband Seismic Network Laboratory, NIED, 2008) were used to estimate the location of the northern edge of the slab contact zone with higher resolution than previously achieved.

The southern edge of the slab contact zone can be estimated from the precise location and configuration of the upper surface of the PAC slab and the lower surface of the PHS slab. Recently, Nakajima et al. (2008) reported an accurate location for the upper surface of the PAC slab based on relocated hypocenter distributions, focal mechanisms including newly determined results, and seismic tomography. Through estimation of the bottom of the PHS slab based on hypocenter distributions and seismic tomography, it was suggested that the PHS slab is approximately 60 km thick in the Kanto region, which is about twice the estimated slab thickness in the Tokai to Shikoku region (e.g., Ishida, 1992). Thick high-velocity zone corresponding to the mantle of the subducted PHS slab beneath Kanto was also detected by seismic tomographic inversions of Kamiya and Kobayashi (2007). These results are consistent with that expected from the slab age in the Kanto region (>48 Ma; Seno and Maruyama, 1984). The contact zone between the PAC and PHS slabs thus estimated is indicated in Fig. 4. From the iso-depth contours of the upper surface of the PHS slab shown in the figure, the upper PHS plate surface can be seen to undergo large fluctuations in the N–S direction, extending into the area north of the Izu Peninsula. This wavy configuration of the PHS slab is inferred to be related to contraction and buckling deformation in the along-arc direction associated with subduction beneath southwest Japan. The anomalous seismic activity associated with slab–slab contact beneath Kanto is discussed later.

3. Interplate earthquakes

3.1. Seismicity on the plate boundary

Most of the small shallow earthquakes in northeast Japan occur beneath the Pacific Ocean between the Kuril and Japan trenches and the Pacific coast (Fig. 2). These shallow events are located mainly along the boundary between the subducting PAC plate and the overlying NAM plate. Large destructive earthquakes, such as the 2003 *M* 8.0 Tokachi-oki earthquake, have occurred frequently along the main thrust zone in the Pacific Ocean (Fig. 3). The downdip limit of these interplate events estimated from the distribution of low-angle thrust-type earthquakes is subparallel to an isodepth contour of ca. 50 km (e.g., Igarashi et al., 2001; Uchida et al., 2003).

In contrast, few small shallow earthquakes occur in the Pacific Ocean between the Nankai Trough and the Pacific coast in southwest Japan (Fig. 2), yet large destructive earthquakes of *M* 8 or greater have occurred repeatedly in this region. The downdip limit of interplate coupling estimated from the source areas of large interplate earthquakes (Fig. 4) is again subparallel to an isodepth contour of 25–30 km on the plate interface. Deep non-volcanic low-frequency tremors/earthquakes, presumably occurring in the transition zone between locked and stably sliding zones of the plate boundary (Obara, 2002; Obara et al., 2004; Shelly et al., 2006; Ide et al., 2007), are located subparallel to an isodepth contour of 30–35 km, close to the downdip limit of the source areas of large interplate earthquakes. This result indicates that the downdip limit of the source areas of large interplate earthquakes appears to coincide with the downdip limit of interplate coupling in this area. The difference in the downdip limit of interplate coupling or interplate earthquakes between the PAC Plate subducting beneath northeast Japan and the PHS Plate subducting beneath southwest Japan is likely to be due to the difference in slab age, since the downdip limit is considered to be controlled by temperature (e.g., Hyndman et al., 1995, 1997; Yoshioka and Murakami, 2007).

The downdip limit of interplate earthquakes appears to deepen locally in the slab contact zone in Kanto for both the PAC and PHS plates, as shown in Fig. 5 (Nakajima et al., 2008). This anomalous deepening of the downdip limit of interplate events above and on the slab contact zone is perhaps due to lower temperature condition there resulting from the contact of the two plates. The contact of the PHS plate with the PAC plate hinders the heating of the PAC slab by hot mantle wedge (Hasegawa et al., 2007a; Noguchi, 2007). The contact with the cold PAC plate immediately below would also keep the PHS plate at lower temperatures (Uchida et al., 2009) as well as inherent low temperature condition because of the subduction of colder forearc materials of older plate (>48 Ma) (Seno and Maruyama, 1984).

3.2. Asperity model

The seismic coupling coefficient at the plate boundary east of northern Tohoku is estimated to be approximately 25%, while that east of southern Tohoku–Kanto is approximately 10% or less (e.g., Kanamori, 1977; Seno, 1979; Kato, 1979; Peterson and Seno, 1984; Pacheco et al., 1993). These coupling rates suggest that much of the interplate slip is accommodated by aseismic slip in these regions. Two end-member models have been used to describe interplate coupling and the recurrence of large interplate earthquakes; the asperity model, and the uniform coupling model. The asperity model divides the plate boundary into areas of strong frictional coupling dominated by stick–slip behavior (asperities), and intervening areas of weak frictional coupling dominated by stable sliding, as shown in Fig. 6 (Kanamori, 1981; Lay and Kanamori, 1981; Lay et al., 1982; Boatwright and Cocco, 1996). According to this model, asperities are persistent features and large slip areas of successive ruptures will occur in the same location on a given section of the plate boundary. In the uniform coupling model, the plate boundary is described as having spatially uniform frictional coupling, where the recurrence of large interplate earthquakes and the slip distributions vary with time.

Seismic waveform inversion studies for northeast Japan have provided clear results in support of the asperity model (Nagai et al., 2001; Yamanaka and Kikuchi, 2003, 2004; Okada et al., 2003; Yagi and Kikuchi, 2003; Yagi et al., 2003). Fig. 7 presents an example of these studies, showing the slip distributions for the 2003 *M* 8.0 and 1952 *M* 8.1 Tokachi-oki earthquakes, which occurred on the plate boundary south of Hokkaido (Yamanaka and Kikuchi, 2003). These two successive large earthquakes ruptured in nearly the same location on the plate boundary, supporting the asperity model. Yamanaka and Kikuchi (2004)

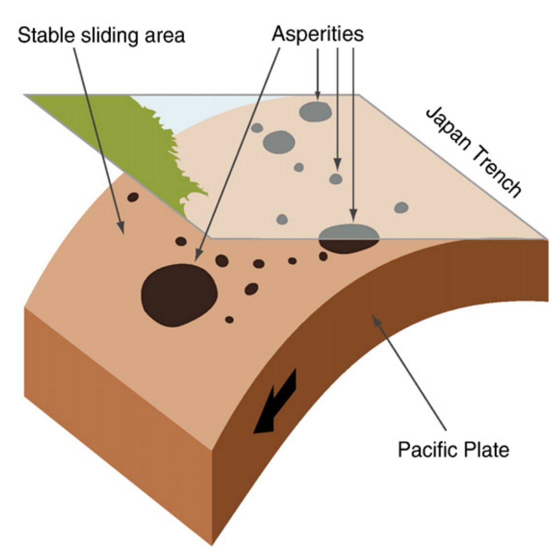

Fig. 6. Schematic illustration of asperities and stable sliding areas on the plate boundary east of NE Japan.

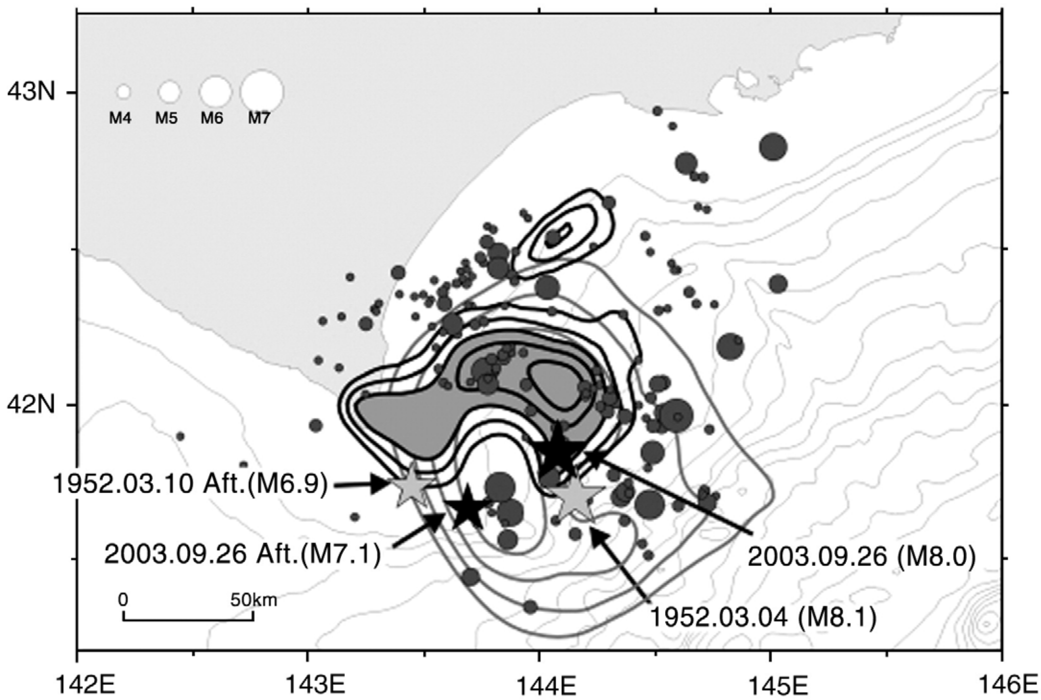

Fig. 7. Slip distribution for the 1952 M 8.1 and 2003 M 8.0 Tokachi–oki earthquakes (Yamanaka and Kikuchi, 2003). Mainshock and largest aftershock are indicated by large and small stars (red, 2003; black, 1952). Solid circles denote aftershocks of the 2003 event.

conducted a systematic study of the slip distributions of large earthquakes occurring in the last 70 years on the upper interface of the subducting PAC plate in northeast Japan based on the inversion of waveforms recorded by strong-motion seismographs. The results reveal a distinct tendency for the large-slip areas of successive large interplate events to occur in the same location on the plate boundary, as expected from the asperity model. It was further found that the typical size of individual asperities correspond to M 7-class events, while M 8-class events are caused by the simultaneous rupture of more than one asperity.

Repeating ruptures for much smaller size asperities have also been identified on the subducting plate boundary in northeast Japan. Matsuzawa et al. (2002) found a repeating earthquake sequence of M 4.8 ± 0.1 that occurred regularly at the same location on the plate boundary offshore of Kamaishi, east of northern Tohoku. These repeating earthquakes, having nearly the same waveforms, were found to recur at a regular interval of 5.35 ± 0.53 years. The estimated slip distributions for the three most recent events (Fig. 8), which are of remarkably similar magnitude (M 4.8 ± 0.1), indicate that the spatial extent of the rupture areas for these events are almost coincident (diameter, 1.0–1.5 km). Based on the asperity model, this earthquake sequence can be interpreted to be generated by the repeated rupture of the same asperity (frictionally locked patch) with a size of approximately 1 km, while the plate boundary surrounding the asperity undergoes stable sliding. There are no historical records of earthquakes of $M > 6$ in the stable sliding area surrounding the inferred asperity for this earthquake sequence, whereas microearthquake activity is very high, suggesting continuous creep on the plate boundary in this area. Based on this interpretation, Matsuzawa et al. (1999) predicted that the next event would occur by the end of November 2001 with 99% probability. An M 4.7 event was subsequently recorded on November 13, 2001 (Matsuzawa et al., 2002). Similarly, the next event was predicted to occur by the end of January 2008 with 68% probability (Matsuzawa et al., 2002), and an M 4.7 event was recorded as expected on January 11, 2008 at the same location on the plate boundary (Fig. 8) (Okada et al., 2003; Shimamura et al., 2008).

The existence of small repeating earthquake sequences, similar to the M 4.8 ± 0.1 earthquake sequence offshore of Kamaishi, was further confirmed in subsequent studies by Igarashi et al. (2003) and Uchida et al. (2003). In those studies, many repeating earthquake sequences of M 3–4 class were identified in the plate boundary region east of Tohoku. These earthquakes sequences were estimated to be caused by the repeated slip of small asperity patches of 0.1–1 km in size surrounded by creep areas on the plate boundary.

The GEONET dense network of global positioning system (GPS) stations in Japan, comprising approximately 1200 permanent stations in all areas of the Japan Islands, provides extensive data on crustal movement, and has revealed important information on the spatial

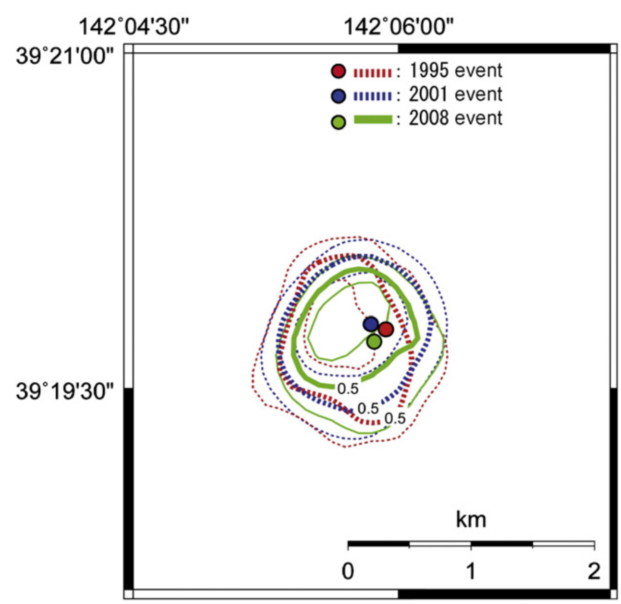

Fig. 8. Slip distribution for the Kamaishi–oki repeating earthquakes (Okada et al., 2003; Shimamura et al., 2008). Red broken, blue broken and green curves show slip distributions of 1995, 2001 and 2008 events. Hypocenters of those events are denoted by red, blue and green circles, respectively.

distribution of interplate coupling. Nishimura et al. (2000) and Ito et al. (2000) estimated the spatial and temporal distribution of seismic and aseismic slip along the plate boundary east of Tohoku through backslip inversion of terrestrial GPS data, updating the previous estimation from geodetic data inversion (El-Fiky and Kato, 1999). Suwa et al. (2006) subsequently derived the spatial distribution of backslip on the plate boundary east of Tohoku and south of Hokkaido based on backslip inversion of GPS data (Fig. 9). These studies revealed two regions of strong interplate coupling on the plate boundary; east of Miyagi, and east of Aomori (Tohoku) extending south of Tokachi (Hokkaido). In these areas, the interplate coupling estimated for the 5 year period from 1997 to 2002 is approximately 75–100%. The large asperities left unruptured (having long lapse times from their last ruptures), of those estimated by Yamanaka and Kikuchi (2004) based on strong motion records for the last 70 years, are located in these two strong coupling areas. It should be noted here that asperities east off south Tohoku were not included in their estimation, and so this area should be disregarded for the comparison. Among the four unruptured asperities shown in the figure, one south of Tokachi ruptured in 2003, producing the *M* 8.0 Tokachi-oki earthquake, the slip distribution of which is shown in Fig. 7.

The spatial distribution of interplate coupling has also been estimated by analysis of repeating earthquakes. Fig. 10 shows the spatial distribution of interplate slip rate estimated from repeating earthquake data (Uchida et al., 2004). In this estimation, the slip rate

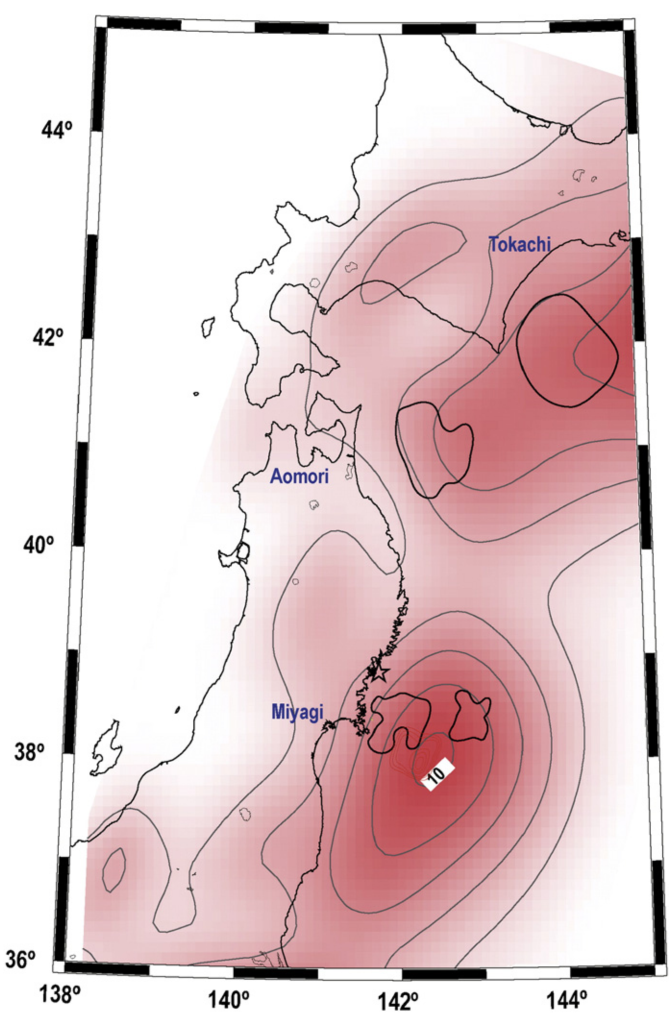

Fig. 9. Backslip distribution along the plate boundary east of NE Japan estimated from GPS data for the period 1997–2002 (Suwa et al., 2006). Contours intervals, 2 cm/yr. Solid outlines east of the Japan Islands denote asperities unruptured prior to 2002.

at each of the small asperities is calculated from the rate of earthquake occurrence assuming a scaling relationship between seismic slip and the scalar moment, as proposed by Nadeau and Johnson (1998). The observations of Igarashi et al. (2003) have shown that repeating earthquake sequences on the subducting plate boundary east of Tohoku satisfy the scaling relationship of Nadeau and Johnson (1998), which was originally developed for repeating earthquake sequences on the transform plate boundary of the San Andreas Fault. Comparison of the slip rate distribution estimated from repeating earthquake data with the backslip distribution estimated from GPS data reveals good consistency between the two data sets. Repeating earthquakes do not occur in areas where the plate boundary is completely locked, which appear as vacant areas in Fig. 10.

These observations indicate that the asperity model is applicable to the process of seismic and aseismic slip on the upper boundary of the subducting PAC plate east of northeast Japan. The spatial distribution of asperities on the plate boundary east of Tohoku is shown schematically in Fig. 6. Large and small isolated asperities are scattered on the plate boundary, surrounded by stably sliding areas. Downdip limit of asperities is located at a depth of ca. 50 km, and no large asperities exist east of Iwate. Asperities are locked during the interseismic period, and the subsequent failure and repeated rupture of these asperities cause interplate earthquakes.

At present, the cause of asperity formation on the plate boundary is unclear. However, seismic tomographic studies have provided important information on this issue. Mishra et al. (2003) and Zhao et al. (2007a,b) conducted tomography studies for the fore arc region of Tohoku by using arrival time data for earthquakes beneath the Pacific Ocean observed at land station (see also Zhao et al., 2007a,b; Zhao and Ohtani, 2009-this issue; Zhao, 2009). In those studies, the focal depths of offshore earthquakes were determined using the sP phase, a depth phase observable at short epicentral distances and originally reported by Umino et al. (1995). The sP phase allows the seismic velocity structure external to the seismic network on land to be resolved. The results reveal strong lateral heterogeneities in the mantle wedge of the fore arc region of Tohoku. Comparison with the source areas of large earthquakes on the plate boundary suggests that large interplate earthquakes mainly occur below the high-velocity areas and not below the low-velocity areas of the hanging wall side of the plate boundary. These low-velocity areas in the fore arc mantle wedge, in direct contact with the subducting PAC plate and immediately above it, tend to have high Vp/Vs and probably correspond to hydrated mantle affected by slab-origin water, which allows aseismic slip to occur on the plate interface due to the stable sliding nature of serpentinized mantle (Peacock and Hyndman, 1999).

Such a tendency of the location of large interplate earthquakes described above was partly confirmed by a seismic tomography study of Yamamoto et al. (2006) using ocean bottom seismograph (OBS) network data. In that study, a precise seismic velocity structure was reported for the region encompassing the source area of the 1978 *M* 7.4 Miyagi-oki earthquake based on a combination of terrestrial and offshore seismic station data. In this area of the plate interface, earthquakes of *M* 7.5 class recur at an interval of approximately 37 years, with the next event predicted with 50% probability to occur 7 years from now (Headquarters for Earthquake Research Promotion, 2003). Due to the high probability of an earthquake in this region, a temporary OBS network has been deployed in the epicentral region of the predicted event in order to allow the seismic velocity structure of the source area to be investigated at high resolution. The results acquired using this temporary network reveal a lateral variation in seismic velocity in the fore arc mantle wedge. The mantle wedge immediately above the rupture area (asperity) of the 1978 event has the normal velocity of peridotite, whereas the mantle wedge north of the asperity exhibits anomalously low velocities, suggesting the existence of serpentinized mantle, similar to the results reported for

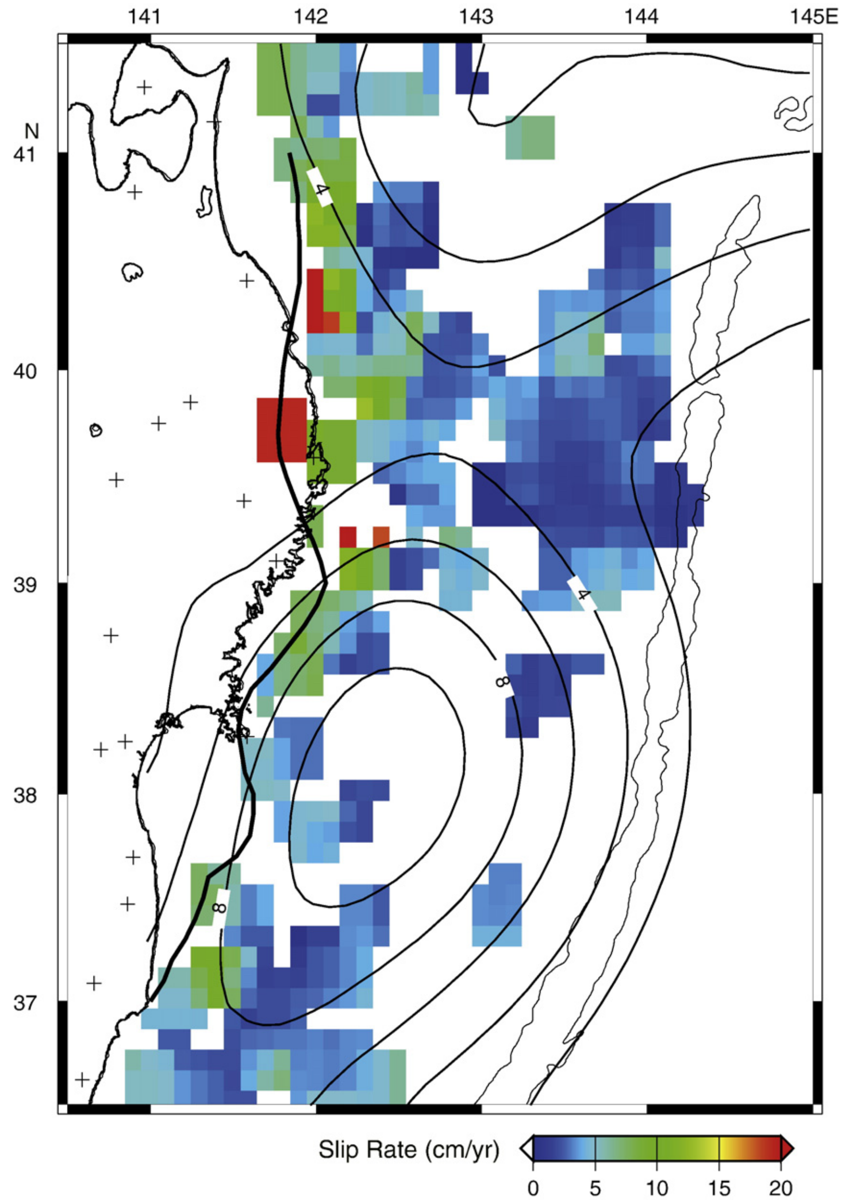

Fig. 10. Slip rate distribution along the plate boundary east of Tohoku estimated from repeating earthquake data for the period 1997–2002 (Uchida et al., 2003). Slip rate is shown by the color scale. Contours denote backslip distribution estimated from GPS data by Suwa et al. (2006), and bold line close to the Pacific coast denotes the downdip limit of interplate earthquakes (Igarashi et al., 2001).

the fore arc region of Tohoku by Mishra et al. (2003) and Zhao et al. (2007a,b).

The temporary deployment of dense seismic networks in the vicinity of recent large inland crustal earthquakes has yielded valuable data that has made it possible to resolve the seismic velocity structure near the fault planes in unprecedented detail. Tomographic inversions of large volumes of high-quality aftershock data have shown that there exists a systematic relationship between areas of high seismic velocity and large slip areas (asperities) associated with the mainshock rupture (e.g., Okada et al., 2006a,b; Kato et al., 2006; Okada et al., 2007). For example, Okada et al. (2006a,b) reconstructed the source area of the 1995 Southern Hyogo (Kobe) earthquake at high resolution by applying the double-difference tomography method of Zhang and Thurber (2003) to aftershock observation data. The results reveal that large slip areas (asperities) of the mainshock rupture estimated by Yoshida et al. (1996) are located in high-velocity zones on the fault plane, and avoid low-velocity zones (Fig. 11). These high-velocity zones extend for at least several kilometers on both sides of the fault plane, suggesting that the formation of asperities is closely related to this inhomogeneous seismic velocity structure surrounding the mainshock fault plane.

3.3. Episodic slow slip and low-frequency tremors/earthquakes on the plate boundary

The dense nationwide GPS network data have provided information on rigid plate motion and interplate coupling in Japan as highlighted above. Back slip inversions of GPS data in this region show a strong interplate coupling on the plate boundary along the Nankai Trough and along the Sagami Trough: almost full coupling at shallow depths (ca. 5–25 km) gradually decreasing to zero at ca. 35 km (e.g., Yoshioka et al., 1993; Sagiya, 1999; Mazzotti et al., 2000; Sagiya, 2004). The GPS data have further revealed the existence of episodic slow slip along the plate boundary, even in the areas surrounding unruptured asperities. These observations suggest that episodic slow slip may play an important role in the stress loading process at asperities.

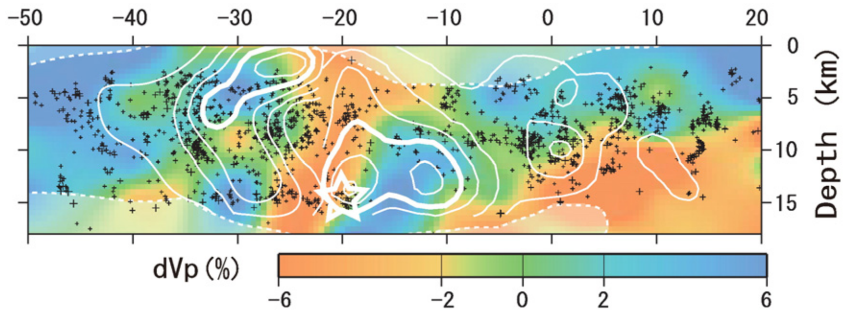

Fig. 11. Vertical cross-section of P-wave velocity perturbations along the fault plane of the 1995 southern Hyogo (Kobe) earthquake of M7.2 (Okada et al., 2007). Perturbations from the average velocity at each depth are shown by the color scale. White contour lines show the slip distribution of the main shock rupture (Yoshida et al., 1996) in an interval of 0.4 m. Deeper color means greater DWS value. Crosses denote aftershocks.

On the upper boundary of the PAC plate in northeast Japan, postseismic slip following interplate earthquakes has been detected for many events by GPS and extensometer networks (e.g., Yagi et al., 2003; Heki, 2007). Such postseismic slip has also been detected using small repeating earthquake data, and it has been shown that most of the $M>6.0$ events analyzed were followed by postseismic slip in the areas surrounding ruptured asperities (Uchida et al., 2003). These observations suggest that postseismic slip is a common phenomenon following interplate earthquakes along the upper surface of the PAC slab in northeast Japan.

On the plate boundary in southwest Japan, slow slip tends to occur in the form of independent episodic slow slip events. In this region, megathrust earthquakes of $M>8$ have occurred repeatedly in association with the subduction of the PHS plate along the Nankai and Sagami troughs. The recorded history of earthquake activity along the Nankai Trough can be traced back more than 1000 years, and reveals that great earthquakes typically occur in pairs separated by a short interval of days to years, with a recurrence interval of 100–200 years (Ando, 1975; Kumagai, 1996; Ishibashi and Satake, 1998; Ishibashi, 2004). Interplate coupling in this region is considered to be very high, seismic coupling coefficient being close to 100% (Kanamori, 1977; Yoshioka, 1991; Ito et al., 1999; Mazzotti et al., 2000; Ito and Hashimoto, 2004), and it is likely that the low earthquake activity along the Nankai Trough at present is related to this high coupling coefficient (see Fig. 2). Using dense GPS network data, episodic slow slip (silent earthquakes) was detected in the Boso Peninsula of Kanto in 1996 (Sagiya, 2004), in Tokai in 2000 (Ozawa et al., 2003), and in the Bungo Channel east of Kyushu in 1997 (Hirose et al., 1999; Ozawa et al., 2007). Investigations of past crustal deformation or GPS data have revealed the quasiperiodic recurrence of these silent earthquakes along the plate boundary over periods of months to years (Kimata and Yamauchi, 1998; Ozawa et al., 2002, 2003, 2004; Yamamoto et al., 2005; Miyazaki et al., 2006).

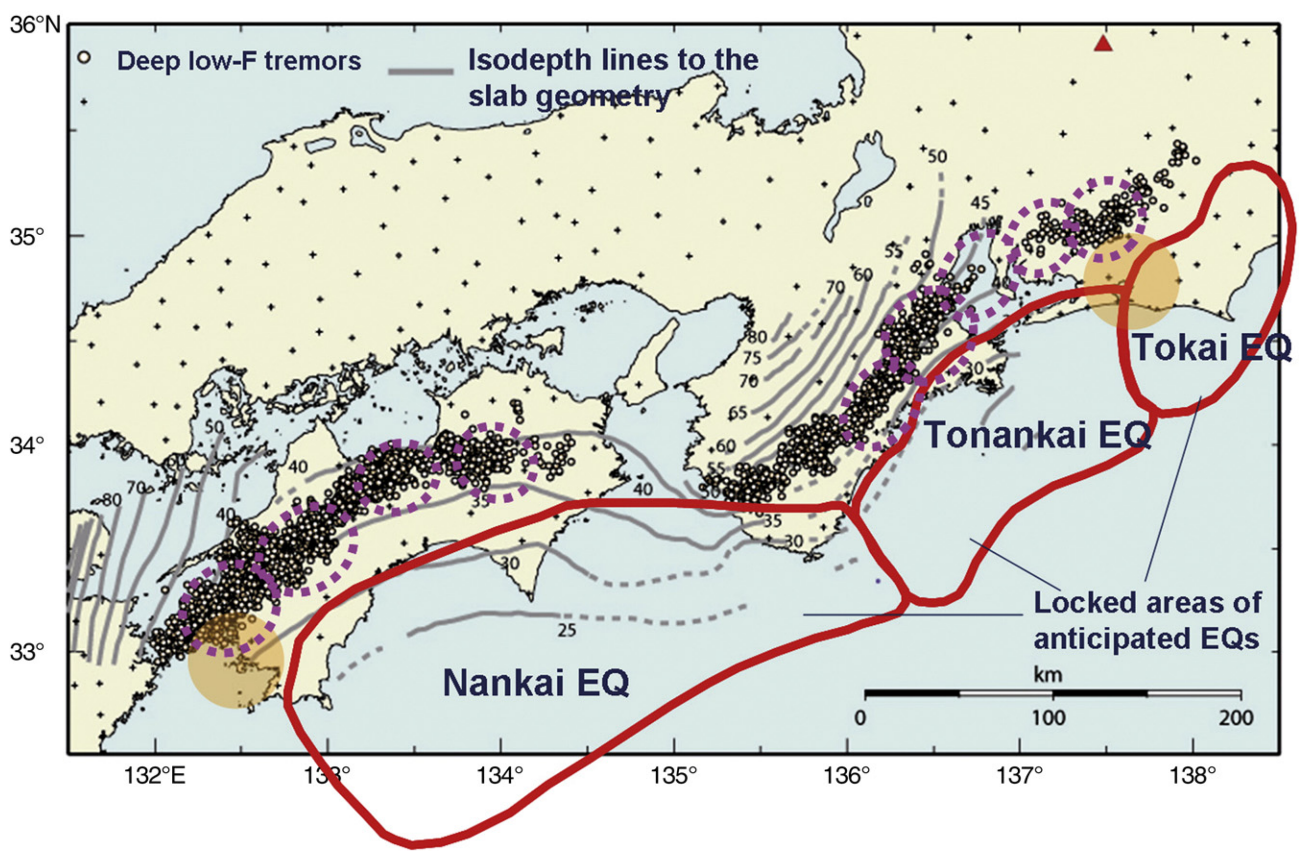

Fig. 12. Epicenter distribution for deep low-frequency tremors (Obara and Hirose, 2006). Broken outlines denote short-term slow slip events accompanying tremors, shaded areas denote long-term slow slip events, and bold outlines denote locked areas of megathrust earthquakes.

In addition to long-term slow slip events lasting months to years, short-term slow slip events accompanying non-volcanic deep low-frequency tremors/earthquakes have been detected at the downdip extension of the locked areas of the Tokai, Tonankai, and Nankai earthquakes (Obara, 2002; Katsumata and Kamaya, 2003; Obara et al., 2004; Obara and Hirose, 2006). Such short-term slow slip events and low-frequency tremor/earthquake activity last for several days and exhibit a distinct periodicity with an interval of approximately half a year, similar to those observed in the Cascadia subduction zone, which recur regularly at average intervals of approximately 14 months (Dragert et al., 2001; Rogers and Dragert, 2003). Fig. 12 presents the epicenter distribution of low-frequency tremors located by Obara (2002). The epicenters are distributed nearly parallel to the iso-depth contours of the subducting PHS slab in regions known to be adjacent to the locked areas of anticipated megathrust earthquakes. Detailed investigations of the mechanism of low-frequency tremors/earthquakes have revealed that such tremors represent swarms of small, low-frequency earthquakes generated by individual shear slip events on the plate interface (Shelly et al., 2006, 2007; Ide et al., 2007). These low-frequency earthquakes therefore appear to be the seismic signature of the associated slow slip events. Tremor (or low-frequency earthquake) and slow slip events can thus be interpreted as being different manifestations of the same process on the plate interface. Another manifestation of this process, shear slip on the plate interface, has recently been detected from the seismograms of tremors, representing very low-frequency earthquakes with a predominant period of approximately 20 s (Ito et al., 2007). This very low-frequency earthquakes also occur in the accretionary prism along the Nankai Trough (Ito and Obara, 2006), suggesting simultaneous occurrence of slow slip events there.

Fig. 13 shows the relative locations of low-frequency tremors, short-term slow slip events, long-term slow slip events, and locked areas of megathrust earthquakes along the Nankai subduction zone (Obara, in press). The periodically occurring slow slip events on the plate interface load stress to the nearby locked areas, incrementally increasing the probability of dynamic rupture of the locked area, resulting in a megathrust earthquake.

4. Intraslab earthquakes

4.1. Double-planed deep seismic zone and dehydration embrittlement

At the depths of intraslab earthquakes, lithostatic pressure is too high to allow brittle faulting. The generation of intraslab earthquakes therefore requires some form of weakening mechanism to be active in such areas. Dehydration embrittlement has been proposed as a possible mechanism for reducing strength and thus allowing the occurrence of earthquakes within subducted slabs (e.g., Raleigh and Paterson, 1965; Raleigh, 1967; Nishiyama, 1992; Kirby, 1995; Green and Houston, 1995; Kirby et al., 1996; Seno and Yamanaka, 1996; Peacock, 2001). If dehydration embrittlement is involved in the generation of intraslab earthquakes, it is expected that earthquakes will occur within the part of the slab containing hydrous minerals. Hacker et al. (2003b) showed that the intermediate-depth earthquakes in four different subduction zones (Tohoku, Costa Rica, Nankai, and Cascadia) only occur within a confined part of the slab that contains hydrous minerals, that is, where dehydration is expected to occur. Such earthquakes were found to be notably absent in the parts of the slabs predicted to be anhydrous.

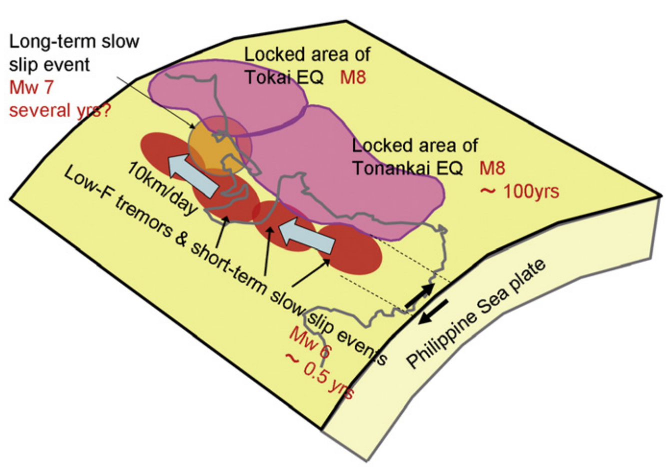

Fig. 13. Schematic illustration showing locations of deep low-frequency tremors, short-term slow slip events, long-term slow slip events, and locked areas of megathrust earthquakes along the plate boundary in the Nankai subduction zone (Obara, in press).

Yamasaki and Seno (2003) estimated the dehydration loci of metamorphosed slab crust and serpentinized slab mantle using experimentally derived phase diagrams for six subduction zones (Tohoku, N Taiwan, N Chile, Cape Mendocino, and E Aleutians). Their estimation showed that the lower plane seismicity of the double seismic zone is located at the lower dehydration loci of serpentine, whereas the upper plane seismicity is located in the slab crust, suggesting the upper plane seismicity is due to dehydration embrittlement of the slab crust. Concentration of earthquakes like this along the dehydration loci of serpentine, i.e., the lower boundary of a hydrated slab mantle, is considered to be the main cause for the formation of the double-planed deep seismic zone. This further implies that dehydration reaction is greater when materials reach the facies boundary in the slab mantle.

Omori et al. (2002) constructed a phase diagram for peridotite in the model system $MgO–Al_2O_3–SiO_2–H_2O$ based on thermodynamic calculations, and compared the estimated dehydration loci with intraslab seismicity for several subduction zones. It was found that the seismic zone estimated assuming that intraslab earthquakes are induced by dehydration embrittlement coincides well with the double seismic zones observed in each subduction zone. The dehydration-induced seismic zones thus estimated further reproduce the locations of earthquake clusters that occur between the upper and lower planes of the double seismic zone in the PAC slab in Tohoku. Extending this result, Omori et al. (2004) proposed that the dehydration embrittlement model is also applicable to deep earthquakes in the subducting slab peridotite.

If the dehydration embrittlement mechanism discussed by Yamasaki and Seno (2003) and Omori et al. (2002) is valid and a deep part of slab mantle is hydrated before its subduction, almost all subduction zones are expected to exhibit a double seismic zone, and the separation between the upper and lower planes is expected to narrow with decreasing age of the subducting oceanic plate. This expectation was confirmed by Brudzinski et al. (2007), who demonstrated that such double seismic zones can be found for all 16 subduction zones investigated, using the global seismicity data accurately relocated by Engdahl et al. (1998) and Engdahl (2006). There is thus considerable evidence supporting the involvement of dehydration embrittlement in the generation of intraslab earthquakes.

4.2. Phase transformation in slab crust and formation of upper-plane seismic belt

The recent resolution of the detailed hypocenter distribution and seismic velocity structure of subducted slabs by the dense seismic network deployed in Japan has provided further evidence supporting the dehydration embrittlement hypothesis. Precise relocation of intermediate-depth earthquakes in Hokkaido and Tohoku using high-quality seismic data and the double-difference hypocenter location method of Waldhauser and Ellsworth (2000) has revealed the existence of a pronounced seismic belt in the upper plane of the double seismic zone, which runs nearly parallel to the iso-depth contours of the upper surface of the PAC slab at depths of 70–90 km (Kita et al., 2006). Fig. 14 shows a cross-arc vertical cross-section of relocated intraslab earthquakes in central Tohoku in relation to the upper surface of the PAC slab, the slab Moho, and the facies boundaries in the oceanic slab crust. The plate interface in this case was estimated by inverting the travel times of seismic waves converted at the plate interface (Zhao et al., 1997a,b). The boundaries between the jadeite lawsonite blueschist (JLB) and lawsonite amphibole eclogite (LAE) facies, and between the LAE and eclogite facies (estimated by Hacker et al., 2003b using temperature distribution of Peacock and Wang, 1999) are also shown in Fig. 14 for the slab crust in central Tohoku. The H_2O contents in the JLB, LAE, and eclogite are 5.4%, 3.0%, and 0.1%, respectively (Hacker et al., 2003a). Dehydration is expected to occur in all H_2O-saturated rocks (i.e., JLB and LAE rocks), since the phase changes occur in a continuous manner due to the effect of solid solutions. Therefore, the generation of intraslab earthquakes is not restricted to the facies boundary, and is expected to occur throughout the slab in regions containing hydrated minerals.

From the data shown in Fig. 14, a significant decrease in earthquake activity in the slab crust can be seen immediately below both facies boundaries, and almost no earthquakes occur in the eclogite facies in the slab crust. This result is consistent with the predictions of the dehydration embrittlement hypothesis. Fig. 14 further shows a concentration of earthquakes in a part of the slab crust that corresponds to the upper-plane seismic belt. This concentration of upper-plane earthquakes lies close to the shallower side of the JLB–LAE facies boundary, which suggests that the dehydration reaction is more prominent as material reaches the facies boundary. Although the pressure and temperature in the above case include ranges of the phase diagram of Hacker et al. (2003a) that are not well defined and there exist considerable errors in the estimation of temperature in the slab, the trends are remarkably consistent with the predictions from the thermal–petrologic model. Of course, this result depends on the phase diagram of the crustal material and temperature distribution within the slab. For example, if we use the phase diagram by Schmidt and Poli (1998), the locations of the facies boundaries in Fig. 14 would move to deeper depths. On the other hand, if we adopt temperature distribution higher than Peacock and Wang (1999) as pointed out by Omori et al. (2009), the facies boundaries would move to shallower depths in turn. Higher temperature model is possible, since the estimation by Peacock and Wang (1999) is based on iso-viscous mantle wedge material. In order to verify the above result, we need more precise data of phase diagram at this pressure–temperature range and of temperature distribution within the subducted crust at this depth range.

Fig. 15 shows the epicenter distribution for earthquakes on the upper plane of the double seismic zone. A pronounced seismic belt aligned nearly parallel to the iso-depth contours of the upper surface of the PAC slab is apparent at depths of 70–90 km as indicated by a thick arrow. This upper-plane seismic belt, probably caused by the supply of H_2O associated with the JLB–LAE phase transformations, can be clearly resolved in Hokkaido and Tohoku. In the Kanto region, however, no such seismic belt oriented parallel to the iso-depth contours of the slab surface is observable, whereas a distinct seismic belt oriented oblique to the iso-depth contours can be recognized as indicated by a thick arrow. Hasegawa et al. (2007a) showed that this seismic belt in Kanto, aligned obliquely to the iso-depth contours and deepening toward the north from a depth of ca. 100 km to 140 km, corresponds to the upper-plane seismic belt.

In the Kanto region, the PAC slab is in direct contact with the subducting PHS slab above. The slab contact zone estimated by Nakajima et al. (2008) is shown in Fig. 15. The downdip limit of this contact zone is located close to and parallel to this obliquely oriented seismic belt, suggesting that the deepening of the seismic belt in the Kanto region is caused by the contact between the PAC and PHS slabs. Contact with the overlying cold PHS slab hinders heating of the PAC

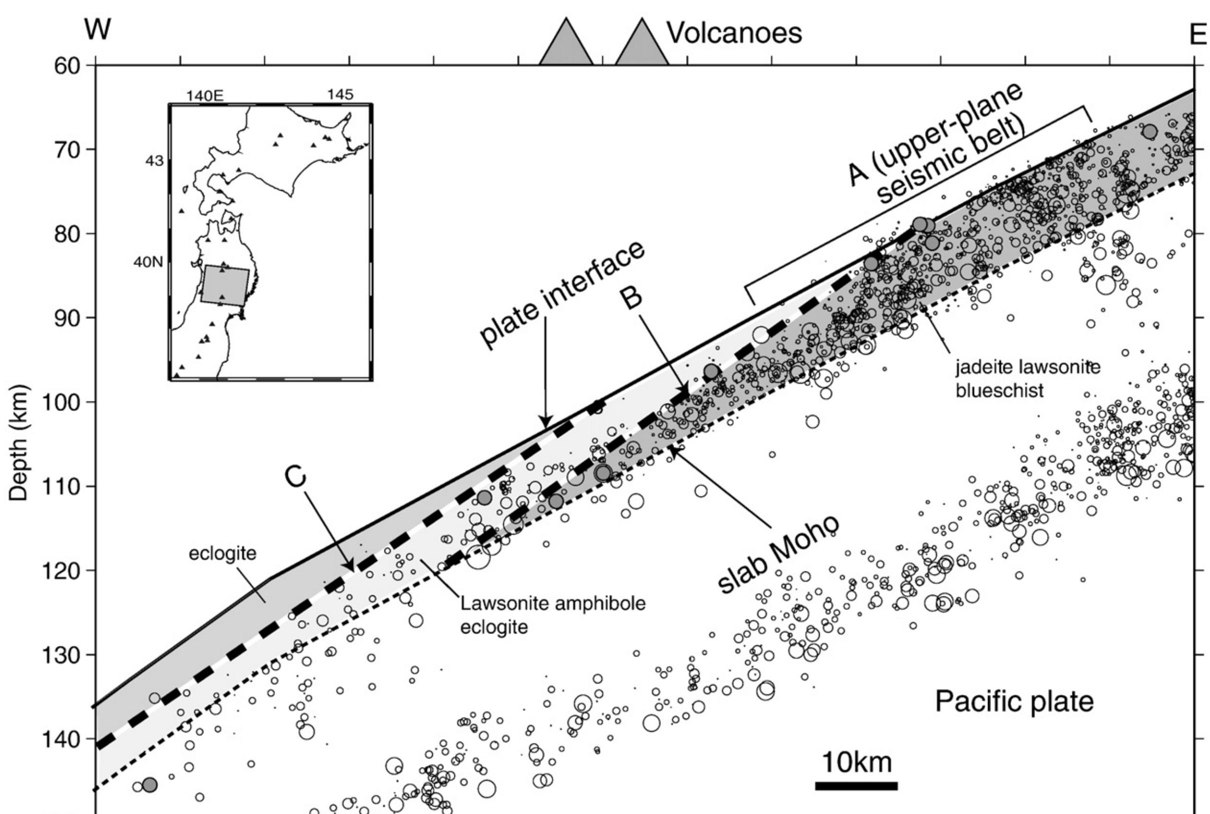

Fig. 14. Cross-arc vertical cross-section of intraslab earthquakes in central Tohoku (Kita et al., 2006). Relocated earthquakes are shown by open circles. A, upper-plane seismic belt; B JLB–LAE phase boundary; C, LAE–eclogite phase boundary (Hacker et al., 2003b).

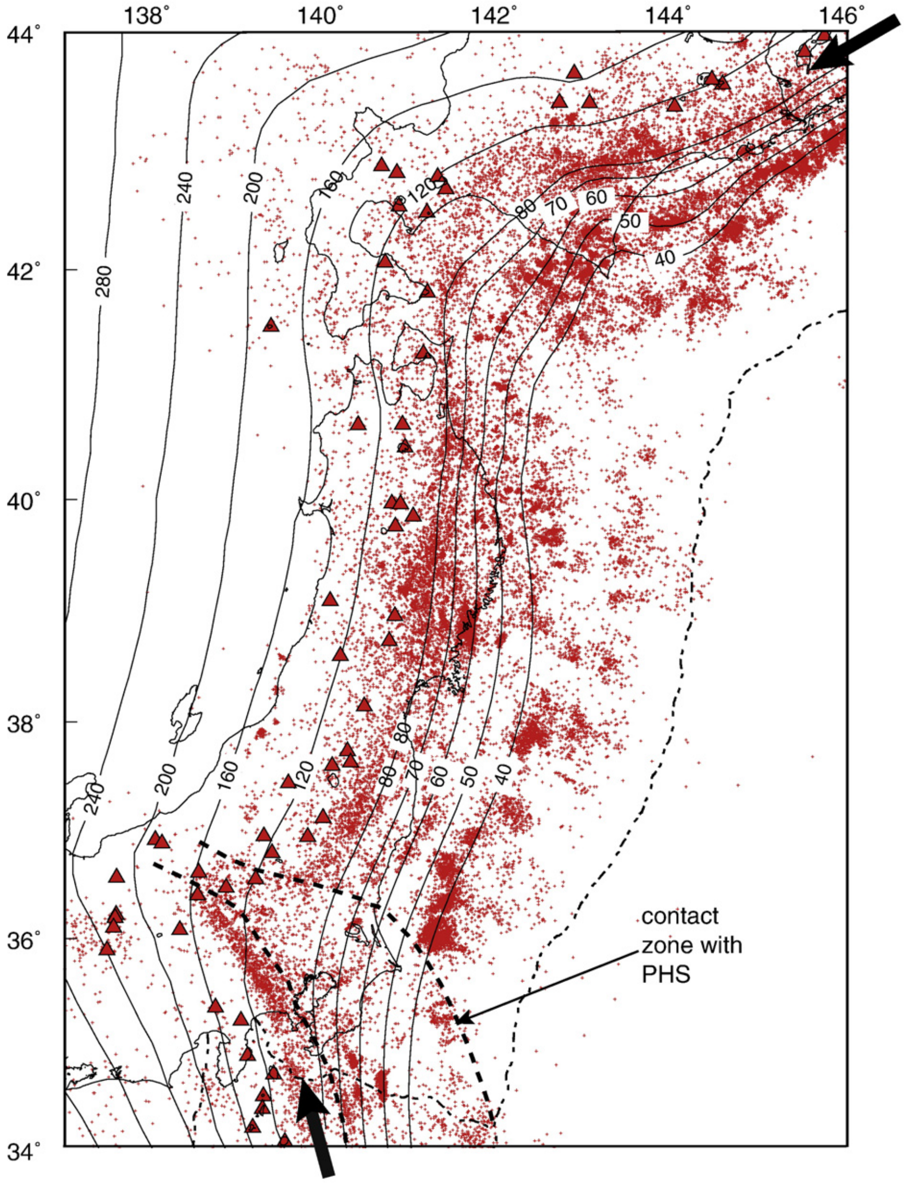

Fig. 15. Epicenter distribution for earthquakes on the upper plane of the double seismic zone in the PAC slab. Earthquakes 0–10 km below the upper plate interface are plotted. Solid curves are iso-depth contours of the upper surface of the Pacific slab. The contact zone with the PHS slab is enclosed by thick broken curves.

slab crust by the hot mantle wedge, delaying the eclogite-forming phase transformations and hence shifting the seismic belt to greater depth. This difference from Hokkaido and Tohoku accounts for the formation of the seismic belt oriented obliquely to the iso-depth contours of the slab surface beneath Kanto.

The upper-plane seismic belt in Kanto is narrower than those in Tohoku and Hokkaido. This can be explained also by the effect of thermal shielding by the overlying PHS slab. The PAC slab, after passing through the contact zone with the overlying PHS slab, comes into direct contact with the mantle wedge, causing rapid heating at the upper surface of the slab. This rapid heating is expected to steepen the dip angle of the JLB–LAE facies boundary in the slab crust compared with that shown in Fig. 14, resulting in a concentration of the seismic belt in a narrower band than that in Tohoku and Hokkaido (Hasegawa et al., 2007a).

Intraslab earthquakes within the crust of the subducted PAC slab form a pronounced seismic belt nearly parallel to 80-km iso-depth contour of the upper surface of the PAC slab beneath Hokkaido and Tohoku, but oblique to it beneath Kanto as described above. We inspected whether or not a seismic belt, similar to the upper-plane seismic belt in the PAC slab, exists within the crust of the PHS slab subducted beneath southwest Japan. Fig. 16 shows the epicenter distribution of earthquakes in the crust of the PHS slab. In the area from Tokai to west Shikoku, where the younger PHS slab (15–27 Ma; Okino et al., 1994, 1999) is subducting, seismicity in the oceanic crust persists down to ca. 40 km. On the contrary, in Kanto and south Kyushu, where the older PHS slab (>48 Ma for Kanto and >50 Ma for south Kyushu; Seno and Maruyama, 1984; Tokuyama, 1995; Iwamori, 2007) is subducting, intraslab earthquakes in the crust occur as deep as ca. 90 km (Kanto) and ca. 180 km (south Kyushu).

A pronounced seismic belt, similar to the upper-plane seismic belt in the PAC slab, is clearly seen nearly parallel to 60-km iso-depth contour of the PHS slab beneath south Kyushu as indicated by a thick arrow. This seismic belt suddenly disappears at a location beneath central Kyushu, at which the deeper extension of the Kyushu–Palau ridge intersects (bold broken line in Fig. 16). This means that the seismic belt formed at depths of ca. 80 km in the crust of the older PHS slab (>60 Ma) beneath south Kyushu does not continue to the crust of the younger PHS slab (ca. 27 Ma) to the north, which strongly suggests that the formation of the seismic belt mainly depends on slab age, or

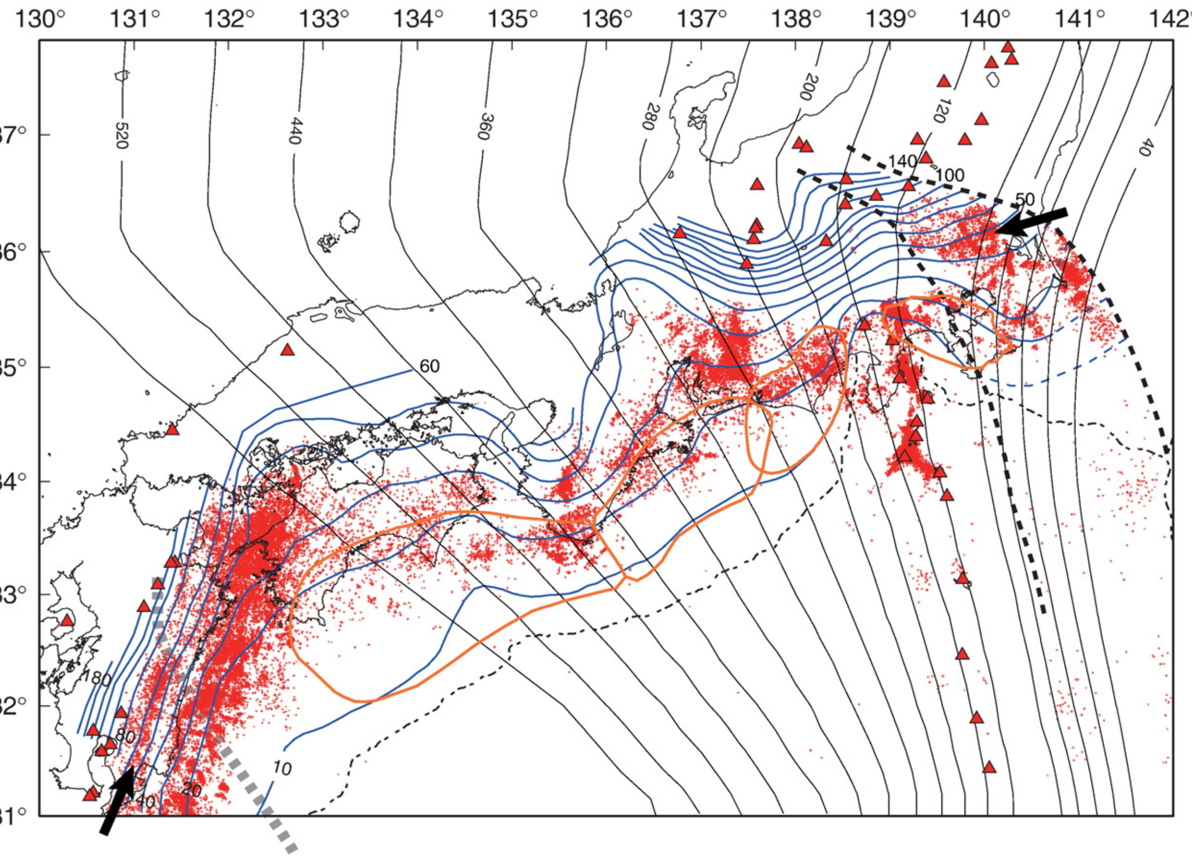

Fig. 16. Epicenter distribution for earthquakes in the crust of the PHS slab. Earthquakes 0–10 km below the upper plate interface are plotted. Estimated location of the subducted Kyushu–Palau ridge is denoted by a bold broken line. Others are the same as in Fig. 4.

temperature in the slab, similar to the case of the upper-plane seismic belt in the PAC slab. Such a clear seismic belt nearly parallel to the iso-depth contours of the upper surface of the PHS slab is not seen for other areas, except Kanto where intraslab earthquakes in the crust have a tendency to concentrate between 40-km and 60-km iso-depth contours as indicated by a thick arrow.

We infer this regional variation is due to the difference in age (or temperature) of the PHS slab. The seismic belt in the crust of the older PHS slab in south Kyushu is formed at depths ca. 20 km shallower than that in the crust of the PAC slab in Hokkaido and Tohoku, and is probably caused by the same dehydration reaction associated with phase transformation of the crustal material. A similar seismic belt might be formed in the crust of the PHS slab in Kanto at depths ca. 35 km shallower than that in Hokkaido and Tohoku. Non existence of such a clear seismic belt in the crust of the younger PHS slab in the area from Tokai to west Shikoku might be due to its warmer temperature condition, which causes a very shallow-angle subduction and therefore a geotherm of the slab crust encountering with several facies boundaries at depths of 10–40 km. This perhaps yields intraslab seismicity more broadly distributed at those depths in the slab crust as actually seen in the area from Tokai to Shikoku. More detailed investigations including detailed seismic velocity structure and temperature distribution in the crust of the PHS slab are necessary for understanding of the dehydration process there.

4.3. Low-velocity slab crust

The seismic velocity of hydrated slab crust is lower than that of the surrounding mantle, and thin low-velocity layers at the top of subducting plates have been detected in several subduction zones (e.g., Fukao et al., 1983; Matsuzawa et al., 1986; Bostock et al., 2002; Abers, 2005; Shiomi et al., 2006; Kawakatsu and Watada, 2007; Rondenay et al., 2008). The seismic velocity of the slab crust is expected to increase with the progress of dehydration reactions, which release H_2O from the subducting material. This increase in seismic velocity has been detected in a recent seismic tomographic study by Tsuji et al. (2008), who estimated the detailed seismic velocity structure in the vicinity of the subducting PAC slab in the Tohoku region by the double-difference tomography method of Zhang and Thurber (2003). The cross-arc vertical cross-section of S-wave velocity thus obtained is shown in Fig. 17. A distinct S-wave low-velocity zone with a thickness of approximately 10 km is resolved at the top of the PAC slab. This zone also has slightly low P-wave velocities and high Vp/Vs (Tsuji et al., 2008). This low-velocity zone gradually becomes less distinct and finally disappears at depths of 70–90 km, suggesting that dehydration reactions become intense at these depths. This result is consistent with the concentration of earthquakes in the slab crust in the Tohoku region, which forms an upper-plane seismic belt at these depths. This observation is also consistent with the prediction of the dehydration embrittlement hypothesis.

Nakajima et al. (2009) reconstructed a detailed seismic velocity structure for the vicinity of the subducting PAC slab over a region that includes both Tohoku and Kanto. A distinct low-seismic velocity zone corresponding to the uppermost part of the subducting PAC slab (i.e., slab crust) was resolved in that study over the entire region, from Tohoku to Kanto. Fig. 18 shows the spatial distribution of S-wave velocities along a curved plane 5 km below the upper surface of the PAC slab, cutting through the middle of the slab crust. A distinct systematic pattern can be seen in the spatial variation of S-wave velocity within the slab crust. In the Tohoku region, the S-wave velocity remains low down to a depth of 70–90 km, corresponding to the upper-plane seismic belt, below which the velocity is considerably

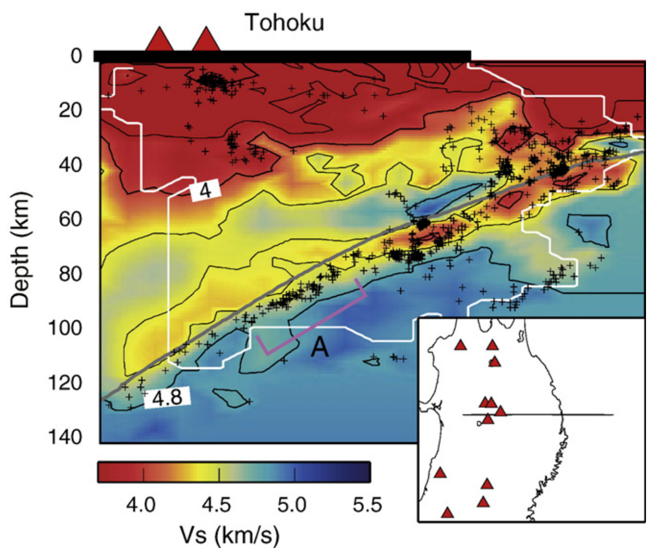

Fig. 17. Cross-arc vertical cross-section of S-wave velocity in central Tohoku (Tsuji et al., 2008). S-wave velocity is shown by color scale at the bottom. Solid line shows the upper surface of the PAC slab. Upper-plane seismic belt is denoted by A.

higher. This velocity structure, which is readily observable throughout Tohoku (see also Fig. 17), is not apparent in the Kanto region. As described above, the PAC slab crust is in contact with the overlying cold PHS slab in the Kanto region, delaying the eclogite-forming phase transformations. The seismic velocity in the slab crust in Kanto should therefore remain low to greater depth than in Tohoku. This can be seen in Fig. 18, where the region of low S-wave velocity in the slab crust beneath Kanto extends to the depths at which the obliquely trending upper-plane seismic belt is formed (100–140 km). This peculiar distribution of the depth limit of low S-wave velocity is consistent with that expected from the prediction that the contact with the cold PHS slab inhibits heating by the mantle wedge and suppresses the temperature rise in the PAC slab crust in Kanto. These observations provide further strong evidence supporting the dehydration embrittlement hypothesis for the generation of intraslab earthquakes. The concentration of intraslab events at the upper-plane seismic belt and the increase in seismic velocity in the slab crust at the same point suggest that the rate of dehydration of the slab crust is particularly high at such depths.

5. Arc magmatism

5.1. Transport of water from slab crust to arc crust and magma generation in Tohoku

Water released by dehydration reactions in the subducting slab migrates into the mantle wedge above. Recent high-resolution seismic observations have provided information on the path of H_2O transport from slab crust to arc crust via the mantle wedge, and the contribution of liberated H_2O to magma generation.

As described in the previous section, intense dehydration reactions are expected to be active in the PAC slab crust in the Tohoku region at depths of 70–90 km, probably corresponding to the JLB–LAE phase transformation of mid-ocean ridge basalt (MORB). Water liberated from the slab crust by these dehydration reactions is expected to migrate upward into the overlying mantle wedge, where it will react with mantle materials to form hydrated minerals such as serpentine, chlorite, and amphibole (e.g., Davies and Stevenson, 1992; Iwamori, 1998). In subduction zones involving old oceanic plates, such as that in Tohoku, the hydrated layer formed immediately above the subducting slab is presumed to be dragged downward by the subducting slab and to thus be transported to a depth where further dehydration can occur (Iwamori, 1998; Schmidt and Poli, 1998; Maruyama and Okamoto, 2007). The depth to which this hydrated layer is transported is dependent on the thermal structure of the mantle wedge. Iwamori (1998) estimated the maximum depth to be 150–200 km in the Tohoku region based on numerical simulations of a plate subduction model. This hydrated layer, lying immediately above the subducting slab in Tohoku, has been detected as a low-velocity layer in recent receiver function analysis and double-difference tomography studies (Kawakatsu and Watada, 2007; Tsuji et al., 2008). Kawakatsu and Watada (2007) reconstructed the structure of the crust and upper mantle in Tohoku by receiver function analysis, and successfully resolved the upper surface of the subducting PAC slab and the slab Moho as distinct velocity discontinuities. A prominent low-velocity layer immediately above the subducting slab was also detected at depths of 70–130 km, and was interpreted to correspond to a hydrated layer containing serpentine and chlorite (Fig. 19). Seismic imaging by Tsuji et al. (2008) using the double-difference tomography method (Zhang and Thurber, 2003), performed at locally higher spatial resolution in the region of earthquake hypocenter concentration, also resolved this hydrated layer as a distinct low-velocity layer in the mantle wedge extending to depths of 70–120 km, immediately overlying the subducting PAC slab (see Fig. 17). These observations

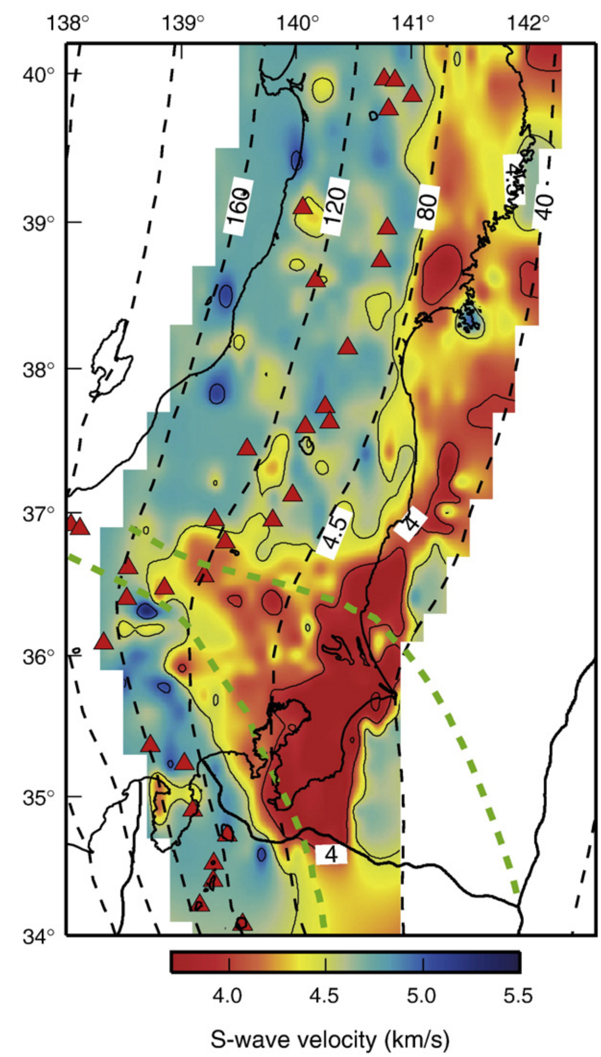

Fig. 18. S-wave velocity distribution in the PAC slab crust on a curved plane 5 km below the upper surface of the PAC slab (Nakajima et al., 2009). Broken curves denote iso-depth contours of the upper surface of the PAC slab. Slab contact zone is enclosed by green broken curves.

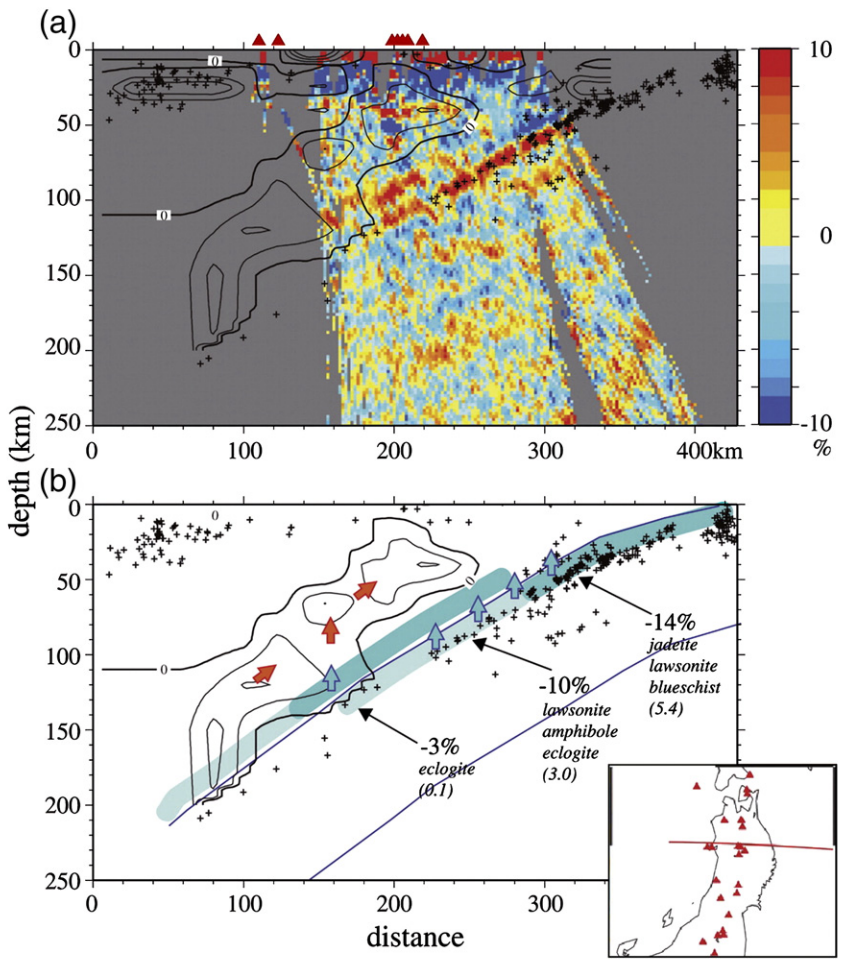

Fig. 19. Cross-arc vertical cross-section of reflectivity beneath NE Japan and a schematic model for water transportation (Kawakatsu and Watada, 2007). (a)Dip-corrected reflectivity image. Reflectivities, measures of local S-wave velocity jumps, are shown by the color scale. Red and blue correspond to velocity increase and decrease with depth, respectively. (b) Inferred pathway of water transportation. Light blue and red colors show nonmagmatic and magmatic paths, respectively. S-wave tomographic image in the mantle wedge above the slab by Nakajima et al. (2001) is also shown by contours with 3% velocity-change intervals. Plus signs show earthquakes with magnitude larger than 3.

offer strong evidence of the transport of water liberated from the hydrated slab crust at shallow depth to the deep mantle in a cold subduction zone environment such as that in Tohoku (Kawakatsu and Watada, 2007). The depth limit of the low-velocity layer detected by seismic observations is 120–130 km, whereas that estimated by numerical simulations (Iwamori, 1998) is 150–200 km. Further investigations are therefore required in order to clarify the depth limit of the hydrated layer, since the resolution of previous studies has been insufficient to resolve such a thin low-velocity layer immediately above the slab at depths greater than 120–130 km.

The hydrated mantle layer containing serpentine and chlorite is in direct contact with the underlying slab, and is transported to depth by entrainment with the subducting slab. This entrainment of the hydrated layer of the mantle with slab subduction results in the migration of mantle material on the back arc side to fill the enclosed space (McKenzie, 1969), thus forming mechanically induced secondary convections in the mantle wedge associated with slab subduction. Seismic tomography studies have also resolved this return flow portion of secondary convection as a distinct zone of low seismic velocity and high attenuation in the mantle wedge of the Tohoku subduction zone (Hasegawa et al., 1991; Zhao et al., 1992; Zhao and Hasegawa, 1993; Zhao et al., 1994; Tsumura et al., 2000; Nakajima et al., 2001; Hasegawa and Nakajima, 2004). Fig. 20 shows a cross-arc vertical cross-section of S-wave velocity perturbations in central Tohoku. A prominent zone of low S-wave velocity can be seen in the mantle wedge, inclined subparallel to the subducting slab and extending from a depth of approximately 100 km to the Moho immediately beneath the volcanic front. This low S-wave velocity zone has been detected throughout the Tohoku region, indicating the existence of a single sheet-like low-velocity zone at depths of 30–150 km in the mantle wedge, inclined nearly parallel to the subducting slab (Nakajima et al., 2001; Hasegawa and Nakajima, 2004). The zone also exhibits low P-wave velocities, high V_p/V_s values (Zhao et al., 1992; Nakajima et al., 2001), and low Q_p values (i.e., high seismic attenuation) (Tsumura et al., 2000). Due to the source–receiver geometry employed in tomographic inversion, the inclined low-velocity zones reported by Nakajima et al. (2001) and Hasegawa and Nakajima (2004) are only resolved down to depths of approximately 150 km. Zhao and Hasegawa (1993) and Zhao et al. (1994) reconstructed the velocity structure for the entire region of the Japan islands by tomographic inversion using both regional and teleseismic data, and thus revealed that there are no prominent inclined low-velocity zones in the mantle wedge in the Tohoku region at depths greater than 150 km.

This inclined sheet-like, low-velocity, high-attenuation zone is presumed to correspond to the upwelling flow portion of the secondary convection accompanying slab subduction. The hydrated layer directly above the slab, corresponding to the downflow, is estimated to undergo dehydration decomposition at a depth of 150–200 km in old plate subduction zones such as that in Tohoku (Iwamori, 1998). The water thus liberated from the hydrated layer immediately above the slab migrates upward, and is expected to encounter the

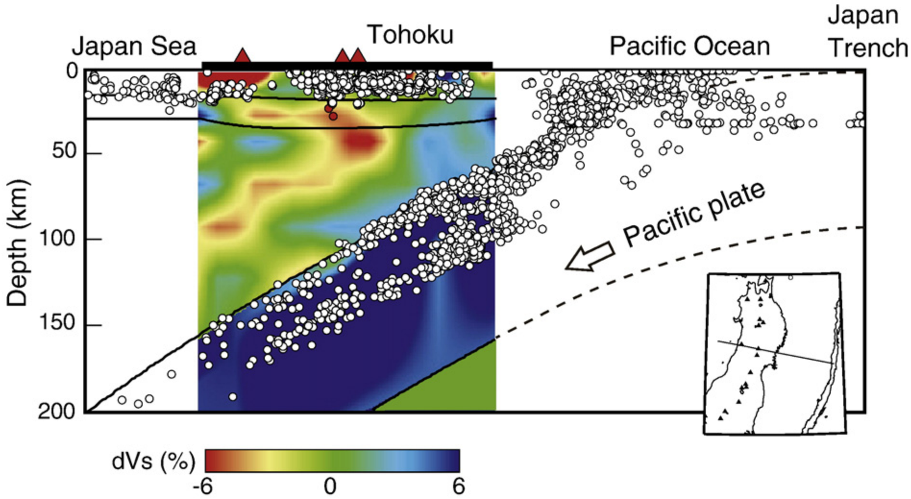

Fig. 20. Cross-arc vertical cross-section of S-wave velocity perturbations in central Tohoku (Nakajima et al., 2001). Red triangles denote active volcanoes, solid black bar denotes land area, open circles denotes earthquakes, red circles denote deep low-frequency earthquakes, solid lines denote the Conrad and Moho of the upper plate and upper and lower surfaces of the PAC slab from top to bottom.

inclined upwelling flow at depths of 100–150 km. Such a supply of water into the upwelling flow is expected to lower the solidus temperature. The temperature of the upwelling flow, estimated by Nakajima and Hasegawa (2003a) by comparison of the seismic attenuation structure with laboratory experiment data, is higher than the wet solidus of peridotite, suggesting the occurrence of partial melting. Nakajima et al. (2005) reported the existence of melt-filled pores with aspect ratios of 0.10–0.01 and melt volume fractions of 0.1 to several percent within this upwelling flow based on a comparison of observed fall-off rates of P-wave and S-wave velocities with a diagram of Takei (2002) showing the relative role of liquid compressibility and pore geometry in determination of the V_p/V_s ratio.

The inferred paths of H_2O transport in Tohoku based on seismic observations are shown schematically in Fig. 21(a). The migration of H_2O into the hot mantle material from below causes partial melting with a volume fraction in the range of 0.1 to several percent within the upwelling flow. Melt is formed both by decompression melting and by melting due to the addition of water. The observation that the inclined low-velocity zone can only be clearly resolved at depths shallower than 150 km (Zhao and Hasegawa, 1993; Zhao et al., 1994) suggests that melting due to the addition of water plays an important role in magma generation in this region. Water originating from the subducting slab is thus eventually incorporated into the melt within the upwelling flow, which accumulates immediately below the arc Moho along the volcanic front. The volume fraction of the accumulated melt, estimated from the fall-off rates of P-wave and S-wave velocities, is approximately 1% (Nakajima et al., 2005). The melt then rises, penetrating the crust, and may ultimately reach the surface to form volcanoes. The volcanic front is thus inferred to be formed at locations where the sheet-like inclined upwelling flow in the mantle wedge reaches the arc Moho (Fig. 21).

Seismic tomography has also provided important information on the variability of magma generation along the Tohoku arc, as well as on the mechanism of back arc volcano formation. Quaternary volcanoes in Tohoku are grouped into 10 volcanic clusters distributed in a long and narrow band oriented perpendicular to the arc (Kondo et al., 1998; Tamura et al., 2002). The clusters extend for approximately 50 km, separated along the arc by volcanic gaps of 30–75 km. Tamura et al. (2002) found that this volcanic clustering is spatially correlated with topography highs in the back arc region, low-velocity zones along the volcanic front beneath the arc Moho, and local negative Bouguer gravity anomalies along the Japan Sea coast. Based on these observations, the clustering of Quaternary volcanoes was

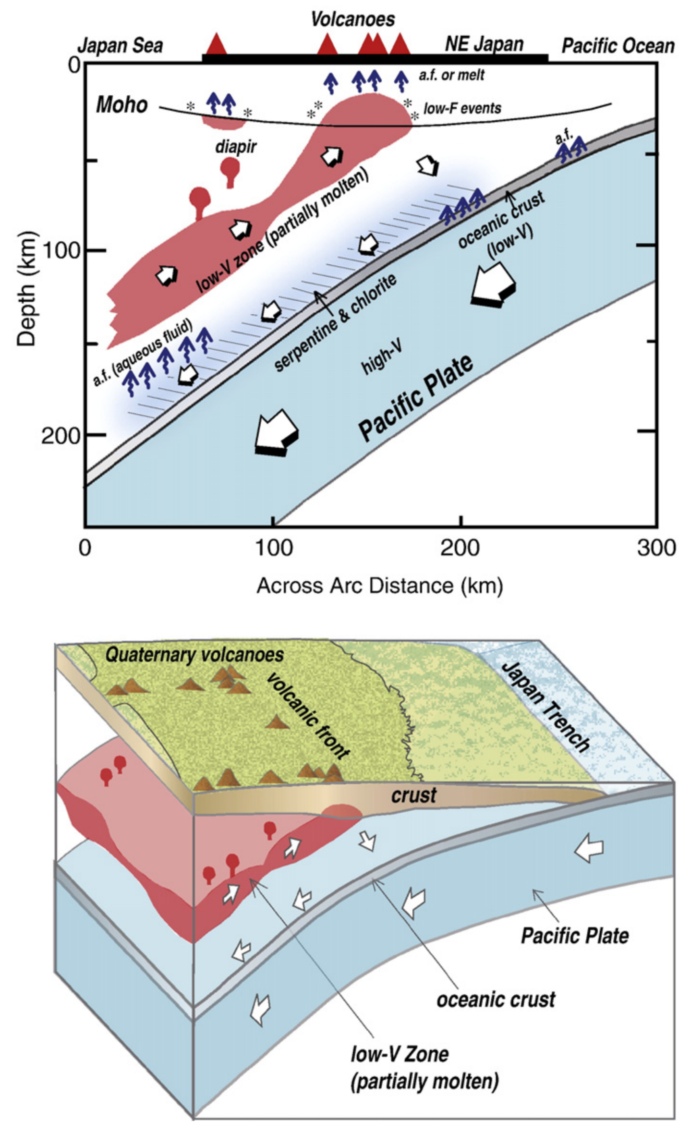

Fig. 21. Schematic illustration showing the inferred pathway of H_2O transportation (Hasegawa and Nakajima, 2004). (a) Cross-arc vertical cross-section of crust and upper mantle in Tohoku. (b) Three-dimensional expression.

inferred to be caused by the local development of hot regions in the mantle wedge in the form of inclined, ca. 50 km-wide fingers extending from a depth of 150 km on the back arc side to 50 km beneath the volcanic front. The repeated supply of magma from these hot fingers in the mantle wedge into the overlying arc crust is thus inferred to cause uplift and volcano formation. Numerical simulations using 3D models of plate subduction with/without pressure and temperature-dependent viscosity by Honda et al. (2002) and Honda and Yoshida (2005) showed the existence of roll-like small-scale convection in the mantle wedge. They suggested that this small-scale convection caused the along-arc variation of magmatism in the Tohoku arc.

More distinct low-velocity zones in the mantle wedge exhibiting good spatial correlation with volcanic clusters were resolved in a seismic tomographic study by Hasegawa and Nakajima (2004). Tomographic images constructed for the inclined low-velocity zone in the mantle wedge reveal distinct along-arc variations in this zone, with areas of particularly low velocity occurring periodically every ca. 80 km along the strike of the arc. These very-low-velocity areas of the inclined low-velocity zone at depths of 30–150 km in the mantle wedge exhibit good spatial correlations with the topographic highs between the Ou backbone range and the back arc side hosting the volcanic clusters. Based on these observations, Hasegawa and Nakajima (2004) proposed a structure for the crust and upper mantle and the upwelling flow in the mantle wedge for Tohoku, as shown schematically in Fig. 21(b). This figure extends the two-dimensional cross-section shown in Fig. 21(a) to three dimensions. The upwelling flow, resolved as an inclined low-velocity zone in the mantle wedge, is sheet-like with locally varying thickness in the along-arc direction. The volcanic front emerges where the upwelling flow reaches the arc Moho. In regions characterized by local thickening of the sheet-like upwelling flow and by large volume fractions of melt, some of the melt may separate from the inclined upwelling flow before reaching the arc Moho, rising vertically from the point of separation in the form of a plume and accumulating beneath the arc Moho. Some of this melt rises further and penetrates into the arc crust, forming volcanoes and causing crustal uplift. Hasegawa and Nakajima (2004) inferred that the clusters of Quaternary volcanoes and the elevations of topography extending from the volcanic front toward the back arc region are formed in this manner, and postulated that this formation process continues even at present.

5.2. Mantle upwelling flow

The hydrated layer immediately above the subducting slab and the inclined upwelling flow in the mantle wedge, which are mechanically induced by slab subduction and which are observable in Tohoku as low-velocity and high-attenuation zones, are considered to be common to all subduction zones. Inclined low-velocity zones oriented subparallel to the subducting slab, probably corresponding to the upwelling flow similar to that in Tohoku, have been detected in the mantle wedge of several other subduction zones, including those in Kamchatka (Gorbatov et al., 1999), Tonga (Zhao et al., 1997a,b), eastern Aleutians (Abers, 1994), Alaska (Zhao et al., 1995), Hokkaido (Wang and Zhao, 2005), and Kyushu (Abdelwahed and Zhao, 2007). Inclined high-attenuation zones similar to that in Tohoku also have been detected in the mantle wedge of Kanto–Tokai (Sekiguchi, 1991), New Zealand (Satake and Hashida, 1989), Tonga–Fiji (Roth et al., 1999) and northern Chile (Haberland and Rietbrock, 2001). Although such inclined low-velocity and high-attenuation zones in the mantle wedge have not been detected in a number of other subduction zones exhibiting arc volcanism, it is probable that the resolution of analyses conducted in those regions was insufficient to resolve the low-velocity zones at the appropriate depth beneath the back arc region. The low-velocity layer immediately above the subducting slab, corresponding to the downgoing flow, has not been detected in subduction zones other than that in Tohoku. This suggests that the detection of such a thin layer at these depths requires very high resolution, which is achieved in the Tohoku region due to the high density of the seismic network deployed in that area.

The idea that arc magmatism is caused by the upwelling flow induced by slab subduction and by the direct addition of water released from the slab also seems to explain the structure of subduction zones without arc volcanism. In subduction zones involving young and warm slabs, the dehydration decomposition of serpentine and chlorite in the hydrated layer immediately overlying the subducting slab will occur at shallower depths (e.g., Iwamori, 1998). If the liberation of water by this process occurs at depths too shallow to meet the upwelling flow, partial melting may not occur, resulting in weak or absent arc volcanism (Kirby et al., 1996). The geometry of the subducting slab may also affect the emergence of arc volcanism, since the upwelling flow in the mantle wedge is presumed to be formed nearly parallel to the slab. For example, the horizontal orientation of the Nazca slab beneath central Peru and central Chile (Hasegawa and Sacks, 1981; Cahill and Isacks, 1992) probably inhibits the upwelling flow that is mechanically induced by slab subduction, with the result that no active volcanoes are generated in this region.

In Hokkaido, where the PAC slab subducts obliquely to the trench axis, active volcanoes are distributed along the volcanic front running subparallel to the iso-depth contours of the PAC slab surface. Low-velocity zones inclined nearly parallel to the subducting slab, similar to

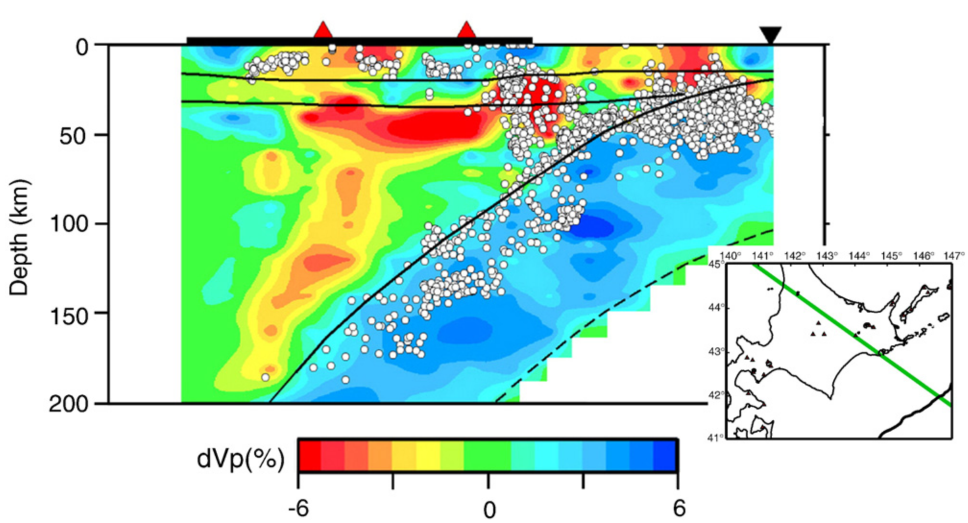

Fig. 22. Cross-arc vertical cross-section of P-wave velocity perturbations in Hokkaido (Wang and Zhao, 2005). Others are the same as in Fig. 20.

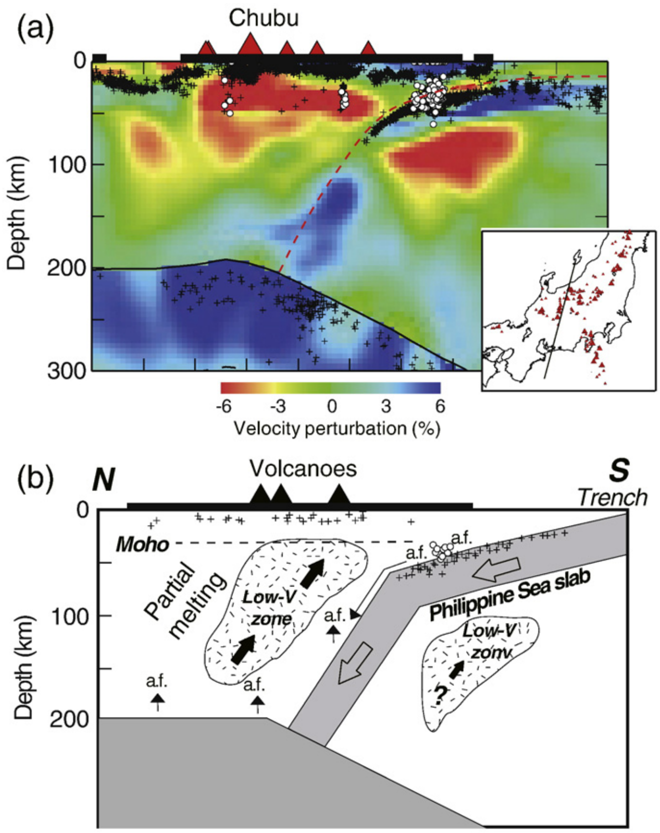

Fig. 23. (a) Vertical cross-section of S-wave velocity perturbations along the line shown in the inset map, and (b) schematic vertical cross-section of slab and mantle structure in Chubu (Nakajima and Hasegawa, 2007a). Crosses denote earthquakes and open circles denote deep low-frequency earthquakes. Others are the same as in Fig. 20.

those beneath Tohoku, have been clearly resolved in a seismic tomographic study by Wang and Zhao (2005), as shown in Fig. 22. The seismic anisotropy structure estimated from shear-wave splitting analysis provides additional information on the pattern of the subduction-induced flow in the mantle wedge. The shear-wave splitting observed in Tohoku and Hokkaido represents a systematic spatial variation in the directions of fast shear waves, indicating different features of anisotropy between the fore arc and back arc regions (Okada et al., 1995; Nakajima and Hasegawa, 2004; Nakajima et al., 2006). The fast direction is observed to be trench-parallel in the fore arc regions of both Tohoku and Hokkaido, consistent with the anisotropy expected from the deformation of the B-type olivine fabric and with that within the subducted slab. In contrast, the fast directions in the back arc regions differ between Tohoku and Hokkaido. In Tohoku, the observed fast direction is trench-perpendicular, which can be explained by the lattice-preferred orientation of minerals caused by flow-induced strain in the mantle wedge (Ribe, 1992; Zhang and Karato, 1995; Tommasi, 1998; Kneller et al., 2007). However, the fast direction observed in the back arc region of Hokkaido is not trench-perpendicular, instead being oriented nearly parallel to the dip direction of the subducting PAC slab. This suggests that the direction of the return flow in the mantle wedge is governed by the local slab geometry, where materials within the upwelling flow in the mantle wedge move in the direction parallel to the maximum dip of the subducting slab, and not in the plate convergence direction (Nakajima et al., 2006).

Receiver function analyses in central Japan by Tonegawa et al. (2008) detected a thin hydrous layer immediately above the subducting PAC slab down to a depth of ca. 500 km, similar to the case of NE Japan detected by Kawakatsu and Watada (2007) but to much greater depths. This observation indicates that water is transported downward to the mantle transition zone with the slab subduction. This deep water subduction probably contributes to generate magma beneath southwestern Japan. In Chubu, central Japan, the volcanic front deflects toward the back arc side, in contrast with the fronts in Hokkaido, Tohoku, and Izu–Bonin, which run nearly parallel to the iso-depth contours of the slab surface. The volcanic chain in northern Chubu emerges on a line corresponding to a depth of 150–300 km to the upper surface of the PAC slab (see Fig. 3). Iwamori (2000) investigated this westward deflection of the volcanic front through two-dimensional numerical modeling of plate subduction, and revealed that heating of the PAC slab by the overlying hot mantle wedge is delayed due to contact with the overlying PHS plate, which shifts the dehydration reactions in the PAC slab, and hence magmatism, to greater depth. The deflection of the volcanic front toward the back arc side in Chubu can be readily understood from the detailed geometry of the contact between the two slabs in Kanto (see Fig. 4). In the Kanto region, the PHS slab subducts at a shallow dip in the south and then with steeper dip in the volcanic area of Chubu. Nakajima and Hasegawa (2007a) showed that an aseismic portion of this PHS slab subducts to a depth of at least 200 km in the volcanic area of Chubu, as shown in Fig. 23, whereas no volcanism occurs in the Kinki region where the slab subducts subhorizontally as far as the Japan Sea. On the basis of this observation, the geometry of the PHS slab is inferred to have a strong influence on arc magmatism, in addition to the supply of water from the underlying PAC slab (Nakamura et al., 2008).

A distinct volcanic front is formed in association with the subduction of the PHS slab in Kyushu, where the slab subducts at a steep dip down to a depth of approximately 200 km (Fig. 4). A seismic tomography study by Abdelwahed and Zhao (2007) indicated the existence of a low-velocity zone inclined subparallel to the subducting slab in the mantle wedge, as shown in Fig. 24. This low-velocity zone is similar to that observed in Tohoku, suggesting that the volcanism in Kyushu is caused by the inclined upwelling flow induced by slab subduction as in Tohoku. Fig. 24 also shows that the subducting PHS slab is aseismic down to a depth of approximately 500 km.

In southwest Japan, Quaternary volcanoes are distributed along the Japan Sea coast near Chugoku (Fig. 4), although the volcanic front is not well defined in this region in contrast to Kyushu. It is known that diverse volcanism occurred in the Chugoku region in the Neogene (Kimura et al., 2003). Here, the PHS slab subducts subhorizontally, which differs from the case of Kyushu. Primary source of magmas for the Quaternary volcanoes in this region is not well known. Iwamori (1992) showed the magmatism here originates from the mantle with a standard potential temperature but with abundant volatile components. Kimura et al. (2005) proposed that slab melting is one of likely candidates for the magmatism in this region. Seismic tomographic studies by Nakajima and Hasegawa (2007b) revealed a large low-velocity anomaly with lateral extent of approximately 300 km and depth extent of about 200 km in the upper mantle below the PHS slab in Chugoku and Shikoku. This anomaly extends to the uppermost mantle immediately beneath the Quaternary volcanoes, passing through the leading edge of the subducting PHS slab (Fig. 25). Nakajima and Hasegawa (2007b) interpreted this low-velocity zone to be an upwelling flow from depth, acting as one of primary sources of magmas for the Quaternary volcanoes in Chugoku.

6. Inland crustal earthquakes

6.1. High strain rate zones estimated from GPS data

Recent investigations have made great progress in clarifying the stress concentration mechanism that leads to interplate earthquakes, as described in Section 3. Aseismic slip in the stably sliding areas surrounding isolated asperities distributed on the plate boundary causes stress accumulation at the asperities. This incremental process eventually culminates in sudden slip at asperities, producing earthquakes. This asperity model is plausible for interplate earthquakes

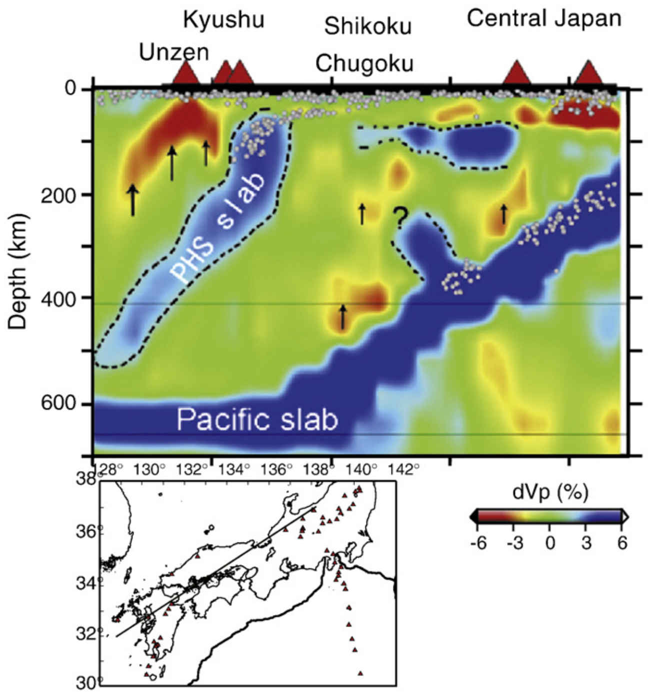

Fig. 24. Vertical cross-section of P-wave velocity perturbations along the line shown in the inset map (Abdelwahed and Zhao, 2007). Others are the same as in Fig. 20.

along the subducting plate boundary in the Pacific Ocean (Nagai et al., 2001; Matsuzawa et al., 2002; Yamanaka and Kikuchi, 2003, 2004; Yagi and Kikuchi, 2003; Yagi et al., 2003; Matsuzawa et al., 2004: Okada et al., 2006a,b; Hasegawa et al., 2007b; Shimamura et al., 2008). However, many large earthquakes also occur in the upper crust of terrestrial Japan, causing severe damage in many instances due to the shallow hypocenter and proximity to inhabited areas. The stress concentration mechanism that leads to these inland crustal earthquakes has yet to be clarified, although the local concentration of arc crust deformation is expected to play an major role in stress accumulation, as pointed out by Iio et al. (2004) and by Hasegawa et al. (2005).

Data acquired by the dense nationwide GPS network have made it possible to resolve in fine detail the pattern of coseismic and postseismic deformation associated with large earthquakes, episodic slow slip events along the subducting plate boundaries, and secular deformation of the Japan islands. The spatial distribution of strain rate estimated from these GPS data reveals that most regions characterized by high strain rates are associated with interplate coupling at the plate boundary and volcanic activity. However, high strain rate regions not correlatable to interplate coupling or volcanic activity have also been identified.

Through an analysis of dense GPS network data, Sagiya et al. (2000) detected a zone of high strain rate extending from Niigata to Kobe along the Japan Sea coast and in northern Chubu and Kinki. This zone of high strain rate is approximately 100 km wide and extends along this line for up to 500 km in a NE–SW orientation, as shown in Fig. 26, and has been termed the Niigata–Kobe Tectonic Zone (NKTZ). The existence of this zone has also been pointed out based on an analysis of triangulation data for the past 100 years (Hashimoto, 1990; Hashimoto and Jackson, 1993). The NKTZ undergoes contraction in the WNW–ESE direction at a rate of approximately 10^{-7}/yr, orders of magnitude greater than that of the surrounding areas. Historically, many large earthquakes have occurred along this tectonic zone, suggesting that the present deformation field represents tectonic deformation that has persisted for at least the last several hundred years (Sagiya et al., 2000).

A zone of high strain rate has also been identified along the Ou backbone range in Tohoku through analysis of dense GPS network data (Sato et al., 2002; Miura et al., 2004). Fig. 27 shows the spatial distribution of E–W components of horizontal strain rates. A high strain rate zone can be recognized passing through the middle of Tohoku in a N–S orientation. The figure also reveals the existence of another high strain rate zone along the Japan Sea coast, which corresponds to the northern extension of the NKTZ. The high strain rate zone along the Ou backbone range is approximately 400 km long and 30–40 km wide, and undergoes contraction in the E–W direction

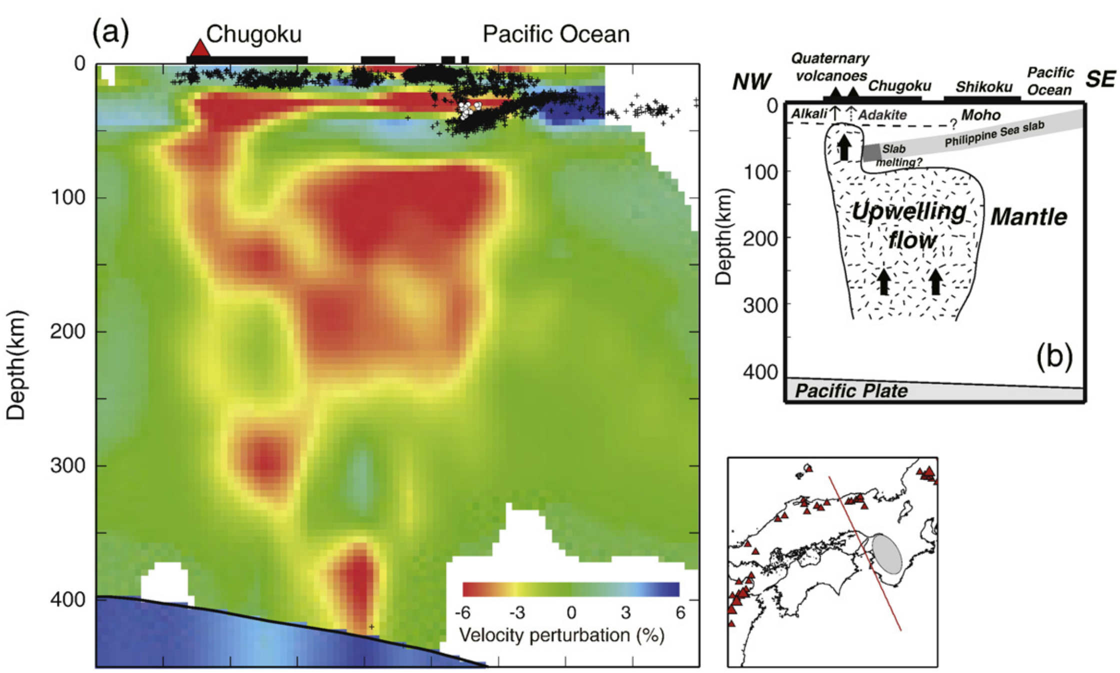

Fig. 25. (a) Cross-arc vertical cross-section of S-wave velocity perturbations along the line shown in the inset map, and (b) schematic illustration of slab and upper mantle structure in Chugoku and Shikoku (Nakajima and Hasegawa, 2007b). Others are the same as in Fig. 23.

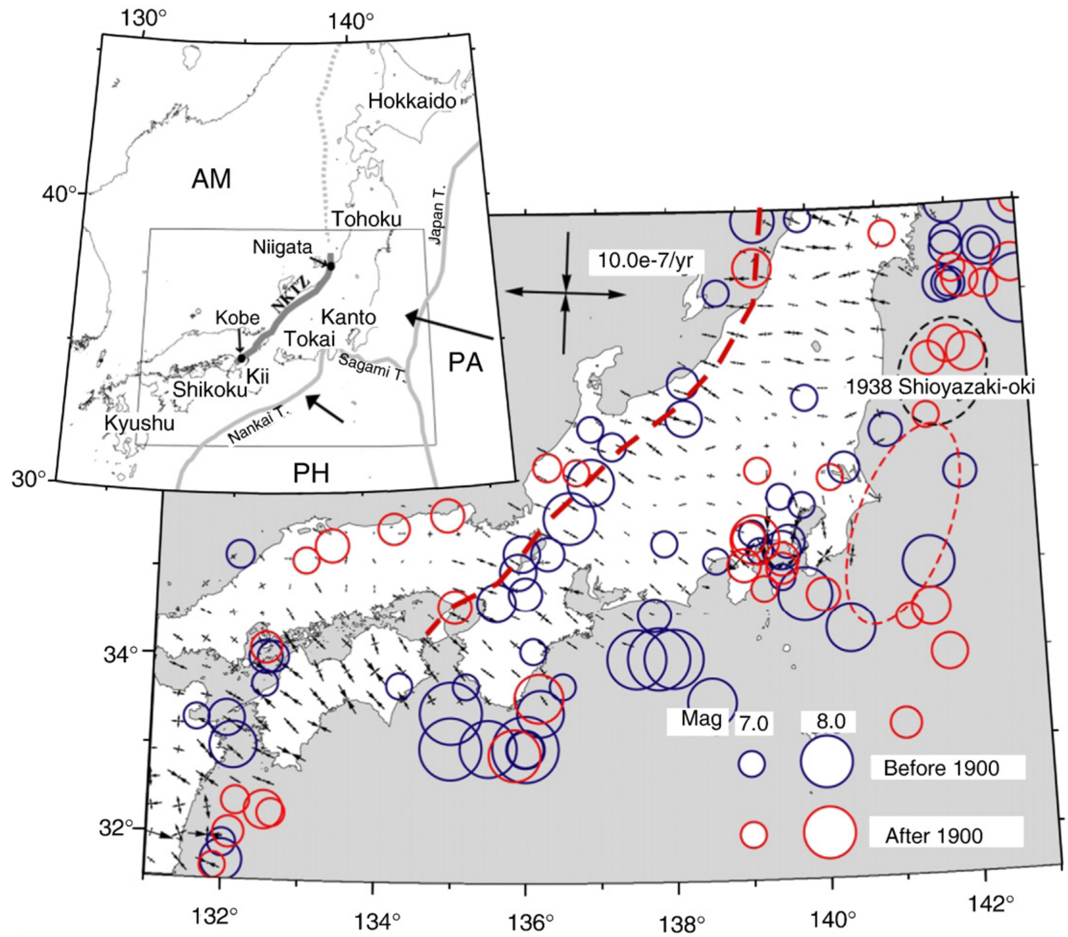

Fig. 26. Distribution of principal strain axes of horizontal strain rates estimated from GPS data for the period February 1996 to June 1999, and epicenters of historical large shallow earthquakes (Hyodo and Hirahara, 2003). Solid broken line and solid gray line in the inset map denote the NKTZ.

(the plate convergence direction in this region). This high strain rate zone along the backbone range coincides with a concentration of microearthquake activity (Fig. 27), suggesting that the microearthquake activity is caused by the present concentration of crustal deformation. It is known that active faults are distributed along the eastern and western edges of the backbone range (Active Fault Research Group, 1991), corresponding to both edges of this high strain rate zone.

6.2. Concentrated deformation of arc crust and generation of inland crustal earthquakes

The local concentration of arc crust deformation, as indicated by the high strain rate zones, is inferred to be caused by aqueous fluid originally liberated from the subducting slab, and is considered to be responsible for large inland crustal earthquakes (Hasegawa et al., 1991, 2000; Iio and Kobayashi, 2002; Iio et al., 2002, 2004; Hasegawa et al., 2005). In Tohoku, the inclined sheet-like upwelling flow in the mantle wedge reaches the arc Moho along the volcanic front, corresponding to the Ou backbone range, as shown schematically in Fig. 21. The melt contained in this upwelling flow either underplates the arc crust or penetrates into the crust, resulting in a locally elevated geotherm along the backbone range. Some of the melt that penetrates into the crust may cool and partially solidify, expelling water. Water of slab origin may thus be supplied continuously into the crust along the Ou backbone range. The presence of water is expected to weaken the crustal material, causing local contractive deformation of the crust under the present compressive E–W stress field.

The deformation pattern of arc crust in Tohoku indicated by these observations is shown schematically in Fig. 28. Melt intruded into the arc crust heats the surrounding rocks, causing a local elevation of the brittle–ductile transition zone. Some of the melt cools and solidifies, expelling water that can move rapidly at lower crustal levels, causing anomalous deep low-frequency microearthquakes (Hasegawa et al., 1991; Hasegawa and Yamamoto, 1994). Bright S-wave reflectors have been widely detected at midcrustal levels along the backbone range (Matsumoto and Hasegawa, 1996; Hori et al., 2004), which may represent manifestations of water accumulated within thin horizontal reservoirs at intermediate crustal depths. Some of the water continues to rise and reaches the upper crust, which might cause plastic contractive deformation in some part of the brittle upper crust. The entire crust along the backbone range is therefore locally weak compared to the surrounding regions, causing locally concentrated contractive deformation and uplift along it under the current compressional stress field in the plate convergence direction. The upper crust, which is being compressed, deforms elastically, but some anelastic deformation may be involved in parts of the upper crust along the backbone range. This local contractive deformation along the backbone range causes stress concentration in the upper crust immediately above. This accumulation of stress will eventually lead to rupture of the entire upper crust, generating a large inland crustal earthquake that returns the deformation to spatial uniformity (Hasegawa et al., 2005). Such anelastic contractive deformation concentrated along the backbone range including the upper crust is likely to be responsible for the extremely high microearthquake activity in this region (see Fig. 27).

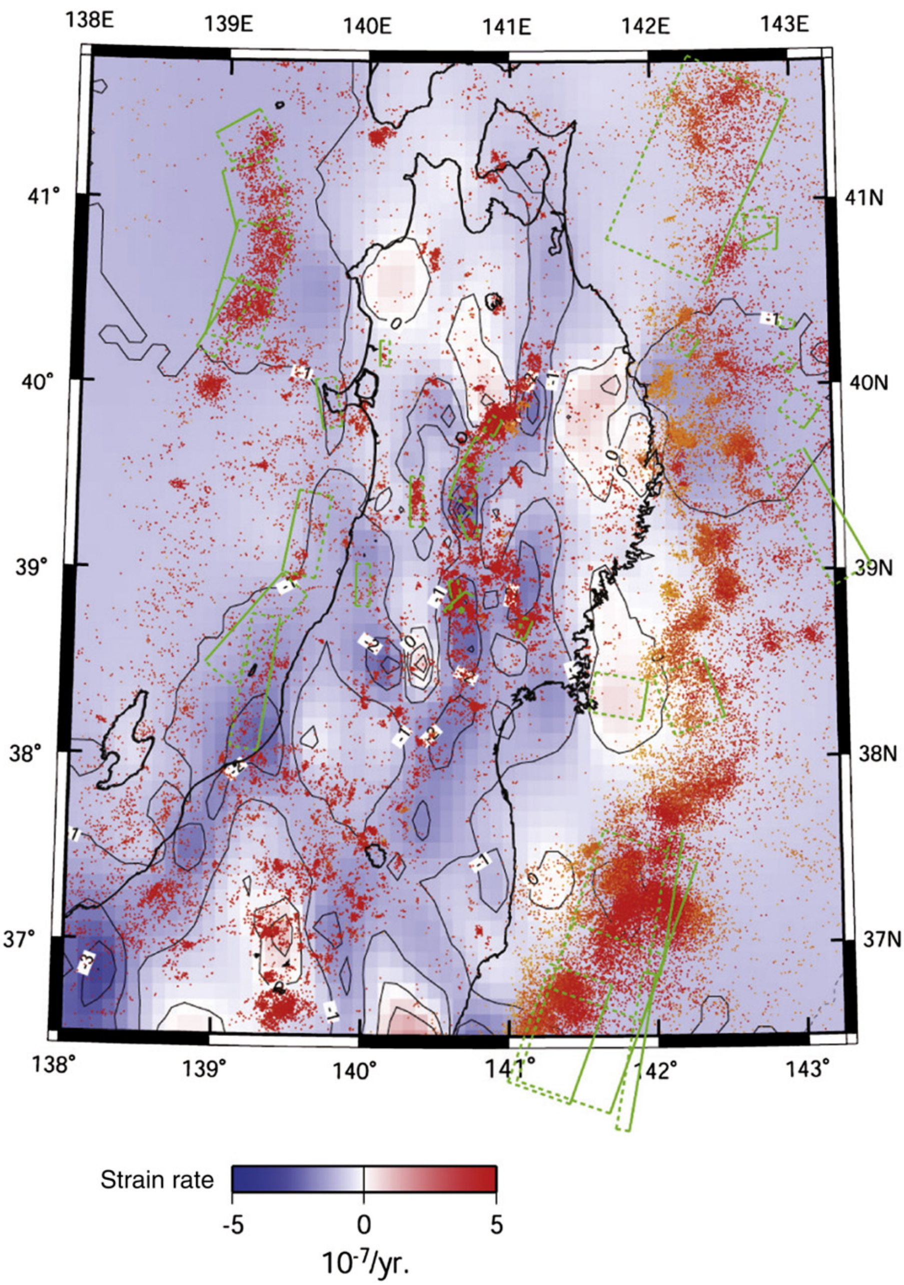

Fig. 27. Distribution of horizontal E–W strain rates estimated from GPS data for the period 1997–2001 (Sato et al., 2002). Red dots denote shallow microearthquakes located by Tohoku University seismic network for the same period, and green outlines denote fault planes of large earthquakes (Sato et al., 1997).

Numerical simulations by Shibazaki et al. (2007, in press) have verified the anelastic contractive deformation model for earthquake generation in the Tohoku region. The models considered in these studies are two-dimensional models with nonlinear viscous flow in the lower crust and faulting by Mohr–Coulomb plasticity in the upper crust, taking into consideration the effect of geothermal structures in the crust under a compressional stress field. The finite element analyses conducted on these models demonstrate that local contractive deformation due to nonlinear viscous flow occurs in the lower crust of the high-temperature region, resulting in shear faulting in the upper crust immediately above. This result indicates that the loading process for large inland crustal earthquakes originates from the non-uniformity of the thermal structure in the crust and uppermost mantle.

The origin of the NKTZ has been investigated by several authors. The high strain rates observed in the NKTZ have been discussed in terms of both interplate deformation (e.g., Shimazaki and Zhao, 2000) and intraplate deformation (e.g., Iio et al., 2002; Hyodo and Hirahara, 2003; Yamasaki and Seno, 2003; Iio et al., 2004). Shimazaki and Zhao (2000) characterized the NKTZ as a colliding plate boundary, and discussed the observed high strain rates along the zone in terms of interplate deformation. Iio et al. (2002, 2004) concluded that the zone is not a plate boundary, and instead proposed a model in which a weak zone with low viscosity exists in the lower crust immediately beneath the NKTZ, as shown schematically in Fig. 29. The weak zone was presumed to be attributable to an elevation of water content resulting from the migration of water liberated by dehydration of the subducting PHS slab. Enhanced fluid flux from the overlapping PHS and PAC slabs in this region (Nakamura et al., 2008) may promote it. Hyodo and Hirahara (2003) examined this model by numerical simulations using a viscoelastic finite element model, and showed that the observed high strain rates along the NKTZ can be satisfactorily explained by an underlying 15 km-thick viscoelastic lower crust with

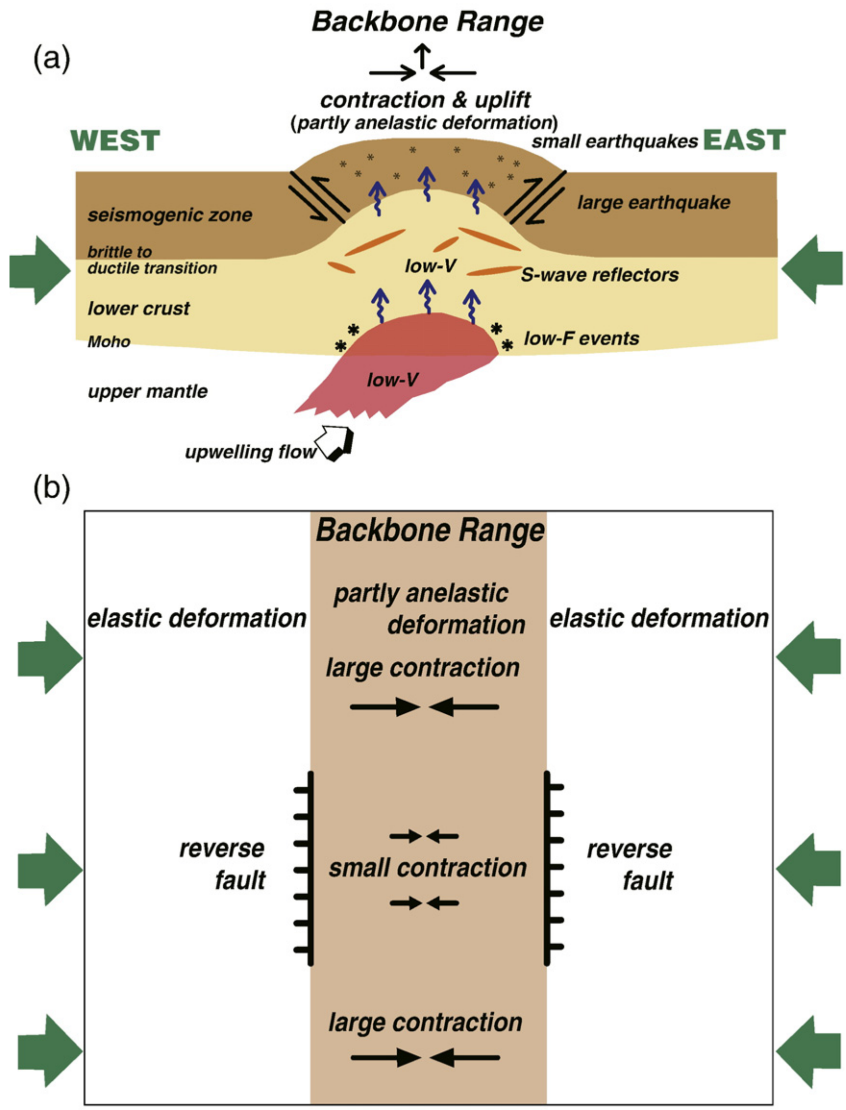

Fig. 28. Schematic illustration of pattern of arc crust deformation and characteristic shallow earthquake activity in Tohoku (Hasegawa et al., 2005). (a) Cross-arc vertical cross-section, and (b) map.

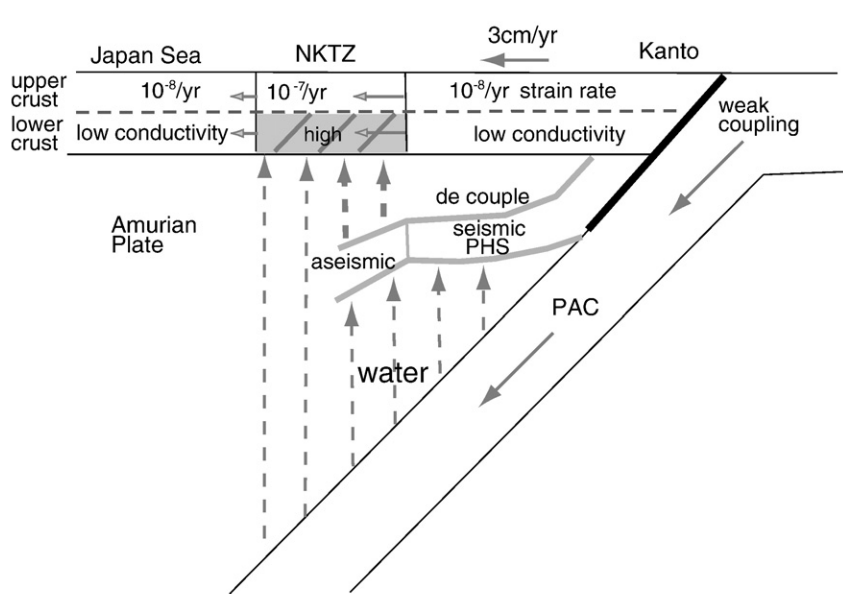

Fig. 29. Schematic illustration of cross-arc vertical cross-section of the crust, mantle, and slab structure in Chubu showing the nature and origin of the NKTZ (Iio et al., 2002).

viscosity as low as that of the uppermost mantle. Yamasaki and Seno (2005) examined the effect of rheological heterogeneities in the lower crust or upper mantle on the surface deformation using a two-dimensional finite element method based on the assumption that the observed high strain rates along the NKTZ are due to loading and unloading of the subducting PHS slab. The results indicate that the observed high strain rates can be reproduced by a low-viscosity upper mantle immediately beneath and trenchward of the NKTZ. It was suggested that such a low-viscosity upper mantle could be realized by the upward movement of water derived from dehydration of the subducting PHS and PAC slabs, by partial melting of the mantle above the PAC slab, and by serpentinization in the mantle wedge above the PHS slab.

Nakajima and Hasegawa (2007c) reported a detailed model of the crustal structure beneath the NKTZ based on a seismic tomography study. The analysis revealed a prominent low-velocity anomaly in the lower crust along the NKTZ in the southwest and central parts, extending to the uppermost mantle along the zone in the NE and central parts. This low-velocity anomaly was attributed to the existence of melt or aqueous fluid derived from the PHS slab, suggesting that the crust and upper mantle along the NKTZ is weakened by the accumulation of fluid. Although the origin of the observed high strain rates along the NKTZ requires further investigation, the studies to date suggest that water derived from the subducting slab or melt produced by the addition of water play an important role in stress accumulation and earthquake generation in the crust of the upper plate. This mechanism is thus similar to that responsible for the high strain rate zone along the Ou backbone range in Tohoku.

In Japan, dense temporary networks of seismographs have been deployed in the source areas of recent large inland crustal earthquakes, and seismic tomography using these post-mainshock data

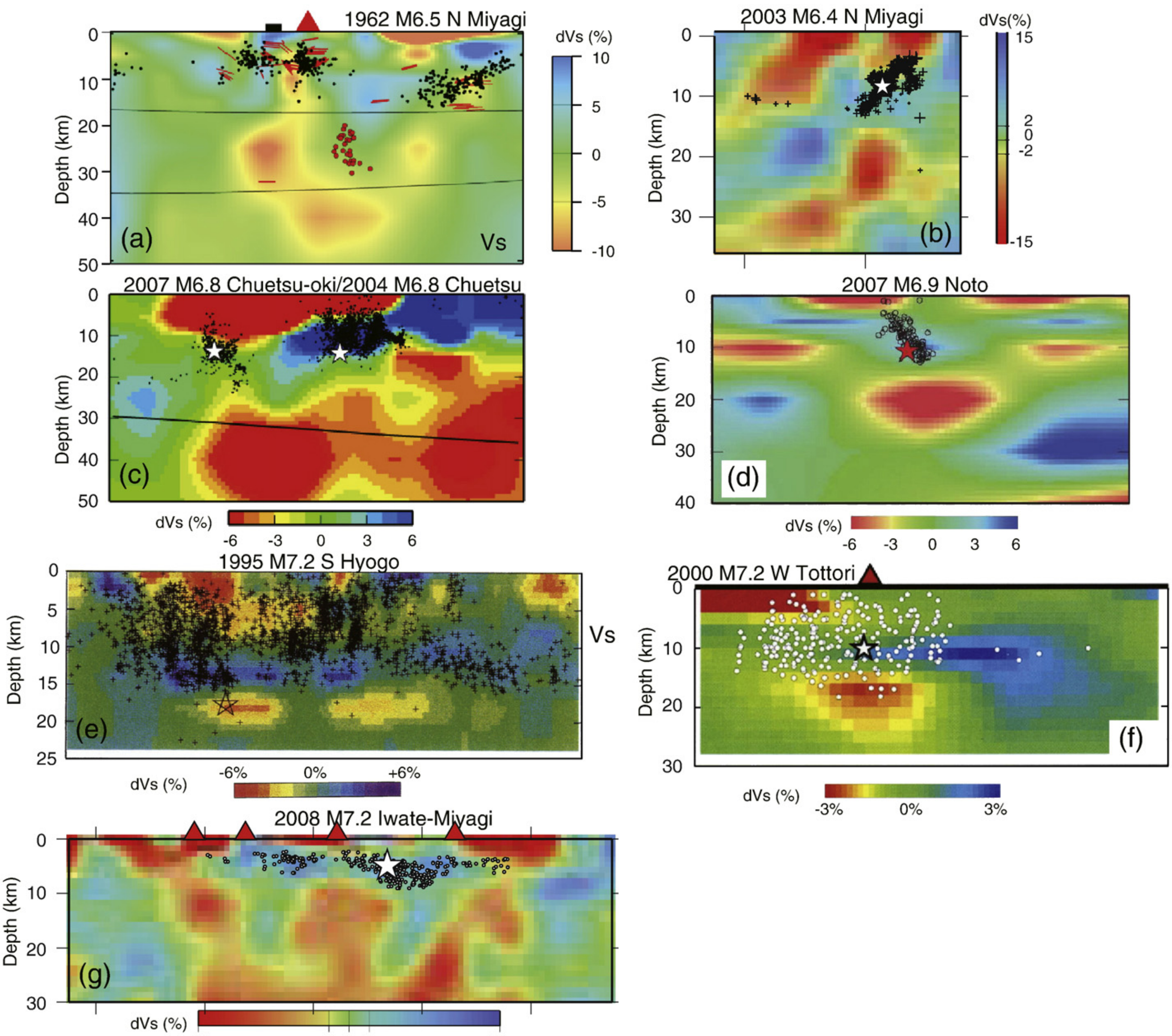

Fig. 30. Seismic velocity structure in the source areas of recent large inland crustal earthquakes in Japan. (a) 1962 *M* 6.5 N Miyagi (Nakajima and Hasegawa, 2003b), (b) 2003 *M* 6.4 N Miyagi (Okada et al., 2008), (c) 2004 *M* 6.8 Niigata–Chuetsu and 2007 *M* 6.8 Niigata–Chuetsu–oki (Nakajima and Hasegawa, 2008), (d) 2007 *M* 6.9 Notohanto–oki, (e) 1995 *M* 7.2 S Hyogo (Kobe) (Zhao et al., 1996), (f) 2000 *M* 7.2 W Tottori (Zhao et al., 2004), and (g) 2008 *M* 7.2 Iwate–Miyagi (Okada et al., 2008). (e), (f), and (g) are vertical cross-sections along the mainshock faults, and other figures are vertical cross-sections across the mainshock faults. Mainshock and aftershocks are denoted by stars and circles or crosses, respectively.

have provided further evidence of the important role of water in earthquake generation. Fig. 30 shows vertical cross-sections of the seismic velocity structure obtained by the temporary observation of aftershocks following seven recent large inland earthquakes: 1995 *M* 7.2 south Hyogo (Kobe), 2000 *M* 7.3 west Tottori, 2003 *M* 6.4 north Miyagi, 2004 *M* 6.8 Niigata–Chuetsu, 2007 *M* 6.8 Off Niigata–Chuetsu, 2007 *M* 6.9 Noto Peninsula, and 2008 *M* 7.2 Iwate–Miyagi. A vertical cross-section of S-wave velocity perturbations across the fault plane of the 1962 *M* 6.5 north Miyagi earthquake is also shown, since a dense temporary observation network has been deployed in the source area of the 1962 earthquake, and aftershock activity associated with that event remains high. Prominent low-velocity zones can be readily identified in the lower crust immediately below the fault planes of all of these large earthquakes. Some of the low-velocity zones extend downward to the upper mantle, suggesting that the low-velocities there are due to aqueous fluids or melts supplied from the upper mantle right below. All these low-velocity zones in the lower crust, including those that do not extend into the upper mantle, are inferred to be caused by water of slab origin, since seismic tomography using data acquired by temporary observation networks covering small areas do not provide high resolution for the upper mantle. Relation between low-velocity zones in the lower crust and source areas of large inland earthquakes shown in Fig. 30 is clear, but they are based on data obtained by dense temporary observation networks and so sampled areas are limited. We expect further studies using data covering a much wider regions with such dense networks to confirm the relation.

7. Conclusions

The acquisition of large volumes of high-resolution seismic data in Japan has made it possible to perform seismic analyses at unprecedented resolution over large areas of Japan, greatly contributing to the understanding of earthquake and magma generation associated with the subduction of the PAC and PHS plates in this region. Recent studies have shown that the PHS plate subducts beneath southwest Japan along the Sagami and Nankai troughs without splitting in the area north of the Izu Peninsula as suggested in earlier studies. The plate reaches a depth of approximately 200 km in north Chubu and south Kyushu, whereas the downdip limit is only approximately 60 km in the region from north Kinki to north Chugoku. In Kanto, the PHS plate comes into contact the subducting PAC plate below, resulting in anomalously deep activity of interplate earthquakes on the upper interfaces of both the PHS and PAC plates. The subducted PHS plate has an undulating configuration, probably caused by contractive deformation in the along-arc direction associated with subduction beneath southwest Japan.

Recent studies based on high-resolution seismic data indicate that interplate earthquakes are generated by dynamic slip at asperities with strong frictional coupling dominated by stick–slip behavior. These isolated asperities are embedded in regions of weak frictional coupling dominated by stable sliding, and it is the aseismic slip in the areas surrounding the asperities that causes incremental stress accumulation at the asperities, eventually leading to dynamic slip and the generation of earthquakes. Frequent episodic slow slip events have been resolved by both GPS and seismic observations in areas surrounding unruptured asperities, in some cases accompanied by deep low-frequency tremors/earthquakes.

The precise determination of the hypocenter distribution of intraslab earthquakes and the seismic velocity structure in the slab crust has provided evidence supporting the dehydration embrittlement hypothesis for the generation of intraslab earthquakes. Intraslab seismicity has a tendency to concentrate at the dehydration loci of metamorphosed slab crust and serpentinized slab mantle. Earthquakes in the PAC slab crust form a seismic belt (upper-plane seismic belt) that is oriented parallel to the iso-depth contours of the upper plate interface, which is located near the dehydration loci of the metamorphosed slab crust. This upper-plane seismic belt deepens locally in Kanto, where the dehydration loci are also expected to deepen due to contact with the cold overlying PHS slab. The downdip limit of the seismic low-velocity slab crust reaches the depth of the upper-plane seismic belt, which is also locally depressed in Kanto, as expected.

The transport of water from the slab to the arc crust beneath Tohoku has also been estimated based on high-resolution seismic observations. Both the ascending and descending portions of secondary convection flow in the mantle wedge, mechanically induced by plate subduction, have been successfully resolved as inclined seismic low-velocity zones. The ascending flow portion is characterized by a melt-filled pore structure with melt volume fraction of 0.1 to several percent, attributable to the addition of water of slab origin and decompression melting. This mantle upwelling flow reaches the arc Moho immediately beneath the volcanic front, indicating that the formation of the volcanic front is caused by this upwelling flow. Such mantle upwelling flow was detected in the mantle wedge in active volcano chains in Tohoku, Hokkaido, north Chubu, Chugoku, and Kyushu, suggesting that the volcano chains in all of these areas are formed by mantle upwelling flow.

In addition to arc magmatism, water of slab origin thus transported to the arc crust appears to cause local contractive deformation of the arc crust leading to inland crustal earthquakes. GPS observations reveal the existence of a high strain rate zone (NKTZ) that extends from Niigata to Kobe and that along the Ou backbone range. Many large inland crustal earthquakes have occurred along these zones. Distinct seismic low-velocity zones were detected in the lower crust and/or upper mantle immediately beneath these high strain rate zones, suggesting the existence of water, which should weaken the arc crust and cause local contractive deformation. The occurrence of large inland crustal earthquakes in such areas is expected to return the deformation to a spatially uniform state in the depth and along-arc directions.

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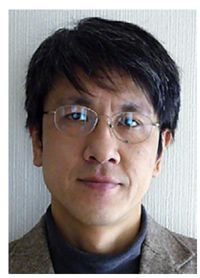

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