Seismological structure of subduction zones and its implications for arc magmatism and dynamics

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Abstract

In this paper I review recent seismological findings on the structure, magmatism, and dynamics of subduction zones. High-resolution seismic tomography has revealed prominent low-velocity (low-V) and high-attenuation (low-Q) zones that exist in the crust and uppermost mantle just beneath active arc volcanoes and extend to 400 km depth in the mantle wedge. The low-V/low-Q zones are located in the central portion of the mantle wedge and lie 30–50 km above the subducted oceanic slab. The mantle wedge low-V/low-Q zones also exhibit strong seismic anisotropy. These results suggest that arc magmatic systems are not limited to the near-surface areas, but are related to the deep processes, such as the convective circulation in the mantle wedge and dehydration reactions in the subducted slab. The low-V/low-Q zones beneath the arc and back-arc are separated at shallow levels but merge at depths >100 km, indicating that the slab components of the arc and back-arc magmas occur through mixing at these depths. These low-V/low-Q bodies form the deep roots and sources of the arc magmatism and volcanism. Large crustal earthquakes in Japan are found to occur around low-V zones that may represent weak sections of the seismogenic crust. The crustal weakening is thought to be closely related to the subduction of the oceanic Pacific and Philippine Sea plates in this region. Along the volcanic front and in back-arc areas, the crustal weakening may be caused by the active volcanism and the presence of magma chambers. In the forearc areas, fluids were detected in the earthquake source areas, which may have contributed to the rupture nucleation and may be related to the dehydration of the subducted slab. These results suggest that large crustal earthquakes may not strike anywhere, but only in anomalous areas which may be detected with geophysical methods. Suggestions are also provided for future directions of seismological research of subduction zones.

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1. Introduction

According to plate tectonics, new oceanic plates are formed at mid-ocean ridges by the upwelling of hot mantle materials, producing mid-ocean ridge magmatism. The same amount of materials returns back to the Earth’s interior at subduction zones where the heavier oceanic plates are descending beneath the more buoyant continental plates, causing the most active seismicity and volcanism on the Earth. Subduction zones are convergent plate boundaries characterized geographically by deep ocean trenches and island arcs or continental margins, seismically by landward dipping Wadati–Benioff zones of deep earthquakes, tectonically by regional-scale crustal faulting and terrane movements, and magmatically by arcuate and linear belts of eruptive centers, the so-called volcanic front. Subduction zones have long been recognized as key el-
elements in plate tectonics. Subduction and arc magmatism are fundamental processes in the evolution of the Earth. They play critical roles in the present day differentiation of Earth's material and are believed to be the major sites of generation of continental crust. Subduction is also significant in the water and carbon cycles.

In contrast to magmatism at mid-ocean ridges where it is generally agreed that the responsible process is adiabatic upwelling, arc magmatism at subduction zones has led to a range of proposals for its source region and process. The fundamental paradox is why the high heat flows and the abundance of melt generation take place right above the relatively cold subducting slab. With the advent of plate tectonics, some researchers proposed that shear and frictional heating led to melting of the oceanic crust (Oxburgh and Turcotte, 1970; Turcotte and Schubert, 1973). Hsui et al. (1983) suggested that the oceanic crust melts as a result of high temperatures arising from the hot induced mantle flow impinging on it. Some researchers proposed that the melting is the result of adiabatic upwelling of either ecologitic crust (Brophy and Marsh, 1986) or mantle wedge (Tatsumi et al., 1983; Plank and Langmuir, 1988), while others argued that the melting results from fluxing of the peridotitic mantle wedge by water from the subducting oceanic crust (Gill, 1981; Tatsumi, 1989; Davies and Stevens, 1992; Iwamori, 1998).

One of the striking observations of arc volcanism is that the active volcanoes nearest the trench form a line parallel to the trench: the volcanic front (Sugihara, 1960). In addition, these volcanoes lie approximately 100 km above the Wadati–Benioff deep seismic zone within the subducting slab (Hasegawa et al., 1978; Tatsumi, 1989; Zhao et al., 1997a,b). These constraints are quite robust, and any model hoping to explain subduction zone volcanism must be able to account for them.

Once an earthquake occurs, it emits seismic waves that propagate through the Earth's interior and can be recorded by seismographs installed at seismic stations when they finally arrive at the Earth's surface. Analyses of the observed seismic waves provide information on the rupture process of the earthquake as well as physical properties of the materials along the wave trajectories in the Earth. The classic studies of seismic waves have led to the discoveries of the layered structure of the Earth. Since the advent of seismic tomography two decades ago (Aki and Lee, 1976; Aki et al., 1977), seismologists have been mapping the heterogeneous structure of the Earth on a broad range of scales. These three-dimensional (3-D) modelings of the Earth's structure promise to answer some basic questions of geodynamics and signify a revolution in Earth science (Dziewonski and Anderson, 1984). Seismic tomography is still developing at an accelerated pace and is expected to be an active field of seismological research for several decades to come (Iyer and Hirahara, 1993). Seismic tomography has also been the most powerful tool for the imaging of subduction zones.

During the last decade, thanks to the advances of seismic imaging techniques like tomography, and high seismic activity and dense seismographic networks installed at some island arcs such as Japan, Alaska, and Tonga, seismologists have made significant progresses in understanding the structure, magmatism and dynamics of subduction zones. Here, I address the recent seismological findings on this topic. Numerous such studies of subduction zones have been published in literature in the last decade, and it would be impossible to make a thorough review of all of them in this limited space. Therefore, I will mainly focus on my own studies and those I am familiar with.

2. Seismic velocity structure

There are basically three seismological parameters to describe the Earth structure: seismic velocity, attenuation, and anisotropy. Seismic velocity has been the best estimated physical parameter of the Earth's interior because it is determined mainly from seismic wave travel times that can be measured with a high precision and a great quantity. Hence, seismic tomography methods have been mainly applied to travel time data to estimate the 3-D velocity structure. Before describing the major findings on the seismic velocity structure of subduction zones, let me first make a brief review of the recent advances in the methodology of seismic tomography at local and regional scales.

2.1. Advances in seismic tomography methodology

Most of the early tomographic studies used the method of Aki and Lee (1976) and Aki et al. (1977),
who divided the medium under study into many cubic blocks and take seismic velocities in every blocks as unknown parameters. Unavoidably, the first tomography method has several drawbacks, e.g. artificial velocity discontinuities are introduced into the model between blocks, seismic rays are assumed to be straight lines within each block, and only one iteration of inversion is conducted. Later, the block method was modified by a number of researchers (see Iyer and Hirasahara, 1993, for a detailed review). Thurber (1983) adopted grid nodes, instead of blocks, to model the Earth structure. Velocities at the nodes are taken as unknown parameters; the velocity at any point in the model is calculated by interpolating the velocities at the eight nodes surrounding that point. Thus, the velocity is continuous everywhere in the model; no velocity discontinuity is allowed to exist in the model even when discontinuities are actually detected in the study area. Thurber (1983) adopted the Cartesian coordinates, hence his method has been applied mainly to local-scale studies.

Japan provides a useful illustration of some of the weaknesses of these earlier methods. Japan is located in a complex region where four plates are interacting with each other (Ishida, 1992). The upper boundary of the subducting Pacific slab is found to be a sharp seismic discontinuity with a complex geometry by detailed studies with reflected and converted waves (Fukao et al., 1978; Hasegawa et al., 1978; Nakanoishi et al., 1981; Matsuzawa et al., 1986, 1990; Zhao et al., 1997a; Ohmi and Hori, 2000). The Moho and the Conrad discontinuities also have large lateral depth variations (Horiuchi et al., 1982a,b; Zhao et al., 1990, 1992b). These discontinuities generate clear converted and reflected waves observable in seismograms and should be taken into account when studying the 3-D Earth structure. In principle, when the block or grid size is sufficiently small, the effects of the discontinuity undulations can be evaluated. The size in actual studies, however, is not small enough because of the sparse distribution of seismic stations. Hence, the tomographic images obtained not only represent velocity variations themselves, but also contain the effects of discontinuity undulations. Without introducing discontinuities into the model, the observed later (converted and reflected) phase data cannot be used. Later phase data contain important information on the Earth structure. The use of them can improve the ray path coverage and the accuracy of earthquake locations. However, neither Aki nor Thurber’s methods can handle the complex discontinuities and the later phase data.

To tackle these problems, Zhao et al. (1992a) developed an improved tomography method. They divided the medium under study into a number of layers bounded by the discontinuities that are known to exist. 3-D grid nets are set individually in every layers. They also developed a fast 3-D ray tracing technique to compute travel times and ray paths in such a model by combining the pseudo-bending technique (Um and Thurber, 1987) and Snell’s law. First, P- and S-waves and later phase data can all be used in the tomographic inversion with the LSQR algorithm (Paige and Saunders, 1982). Later, Zhao et al. (1994, 1997b) improved this method to make it use data from local, regional and teleseismic events simultaneously to determine the 3-D velocity structure under a region. Recently, Zhao (1999) extended this method to the global scale to estimate the whole mantle 3-D structure with the data set compiled by Engdahl et al. (1998). He took into account the depth variations (up to 36 km) of the 410 and 660 km discontinuities (Flanagan and Shearer, 1998) in the inversion and found that the discontinuity topography has a significant influence on the tomography of the mantle transition zone.

In the early 1980s, the introduction of iterative matrix solvers (e.g. SIRT, LSQR) by Clayton and Comer (1983) and Nolet (1985) proved to be an important step toward the refining of the Earth structure. It allowed for the implementation of larger numbers of both seismic data and model parameters in tomographic inversions. Recently, a few researchers have attempted to estimate resolution and covariance matrices using the iterative algorithms (like LSQR) without explicitly performing the operations involving the large and complex matrices (Zhang and McMechan, 1995; Nolet et al., 1999; Yao et al., 1990). Although there are still some debates on the technical details, this approach seems very promising and will become a useful tool in future tomographic studies if a consensus can be achieved.

Great advances have been made in the last two decades on the development of fast 3-D ray tracing algorithms (Thurber and Ellsworth, 1980; Thurber, 1983; Um and Thurber, 1987; Vidale, 1990; Moser, 1991; Zhao et al., 1992a; Koketsu and Sekine, 1998; Sadeghi et al., 1999), which make it possible to con-
duct nonlinear and iterative tomographic inversions. Some researchers adopted irregular parameterizations to account for the inhomogeneous sampling of the Earth by seismic rays (Abers and Roecker, 1991; Vesnauer, 1996; Bijwaard et al., 1998). Hu et al. (1994) and Steck (1995) combined teleseismic polarization data with arrival times to conduct tomographic inversions. Hirahara (1988) and Wu and Lees (1999) proposed to detect 3-D velocity anisotropy in the study area using travel time data, and formulated the problem of travel-time inversion for both velocity heterogeneity and anisotropy in each block. Other examples include Lees and van Decar (1991), who used gravity data to constrain travel-time inversion, and Neele et al. (1993), who used teleseismic P-wave amplitudes together with travel times to determine the upper mantle velocity structure.

2.2. Subducting slab and arc magmatism

Zhao et al. (1992a, 1994) applied their tomography method to over 50,000 arrival times from local, regional and teleseismic events to determine a detailed 3-D P-wave velocity structure of the Japan subduction zone. In addition to first P- and S-wave data, they also used converted waves at the upper slab boundary and the Moho discontinuity. They picked or double checked all the arrival times directly from the seismograms, which ensured the high accuracy of their data. Fig. 1 shows the P-wave tomography in northeast Japan they obtained. The spatial resolution of the tomographic images is 25–30 km in horizontal direction and 10–25 km in depth. The subducting Pacific plate is clearly imaged as a cold slab that is 90 km thick and has a velocity 4–6% higher than the normal mantle. Intermediate-depth earthquakes occur in the subducting slab and form two distinct seismic planes, the so-called double seismic zone (Hasegawa et al., 1978). The upper-plane earthquakes occur at the top of the slab; the lower-plane earthquakes occur in the central portion of the slab. Low velocity (low-V) zones are visible just beneath the active volcanoes, and extend to 150 km depth in the mantle wedge. Anomalous low-frequency microearthquakes occur around the low-V zones beneath active volcanoes in the depth range of 25–47 km, which are caused by the magmatic and volcanic activity (Hasegawa and Zhao, 1994). Distinct shear-wave reflectors are also found near the low-V zones in the lower crust, representing the upper surface of hot magma chambers with a thickness of about 100 m (Hasegawa and Zhao, 1994; Matsumoto and Hasegawa, 1996). Based on these observations, the low-V zones in the crust and mantle wedge are considered to be associated with the magma chambers which form the roots of the arc volcanoes.

Mineral physics studies of phase changes of some mantle minerals, such as olivine, wadsleyite and ringwoodite at high-temperature and high-pressure conditions found that the dehydration of the subducted oceanic crust plays an important role in the generation of melts and magma in the mantle wedge subduction zones (Ringwood, 1982; Irifune, 1993; Iwamori, 1998).

Fig. 1 shows some heterogeneities within the subducting Pacific slab. It is found from map views that earthquakes within the slab (the lower-plane events in the double seismic zone) occurred in the relatively higher velocity areas of the slab (Fig. 18 in Zhao et al., 1992b). There may be two origins for the heterogeneity in the slab. One is the heterogeneity in the oceanic plate before subduction, for instance, seamounts, ridges, and fracture zones, as can be seen from maps of sea floors. The other may be the phase changes of the rocks within the slab due to the increasing temperature and pressure during the subduction of the oceanic plate (e.g. Kao and Liu, 1995).

Analyses of converted waves at the upper boundary of the subducting Pacific slab revealed a thin low-V layer on the top of the slab, which is interpreted to be the subducted oceanic crust (Nakanishi et al., 1981; Matsuzawa et al., 1986; Abers, 2000). The low-V layer is thinner than 10 km and about 6% slower than the normal mantle. The depth extent of the low-V layer is unclear, but is estimated to be deeper than 100 km, and earthquakes in the upper plane of the double seismic zone occur within it (Matsuzawa et al., 1986). The tomographic images in Fig. 1 do not have enough spatial resolution to detect such a thin low-V layer on the top of the slab.

Iidaka and Suefutsumu (1992) suggested the existence of a low-V metastable olivine wedge within the subducting Pacific slab in the Izu–Bonin region. In the Tonga slab, however, Koper et al. (1998) did not find such an olivine wedge though they have used a much better data set and it is more likely that such an olivine wedge would exist in the Tonga slab than in
Fig. 1. Vertical cross sections of P-wave velocity structure beneath northeast Japan from the Earth’s surface to 200 km depth along the profiles shown in the insert map. Red and blue colors denote slow and fast velocities, respectively. The velocity perturbation scale is shown at the bottom. Open circles denote earthquakes that occurred within a 20-km width along each profile. Red circles show low-frequency microearthquakes that occurred around the Moho discontinuity, due to the magmatic and volcanic activity. Red triangles, active volcanoes; reverse triangle, the location of the Japan Trench. The horizontal bar at the top of each cross section shows the land area where seismic stations exist. The three thick curved lines denote the Conrad and Moho discontinuities and the upper boundary of the subducting Pacific slab. The dashed line shows the lower boundary of the slab (after Zhao et al., 1992a).
the Izu–Bonin slab because of the much faster convergence rate. Although the feasibility of a metastable olivine wedge within the subducting slab has been established by mineral physicists (Green and Burnley, 1989; Kirby, 1991), it is still a controversial issue to seismologists. This is an important subject for future seismological research since it has important implications for the mechanism of deep earthquakes and dynamics of subduction zones.

2.3. Deep dehydration and back-arc spreading

Most of the seismic stations in Japan are located on the narrow land area of approximately 200 km in width, hence high-resolution tomographic images are determined only down to about 200 km depth (Fig. 1). The depth extent of the low-V zones in the mantle wedge is unclear in the back-arc region. Recently, Zhao et al. (1997b) used the data recorded by land seismic stations and ocean bottom seismographs (OBS) to determine a detailed 3-D structure down to 700 km depth beneath the Tonga arc and the Lau back-arc (Fig. 2). The subducting Tonga slab is imaged as a 100-km thick zone with a P-wave velocity 4–6% higher than the surrounding mantle. Beneath the Tonga arc and the Lau back-arc, low-V anomalies of up to 6% are visible. The low-V anomaly beneath the Tonga arc represents a dipping zone about 30–50 km above the slab, extending from the surface to about 140 km depth. This feature is similar to the low-V zones found beneath the Japan and Alaska volcanic fronts (Zhao et al., 1992a, 1995). Beneath 100 km depth, the amplitude of the back-arc anomalies is reduced, but a moderately slow anomaly (−2 to −4%) exists down to a depth of at least 400 km. This deep extent of the mantle wedge slow anomalies has been confirmed by detailed resolution analyses (Zhao et al., 1997b), seismic attenuation tomography (Roth et al., 1999, 2000) and waveform modeling studies (Xu and Wiens, 1997). These results indicate that geodynamic systems associated with the back-arc spreading are not limited to the near-surface areas, but are related to deep processes. The slow velocity anomalies at depths of 300–400 km in the mantle wedge (Fig. 2) could be caused by upwelling flow patterns in the mantle wedge or by volatiles resulting from the deep dehydration reactions occurring in the subducting slab (Nolet, 1995).

Volatile would have the effect of lowering the melting temperature and the seismic velocity, and may produce small amounts of partial melt (Collier and Sinha, 1992). Temperatures in fast subducting slabs like Tonga are low enough for water to reach the stability depths of dense hydrous magnesian silicate phases (Parson and Wright, 1996; Taylor et al., 1996), which may allow water penetration down to depths of 660 km. The phase diagrams of important hydrous phases, the associated reaction kinetics, and the relevant mantle conditions (slab temperature and composition) are not known sufficiently well to predict the depth at which dehydration would occur. Partial melting of the mantle wedge by volatiles from the deep slab may be important in localizing low seismic velocities; the slow anomalies we observe at depths of 300–400 km (Fig. 2) may represent this process.

The slow velocity regions beneath the Tonga arc and the Lau back-arc seem to be separated at the shallow levels, but merge at depths >100 km (Fig. 2). This suggests that although the arc and back-arc magma systems are separated at shallow levels, where most of the magma is generated, there may be some interchange between the magma systems at depths >100 km. Interchange with slab-derived volatiles at depths >100 km may help to explain some of the unique features in the petrology of back-arc magmas relative to typical mid-ocean ridge basalts, including excess volatiles and large ion lithophile enrichment (Faul et al., 1994).

The spatial resolution of a tomographic experiment depends on the density and homogeneity of crisscrossing seismic rays available in the inversion. The reliability (quality) of a tomographic solution depends on the data accuracy as well as the density of rays used. Rays used in the Japan tomography are those from the local events in the crust and the subducting slab as well as teleseismic rays which were recorded by a dense seismic network located only on the island area of about 200 km in width. In contrast, rays in the Tonga tomography are from local and teleseismic events recorded by a seismic array that is less denser but has a much wider aperture than that in Japan. In addition, the abundant deep earthquakes in Tonga provided many rays sampling the deep parts of the mantle wedge. The data in both experiments have high quality. Therefore, the resolution of the
Fig. 2. East–west vertical cross section of P-wave velocity image from 0 to 700 km depth beneath the Tonga arc and Lau back-arc region. Red and blue colors denote slow and fast velocities, respectively. Solid triangles denote active volcanoes. Earthquakes within a 40-km width from the cross section are shown in open circles. The velocity perturbation scale is shown at the bottom (after Zhao et al., 1997b).

Fig. 3. P-wave velocity image at a depth of 40 km beneath (a) northeast and (b) central Japan. Red and blue colors denote low and high velocities, respectively. Circles denote earthquakes (M 5.7–8.0, depths 0–20 km) that occurred during a period of 115 years from 1885 to 1999. Solid triangles denote active volcanoes. Active faults are shown by thick lines. The velocity perturbation scale and the earthquake magnitude scale are shown on the right and at the bottom, respectively. Crosses and open squares in (a) show low-frequency microearthquakes and S-wave reflectors in the mid-crust, respectively.
Fig. 4. East–west vertical cross section of microearthquakes and their cutoff depth (solid lines) beneath Unzen Volcano in Kyushu (after Matsuo, 1985).

Fig. 5. Along-arc vertical cross section of microearthquakes that occurred during 1 November 1988 to 30 June 1996 and the estimated crustal temperature in the central part of northeast Japan. Events located in the rectangular area MN in the insert map are plotted in the cross section. Solid triangles and crosses show active volcanoes and deep, low-frequency microearthquakes, respectively. On the top the thin line (Ak) denotes the aftershock area of the 17 October 1970, Southeast-Akita earthquake (M 6.2); the short bar (Kw) and two bars (Sy) denote Kawafune Fault and Senya Fault which were ruptured by the 31 August 1896, Rikuu earthquake (M 7.2). The temperature of the crust is shown in isotherms of 300, 400 and 500°C, which was estimated from seismic velocities determined by seismic tomography (after Hasegawa et al., 2000; Zhao et al., 2000a,b).
Fig. 6. Schematic illustration of across-arc vertical cross section of the crust and uppermost mantle in volcanic areas of Japan, showing the cause of large crustal earthquakes and its relation to low-\(V\) zones and magma chambers in the uppermost mantle.

Japan tomography is very high (10–30 km) down to 200 km depth right beneath the volcanic arc (Fig. 1), but very low under the forearc, back-arc and deeper areas. In contrast, the Tonga tomography has a resolution of 40–50 km for a much wider area under the Tonga arc and the Lau back-arc down to 700 km depth (Fig. 2).

2.4. Large earthquakes in arc and back-arc: influence of magma

Countries located in subduction zone regions, such as Japan, have suffered heavily and frequently from earthquake hazards. Interplate earthquakes occur in mega-thrust zones along oceanic trenches. Intraplate

Fig. 7. (a) Epicentral distribution of 3634 events used in the tomographic imaging of the source area of the 1995 Kobe, Japan, earthquake (star). Crosses denote the events that occurred after 17 January 1995; most of them were aftershocks of the Kobe earthquake along the fault zone (parallel to cross section line A–B). Circles denote microearthquakes that occurred from January 1990 to December 1994. (b) Distribution of seismic stations that recorded the earthquakes in (a). Solid triangles denote portable stations that were set up following the Kobe mainshock. Solid squares denote permanent stations. Solid lines represent the surface traces of the Nojima, Suma, and Suwayama faults (after Zhao and Negishi, 1998).
earthquakes within the continental plate take place in the crust beneath the island arc or continental margin. Although large intraplate earthquakes do not occur so frequently as the interplate ones, they generally inflict greater damages because they are shallow and near the densely populated areas. A recent example of an inland crustal earthquake is the 1995 Kobe earthquake (M 7.2) in southwest Japan, which caused over 6400 fatalities and tremendous property losses. A good understanding of where large crustal earthquakes occur is important for the clarification of physics of earthquake generation and for the mitigation of seismic hazards.

Zhao et al. (2000a) investigated the relationship between the 3-D crustal structure and the distribution of large crustal earthquakes in Japan in the recent history. They found that most of the 160 large crustal earthquakes (M 5.7–8.0, depth 0–20 km) during a period of 115 years from 1885 to 1999 in Japan occurred around zones of low seismic velocity revealed by seismic tomography (Fig. 3). The low-\( V \) zones

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Fig. 8. (a) Distribution of aftershocks (dots) of the 1995 Kobe earthquake (star) within a 5-km width from the line A–B in Fig. 7. (b) P-wave velocity (\( V_p \)), (c) S-wave velocity (\( V_s \)), and (d) Poisson’s ratio images along the line A–B. Slow velocity and high Poisson’s ratio are shown in circles; fast velocity and low Poisson’s ratio are shown in crosses. \( V_p \) and \( V_s \) perturbations range from –6 to 6% from the 1-D velocity model. Poisson’s ratio ranges from 0.225 to 0.27 (–10 to 8% from the average value). The vertical exaggeration is 2:1 (after (Zhao and Negishi, 1998)).
may represent weak sections of the seismogenic crust. Along the volcanic front and in back-arc areas, the crustal weakening may be caused by active volcanism and the presence of magma chambers. low-V zones in the uppermost mantle may be the manifestation of mantle diapirs associated with the ascending flow of subduction-induced convection in the mantle wedge and dehydration reactions in the subducting slab (Zhao et al., 1992a) (Fig. 1). Magmas further rising from the mantle diapirs to the crust may cause low-frequency microearthquakes at levels of the lower crust and uppermost mantle, and make their appearance as S-wave reflectors at midcrustal levels. Their upward intrusion raises the temperature and reduces the seismic velocity of crustal materials around them, causing the brittle seismogenic layer above them to become locally thinner and weaker. This can be seen clearly from the distribution of crustal seismicity in volcanic areas (Figs. 4 and 5). The cut-off depth of the crustal earthquakes is elevated toward the volcanoes, indicating a higher temperature and a thinner brittle seismogenic layer beneath the active volcanoes (Matsuo, 1985; Ito, 1993).

Subject to horizontally compressional stresses in the plate convergence direction, contractive deformations will take place mainly in the low-\(V\), low-\(Q\) areas because of the thinner brittle seismogenic layer and the weaker crust and uppermost mantle there due to the higher temperature. The deformation proceeds partially in small earthquakes but mainly in plastic deformation, causing the crustal shortening, upheaval and mountain building there, as evidenced by the geodetic measurements and GPS studies (Hasegawa et al., 2000). Large crustal earthquakes cannot occur within the weak low-\(V\) zones but in their edge portions where the mechanical strength of materials is stronger than those of the low-\(V\) zones but still weaker than the normal sections of the seismogenic layer. Thus, the edge

Fig. 9. (a) Vertical cross section of P-wave velocity structure down to a depth of 100 km along the line AA' in the insert map. Blue and red colors denote fast and slow velocities, respectively. The velocity perturbation scale is shown at the bottom. The star symbol shows the hypocenter of the 1995 Kobe mainshock (M 7.2). White dots show the microearthquakes within a 20-km width from the line AA', which occurred during 1985 to 1993. The thick lines on the top show the land areas, the Chugoku District and Awaji Island. The open triangle denotes the Kannabe Quaternary volcano in Chugoku.
2.5. Large earthquakes in forearc: effects of slab dehydration

Zhao et al. (2000a) found that large crustal earthquakes in the forearc regions in Japan are also located in or around low-\(V\) zones. Those low-\(V\) zones, however, may not be caused by high temperature because no volcano exists there (Yokoyama et al., 1987) and the surface heat flow is low (Okubo et al., 1989). The 1995 Kobe earthquake (M 7.2) is such an example. Kobe is located in the forearc region where the Philippine Sea plate is subducting beneath the Eurasian plate (Fig. 7).

Zhao et al. (1996) and Zhao and Negishi (1998) determined 3-D P- and S-velocity and Poisson’s ratio structures in the Kobe source area with a spatial resolution of 4–5 km (Fig. 8). They found that the Kobe mainshock is located in a distinctive zone characterized by low velocity and high Poisson’s ratio, which is interpreted to be a fluid-filled, fractured rock matrix that contributed to the initiation of the Kobe earthquake.

There may be two origins of fluids in the Kobe fault zone: one has a shallow origin, such as fluids trapped in the pore space, crustal mineral dehydration, and the permeation of the meteoric and sea waters down to the deep crust through the active faults that would have been ruptured during many earthquake cycles (Zhao and Mizuno, 1999); the other has a deep origin, such as the dehydration of the subducting Philippine Sea slab.

Zhao et al. (2000a) determined the detailed 3-D crust and upper mantle structure under Shikoku and Chugoku (Fig. 9). The subducting Philippine Sea slab is imaged clearly. It has a thickness of 30–35 km and a P-velocity 3–5% higher than that of the normal mantle. A prominent low-\(V\) zone exists in the lower crust and right above the Philippine Sea slab. This low-\(V\) zone has properties as the anomaly at the Kobe hypocenter that Zhao et al. (1996) detected in their high-resolution imaging which shows low \(V_p\), low \(V_S\) and high Poisson’s ratio (Fig. 8). These results suggest that the fluids that contributed to the initiation of the 1995 Kobe earthquake may be related to the dehydration process of the subducted Philippine Sea slab, though the fluids may also have the shallow origins as mentioned above. It is generally considered that fluids widely exist in the crust and uppermost mantle in the forearc regions of subduction zones (Tatsumi, 1989; Iwamori, 1998). The Philippine Sea slab is descending at a very small dip angle in Shikoku and eastern

Fig. 10. Schematic illustration on the effects of slab dehydration on the generation of large crustal earthquakes in the forearc region of the Nankai subduction zone. According to mineral physics studies, dehydration reactions would take place in the subducting oceanic lithosphere when it descends into the mantle due to the increasing temperature and pressure. The Philippine Sea slab is descending at a small dip angle beneath southwest Japan, and it is located right under the crust, thus the fluids from the slab dehydration may easily migrate up to the crust. When the fluids enter the active faults in the crust, pore pressures will increase and fault zone frictions will decrease. Thus, active faults may be triggered to move to generate large crustal earthquakes.
Kii Peninsula, and the slab is located right under the crust (Ishida, 1992), thus the fluids from the slab dehydration may easily migrate up to the crust. When the fluids enter the active faults (such as the Nojima Fault which generated the 1995 Kobe earthquake) pore pressures will increase and fault zone friction will decrease. Thus, active faults can be triggered to move to generate large crustal earthquakes (Fig. 10).

These results suggest that the generation of a large crustal earthquake is closely related to the surrounding tectonic environment such as plate subduction and physical/chemical properties of crustal materials, such as magmas, fluids, etc. The rupture nucleation zone should have a 3-D spatial extent, not just limited to the 2-D surface of a fault, as suggested earlier by Tsuboi (1956) in the concept of “earthquake volume”. Complex physical and chemical reactions may take place in the source zone of a future earthquake, causing heterogeneities in the material property and stress field. The source zone of a M 6–8 earthquake extends from about 10 to over 100 km (Kanamori and Anderson, 1975). The resolution of our tomographic imaging is close to that scale of the earthquake sources, which may have enabled us to image the earthquake-related heterogeneities in the crust and uppermost mantle in Japan.

3. Seismic attenuation structure

Seismic attenuation and its variations in the Earth are useful for determining the type and state of the rocks and minerals composing the Earth. The attenuation of seismic waves is due to three effects: geometric spreading, intrinsic attenuation, and scattering attenuation. Geometric spreading is simply the energy density decrease that occurs as an elastic wavefront expands. Intrinsic attenuation is energy lost to heat and internal friction during the passage of an elastic wave. Scattering attenuation is not true energy loss in this sense. Elastic energy is not converted into heat but is redistributed into angular directions away from the receiver or converted into wave types arriving in different time windows at the receiver. Scattering takes place by reflection, refraction, and conversion of elastic energy by the medium heterogeneities that are discontinuous or rapid variations in the velocity and/or density of the medium. Intrinsic attenuation structure of the Earth can be estimated from observed data of seismic wave attenuation with a method similar to velocity tomography. The data used are amplitudes or amplitude spectra of seismic waves. Attenuating bodies reduce the amplitude or change the spectral content of seismic waves that pass through them, just as low-V bodies increase the travel times of seismic waves. The parameter used to quantify seismic attenuation is seismic wave quality factor, $Q$, value. High-$Q$ and low-$Q$ values indicate weak and strong seismic attenuations, respectively. Compared with velocity tomography, attenuation ($Q$) tomography has, in general, a lower spatial resolution because the amplitude or amplitude spectrum data are usually measured by investigators manually or semi-automatically and thus their measurements are much fewer than arrival time measures. The $Q$ images are usually less accurate than velocity images because the amplitude (spectra) measurements are noisier. Moreover, these inaccuracies are somewhat compensated by the much larger variations in $Q$ than velocity anomalies. The velocity may vary 5–10% between the slab and the mantle wedge, but $Q$ varies by 100% (factors of 10). There are also much fewer $Q$ studies than velocity studies.

During the recent years, a few 3-D attenuation models have been determined for subduction zone regions, mainly in the Japanese Islands (e.g. Umino and Hasegawa, 1984; Furumura and Moriya, 1990; Sekiguchi, 1991; Tsumura et al., 2000). These studies revealed that the subducting Pacific slab shows very small attenuation and has $Q$ values of up to 1000 to 1500. In contrast, the mantle wedge and the crust beneath active volcanoes show very high attenuation with $Q$ values of 100 or smaller. Recently, Tsumura et al. (2000) obtained an updated 3-D attenuation structure in northeast Japan. Their model has a spatial resolution of about 40 km, and shows clearly the high-$Q$ Pacific slab and the low-$Q$ anomalies in the crust and mantle wedge beneath active volcanoes. The general pattern of the $Q$ variations is quite similar to that of velocity variations in this region (Fig. 1) determined by Zhao et al. (1992a). Roth et al. (1999) estimated the 3-D $Q$ structure of the Tonga–Fiji region and found that low-$Q$ anomalies extend to 400 km depth in the mantle wedge above the high-$Q$ Tonga slab, in good agreement with the velocity images by Zhao et al. (1997b) and Roth et al. (2000).
Attenuation structure of subduction zones has also been estimated from the seismic intensity data (e.g. Hashida, 1989). Seismic intensity is usually measured on the basis of human perception and movement of objects observed by experts without instruments. The estimated attenuation structure shows similarity to the $Q$ models determined with the precise seismological data such as seismic wave amplitudes and amplitude spectra. Apparently, however, the intensity data cannot be expected to determine high-resolution and accurate $Q$ structure because of its subjective nature.

As a whole, the images of the low-$Q$ zones in the crust and mantle wedge beneath active volcanoes and the high-$Q$ zones corresponding to the subducting slab are similar to those of the low-$V$ and high-$V$ zones revealed by travel time tomography (Tsumura et al., 2000; Roth et al., 2000). These 3-D seismic attenuation models have provided additional information on the physical properties of subduction zones, and show the mantle wedge anomalous bodies associated with the magmatism of the island arc and back-arc.

4. Seismic anisotropy

Seismic anisotropy is the direction-dependent nature of the propagation velocity of seismic waves. Natural minerals usually have some crystallographic structure. Under the physical conditions of high temperature and high pressure in the Earth’s deep interior, rocks undergo slow plastic deformation. Through the deformation process, crystallographic axes of minerals are realigned to a particular direction due to uniaxial tectonic stress. Rocks may thus exhibit petrofabric structure, which in turn will produce seismic anisotropy on a macroscopic scale. Anisotropy can also form due to the presence of aligned cracks. It is found that seismic anisotropy seems to exist in almost every portions of the Earth, from the crust, mantle, to inner core.

Shear wave splitting has been the most useful tool to detect seismic anisotropy, and numerous researchers have made studies using this approach for subduction zones (e.g. Ando et al., 1980; Kaneshima and Silver, 1992; Okada et al., 1995; Yang et al., 1995; Hiramatsu et al., 1997; Fischer et al., 1998). Shear wave splitting studies show the existence of anisotropy in the crust, mantle wedge and the subducted slab, among them the mantle wedge seems to be the most anisotropic portion of subduction zones.

Okada et al. (1995) made a detailed investigation of seismic anisotropy in northeast Japan using shear wave splitting data from local shallow and deep earthquakes. They found that shear waves from the earthquakes in the subducting slab show significant splittings (strong anisotropy) of up to 1 s when the waves pass through the low-$V$ and low-$Q$ zones in the mantle wedge as imaged by the velocity and attenuation tomography (Zhao et al., 1992a; Tsumura et al., 2000). The shear waves show little splitting when they propagate the normal areas in the crust and mantle wedge. They suggested that the anisotropy was caused by the planar alignment of melts within the low-$V$/low-$Q$ zones in the crust and mantle wedge beneath the active volcanoes. The fraction of melts is estimated to be about 2% from the degree of anisotropy and attenuation, and from the velocity reduction.

The local seismic rays from the shallow and deep events used by Okada et al. (1995) show that the anisotropy of the crust and the subducting slab is insignificant, while Hiramatsu et al. (1997) suggested that the subducting slab shows strong anisotropy resulting from phase changes in the slab. Hiramatsu et al. (1997) used ScS waves which travel from the events in the slab down to the deep mantle and are bounced back from the core-mantle boundary. ScS waves propagate a long distance in the mantle and may be affected by the anisotropy in the deep mantle and/or the D'' layer above the core-mantle boundary (Iidaka and Niu, 1998; Vinnik et al., 1998). It is also possible that the local rays of Okada et al. (1995) did not sample the deep and inner portion of the slab where phase changes take place and seismic anisotropy exists. Future studies are needed to clarify whether the subducting slab is anisotropic or not.

5. Concluding remarks

Although great advances have been made in the seismic imaging of subduction zones in the last decade, a great deal remains to be done in the future in both the theoretical and observational aspects of seismology. The tomographic methods that have been the most powerful tool to imaging subduction zones need further improvement. The existing schemes for model
parameterization, ray tracing and inversion need to be thoroughly compared and examined (e.g. Spakman and Nolet, 1988; Boschi and Dziewonski, 1999); new and more efficient techniques should be explored continuously. New theory and technologies are needed to use amplitudes, polarizations and waveforms together with travel times in the tomographic inversion for seismic velocity, attenuation and anisotropy structures.

Much more work remains to be done on the application of tomography and other seismic methods to the imaging of subduction zones, which largely depends on the installation of new seismic networks or the expansion of the existing networks. Most of the existing seismic stations are located on island arcs or continental margins, thus only the structure in or around the volcanic front can be well imaged. To study the forearc and back-arc regions, installing of OBS stations is crucial. Precise images of the forearc regions of subduction zones are important for clarifying the initiation of subduction, seismic and mechanic coupling of the subducting oceanic slab and the overlying continental plate, and the frequent occurrence of destructive thrust-type earthquakes (Zhao et al., 1997a; Ito et al., 2000). Detailed structure of the back-arc regions of subduction zones is needed to understand the formation of back-arc spreading centers and its relationship to the subduction process (Zhao et al., 1997b; Roth et al., 1999, 2000).

Other seismological puzzles and research topics of subduction zones include: the formation of the double seismic zone and its relationship to the stress regime and structural heterogeneities in the shallow portion of the subducting slab (Kao and Liu, 1995); the existence or absence of a metastable olivine wedge within the deep portion of the slab and its relationship to the generation of deep earthquakes (Idaka and Suetsumu, 1992; Koper et al., 1998); the geometry of the mantle discontinuities near the subducting slab (Collier and Helffrich, 1997); the detailed morphology and structure of the subducting slab around the 670 km discontinuity (Zhao and Clayton, 1990; van der Hilse et al., 1991; van der Hilse, 1995); high-resolution imaging of the mantle wedge to clarify the relationship between the arc magmatism and slab dehydration (Iwamori and Zhao, 2000; Zhao et al., 2000b); the detailed structure of the mantle below the subducting slab and the nature of the lower boundary of the slab (Hasegawa et al., 1994; Zhao et al., 1994), among others.

So far, most of the seismological studies on large earthquake sources have concentrated on the coseismic rupture process on the fault plane through analyzing the seismic waves generated during the faulting, which provides little information on the preparatory process of the earthquake generation. To tackle this problem, it is useful to conduct high-resolution tomographic imagings of the earthquake source areas and active fault zones. The effects of inelastic structures and processes such as magmas and fluids in the crust and lithosphere need to be paid special attention on the generation of earthquakes.

The challenge is great, but with ingenuity, striving, and collaboration, we can anticipate exciting new advances in understanding the structure and dynamics of subduction zones as well as the entire Earth’s interior in the 21st century.

Uncited reference


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