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Role of Water in Conductive Anomalies and Seismic Reflections in the Lower Crust

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Abstract

Relationships between electrical conductivity and seismic reflection in the lower crust are discussed in terms of the presence of aqueous fluids. Tectonic/geological interpretations of the lower crustal structures revealed by recent geophysical surveys are given for the region of stable cratons, convergent margins and rift zones, respectively. Most of the lower continental crusts of the stable craton are electrically resistive, and there are little correlations between electrically conductive zones and seismically reflective zones. In the convergent plate margins, the lower crust is conductive and seismic reflectors (or bright spots) are observed at mid-crustal depth. The rift zones are also characterized by both strong anomalies in electrically conductivity and seismic reflectors throughout the lower crustal column. The strong correlation between the conductive zones and the reflective zones observed from the active convergent margins and young rift zones is possibly explained by the presence of the free fluid phase.

The distribution of fluid in the lower crust is discussed from the phase relation and wetting properties of constituent minerals. The presence of anhydrous granulites in the outcrops of the lower crustal sections and xenoliths derived from the lower crust indicates that the lower crust is generally dry. In the exposed lower crustal section of the Kohistan Arc hydrous minerals are localized in the vicinity of veins and ductile shear zones developed in the host anhydrous lithology. On the other hand, wetting behavior of plagioclase the most dominant mineral in the lower crust, indicates the absence of interconnecting aqueous and CO_2-rich fluids through the equilibrium intergranular channels under lower crustal *P-T* conditions. It suggests that fluid flow in the lower crust is not pervasive but is localized in cracks and ductile shear zones. Ductile shear zones developed in the lower crust are the most probable sites for water pathways which are responsible for conductive anomalies and seismic reflectors in the continental crust revealed by geophysical observations.

Key words: fluid, lower crust, geophysical anomaly, ductile shear zone, wetting property

1. Introduction

Seismic tomography studies demonstrate that the low velocity zones correspond to weak sections of the seismogenic crust (Zhao *et al.*, 1996; 2000). Since fluids can contribute to crustal weakening and rupture nucleation, generation of earthquakes may be closely related to the presence of the free fluid phase in the crust. Source regions of fluids located on or adjacent to major fault systems, suggests that the faults act as fluid migration pathways through the crust, and geophysical data show that the regional-scale surface faults extend to the highly reflective, bright layer and the high electrical conductivity zone in the lower crust (e.g., Zhao *et al.*, 1996; Ryberg and Fuis, 1998; Ichiki *et al*, 1999; Zhu, 2000; Ogawa *et al.*, 2001). These observations are thought to require the presence and supply of free water in the lower crust beneath the brittle-ductile transition zone to

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trigger a crustal earthquake. Indeed, most of geophysicists have interpreted electrical conductivity anomalies and seismic reflectors of the lower crust in terms of the presence of free water (e.g., Gough, 1986; Hyndman and Shearer, 1989). However, the question of whether or not the lower crust contains water remains as an unsolved problem because petrological evidence denies the presence of a free aqueous fluid phase in the lower crust (Yardley, 1986; Yardley and Valley, 1994; 1997).

This paper aims to provide the best possible explanation concerning the role of water in electrical conductive anomalies and seismic reflectors of the lower crust on the basis of our present knowledge. Evidence for the presence of fluids in the lower crust can be acquried in three ways: (1) studies of the modern crust by geophysical measurements of properties such as seismic velocity and electrical conductivity, (2) petrological and geochemical studies of deep crustal materials obtained from exposed rocks or xenoliths and (3) experimental studies of the conditions under which lower crustal mineral assemblages are stable. This paper reviews these approaches. These considerations suggest that free aqueous fluid is restricted in heterogeneous structures localized in the lower crust. The present paper will propose the hypothesis that water-saturated ductile shear zones are the cause of the electrical conductivity anomalies and seismic reflectors in the lower crust and will discuss the validity of free water in the ductile shear zone.

2. Geophysical anomalies in the lower crust

2. 1. Electrical conductivity anomaly

Electrical conductivity is sensitive to very small changes in minor constituents of a rock, and hence is used complementary to other geophysical techniques, such as gravity and seismic velocities, which are sensitive to bulk properties of rock. As silicate minerals have very low electrical conductivity, the conductivity of bulk rock is generally a function of the interconnection of a minor electrically conductive constituent, such as saline fluids, partial melts, sulphides and graphite (e.g., Brace, 1971; Waff, 1974; Hermance, 1979; Duba and Shankland, 1982; Sato and Ida, 1984; Frost *et al.*, 1989; Glover and Vine, 1992; 1994; Duba *et al.*, 1994; Shankland *et al.*, 1997; Raab *et al.*, 1998). Therefore, determination of the electrical structure in the lower crust should provide information about minor conductive phase distribution.

Electrical conductivity distribution within the continental lower crust has been previously determined by four methods: the magnetotelluric (MT) method, the geomagnetic depth sounding (GDS) method, the horizontal spatial gradient (HSG) method, and the controlled source electric-magnetic field (CSEM) method. The most popular technique for determining conductivity distribution is the MT method, which is a natural-source procedure utilizing time-varying electromagnetic fields that result from electric storms and solar activity (Jones, 1992). However, a resolution of MT measurements for the lower crust is more than 10 km, and this decreases with increasing depth because of the diffusive propagation of electromagnetic waves in the Earth. Since measured electrical conductivity represents bulk conductivity within the resolution, a highly conductive layer at a scale less than 10 km may exist in a more resistive matrix.

A number of compilations of lower crustal conductivity data have been carried out to provide a representative absolute conductivity and correlations with geothermal structures (e.g., Shankland and Ander, 1983; Haak and Hutton, 1986; Jones, 1992). These classifications do not differ significantly from each other. In this paper, the author has classified conductivity distribution into three categories; stable cratons (or shields), rift zones and convergent margins including subduction and collision zones.

In stable cratons, where tectonic activity such as volcanism or metamorphism is inactive and surface heat flow is very low, generally speaking, the lower crust is relatively resistive when compared to the Phanerozoic mobile belts. According to Hyndman and Shearer (1989) the mean resistivity for Phanerozoic sites is 20–30 Ωm, with most values lying between 3 and 100 Ωm, whereas the Precambrian shield areas have a mean resistivity of about 500 Ωm. Furthermore, deep crustal conductivity tends to increase markedly from the early Archean through the Proterozoic. The highest resistivities (10^3 Ωm or higher) in the Precambrian lower crust are located around the center of the stable cratons. For example, continental lower crusts of the Canadian shield could be zoned into three general types with the central part highly resistive (10^3–10^4 Ωm) zone, an edge of intermediate resistivity (100–300 Ωm), and some zones of anomalously high conductivity (10–50 Ωm) in association with the suture zone (Jones, 1981; Mareschal *et al.*, 1995). Although the continental lower crust resistivity beneath stable cratons is generally high (in the order of 10^4 Ωm), there is distinct evidence that the resistivity can be an order of magnitude lower, such as found beneath the eastern part of the Siberian shield (Vanyan *et al.*, 1989), the southern part of the Baltic Shield (Jones *et al.*, 1983), and northern Wisconsin (Dowling, 1970), but these have been extensively reworked during the Proterozoic (Goodwin, 1996). Wannamaker (2000) reconstructed the 2-D SVEKA model of Korja and Koivukoski (1994) along Finland (from Proterozoic to the Archern transect), together with the 2-D model isotherms and Moho depth from seismic refraction. The results show an inconsistency of thermal structure with resistivity structure. Therefore, the enhanced conductivity of these old regions could not be explained by a simple function of continental geotherm.

Electrical structures of rift zones are characterized by a zone of high conductive anomalies (less than 50 Ωm) at 10–30 km depth. The modern rift zone is the most conductive tectonic environment. For example Rio Grande rift two-dimensional modeling shows some variations ranging from a thick zone of low resistivity (10 Ωm) in the middle to lower crust to a relative resistive zone (400 Ωm) (Jiracek *et al.*, 1983). A two-dimensional resistivity model across the Miocene extensional detachment of the Betic Chain in Spain shows a highly conductive zone (<5 Ωm) at lower crustal levels, which is interpreted as partial melting of a southeast dipping subducted Iberian lower crust (Pous *et al.*, 1999). In most modern rift locations the highly conductive lower crust could be interpreted as interconnected partial melt zones based on the anomalous high heat flow.

The electrical resistivity structures between the extending zone and the surrounding tectonically inactive zone can be compared with the aid of MT profiles across the young rift zone. Popov (1987) illustrates that the resistivity of the Baikal region in Siberia decreases from 200 to 30 Ωm toward the rift zone. Recent deep conductivity profiles from two MT transects in the Great Basin, western USA, show that the lower crust beneath the eastern area with modern active extension is more conductive (7 Ωm), while the central part, where the extension has become quiescent since 5–10 Ma is moderately conductive (20 Ωm) (Wannamaker *et al.*, 1997). The upper levels of each conductor are interpreted to be high-salinity brines with a minimal activity of H_2O (Wannamaker, 2000). This is because the temperature estimated from the typical geotherm of the area at the tops of the conductors is ~550℃ corresponding to amphibolite facies metamorphic grade, which is lower than the solidus temperature of wet granite.

Most of the past rift zones dating before Proterozoic also have a conductive nature. Results in the MT data from the 1.1 Ga Keweenawan Midcontinent Rift in North America show asymmetrical conductivity structures, which are composed of easterly dipping conductive zone (<10 Ωm), that are related to a low angle detachment fault (Serpa *et al.*, 1984) and a resistive zone greater than 10^4 Ωm (Wunderman *et al.*, 1985). The preservation of high conductivity anomalies under reduced thermal gradient due to the termination of the extensional event indicates that the cause of conductive anomalies cannot be explained by partial melting alone.

Imaging the conductivity structure in the vicinity of modern continental active margins such as found in the subduction zone is difficult due to the coastal effect, because conductive seawater masks the minimal conductance of crustal rocks. However, recent studies also show evidence of a conductive lower crust at modern subduction zones. Mostly with the aid of the geomagnetic depth sounding (GDS) method, Utada (1987) discovered the conductive anomaly in the lower crust of Northeastern and Central Japan. Although recent MT results show a resistive lower crust of Southwestern Japan (Shiozaki *et al.*, 1998), Shiozaki *et al.* (1999) also reported the presence of conductive lower crust (<10^2 Ωm) beneath the coastal region of the Japan Sea. The crustal structure of the Kii Peninsula in the forearc shows a presence of a conductive layer in the lower crust at 15–25 km depth (Fujita *et al.*, 1997). The conductive zone distributed on the fault plane at depths from 2 to 16 km in the northern part of Miyagi Prefecture, northeastern Japan correlates to the hypocentral zone and to the lower crust, and to the relatively conductive blocks (lower than 5 Ωm) existing below a depth of 15 km (Ichiki *et al.*, 1999). The most precise, EM investigations on a continental margin have been carried out along the west coast of North America, where the young Juan de Fuca plate subducts beneath the continental plate (Kurz *et al.*, 1986). The results show a gently-dipping conductive zone (30 Ωm) suggesting a subducting oceanic crust (Kurz *et al.*, 1986) or being within an accretionary wedge above the top of a subducting plate (Hyndman, 1988).

For a typical arc-continent collision, there is a conductive zone in depths of 10–20 km beneath the Island of Taiwan (Chen and Chen, 1998). This conductive zone was interpreted as mobilized fluids derived from dehydration reactions beneath Taiwan. The 2-D section of the Himalayas of Central Nepal shows high conductivity in the foreland basin ($\sim$30 Ωm) that contrasts with the resistive Indian basement ($>$300 Ωm) and Lesser Himalayas ($>$ 1,000 Ωm) (Lemonnier *et al.*, 1999). It is interpreted that the high conductivity zone reflects metamorphic fluids that were released during underthrusting of the Indian basement. In summary, the observed high electrical conductivity anomalies at convergent boundaries are generally interpreted as a zone of trapped free water, since the process of a convergent zone leads to the release of dehydrated water from subducting-slab components.

There are general indications as to the electrical structure of the crust obtained from the geophysical surveys as follows. (1) The continental lower crust is generally more conductive ($<10^4$ Ωm) in comparison with the expected conductivity of dry crystalline rocks (Kariya and Shankland, 1983). (2) The electrical conductivity increases sharply at mid-crustal depths (15–20 km) (Jones, 1992) and then in most cases decreases in lower crustal depths. (3) Electrical conductivities beneath the modern tectonically active zone such as the rift zone, the collision zone and the subduction zone are even higher than those beneath the inactive zone. (4) The rift zone shows the most conductive nature (<10 Ωm). (5) The lower crust, on the average, is more conductive in the Phanerozoic areas than in the Precambrian areas (Hyndman and Shearer, 1989), although traceable structures of ancient tectonic activities could often be observed as moderately conductive anomalies in the ancient tectonic belts.

2. 2. Seismic reflection

Deep seismic profiling has been used as a standard tool to visualize the deep structure in the continental crust and occasionally in the lithospheric mantle since the 1980's, for example COCORP in the USA, BIRPS in the UK, LITHOPROBE in Canada and INDEPTH in the Tibet-Himalayas. Recently multichannel seismic profiling has been applied to the island arc to clarify deep crustal structures of the arc-arc collision zone (e.g., Arita *et al.*, 1998; Tsumura *et al.*, 1999). The continental lower crust exhibits bright, sub-horizontal, laterally extensive, and multicyclic layered reflections. In contrast, this reflection signature is rare in the continental upper crystalline crust and upper mantle, and in the oceanic lithosphere. It has been shown that bright reflections in the lower crust have normal incident reflection coefficients of around 0.1; that is, they require an acoustic impedance contrast of approximately 20%. Although many explanations for layered reflections within the lower crust have been proposed, only three explanations have been intended to provide a more general application. The reflections are mainly due to: (1) the existence of free aqueous fluids, (2) ductile shearing, and (3) compositional layering.

This section summarizes several review papers (Klemperer, 1987; Warner, 1990; Mooney and Meissner, 1992) with the additional results of recent geophysical research. Tectonic settings of reflective lower crusts are also classified into three categories; stable cratons, rift zones and convergent zones. The reflective patterns of stable cratons show the most complicated feature compared to other tectonic settings. Early seismic reflection profiles in stable cratons showed a decrease in the number of reflections with depth, and the Moho was indicated only by a few weak and discontinuous reflections (Smithson, 1989). However, these seismic property concepts of the stable craton have undergone a significant reevaluation because of the change from conventional mechanical seismic sources to more powerful sources. These profiles, Arunta block in Australia (Goleby *et al.*, 1989), Superior Province in Canada and the USA (Jackson *et al.*, 1990), and the Baltic Shield (Behrens *et al.*, 1989), appear to possess various seismic reflectivity responses, ranging from transparent to highly reflective laminae. The various reflective structures may be derived from the superposition of successive tectonic events throughout the history of crust. Therefore these profiles must have been formed by the integration of some tectonic events such as collisions, extensions and subductions.

In this section, rift zones imply areas where the last tectonic event was extensional, such as some grabens in Western Europe, the Basin and Range Province of the USA, and many passive margins. In these areas the lower crust (below a depth of about 10–14 km) is highly reflective, in contrast to the less reflective appearance of the crystalline upper crust and upper mantle. The reflections of Moho are nearly flat, and upper crustal faults, where they can be traced into the lower crust, are listric at depth and merge into the reflective lower crust. The BIRPS profiles typically show bright lower crustal seismic layering from the continental shelf south of west of Britain, where the area underwent post-Caledonian or post-Variscan crustal extension (Peddy *et al.*, 1989). While the majority of the BIRPS profiles resemble each other, the geometry and continuity of the layering is highly variable (Warner, 1990). Some reflections show individual continuous reflections of 10–20 km length, while others show fragmental reflections traceable for only 1–2 km. The Basin and Range Province underwent Cenozoic extension that generated a metamorphic core complex that is associated with listric and high-angle normal faults (Wernicke, 1981; Davis *et al.*, 1986). This section also shows a reflective lower crust and a remarkably planer Moho (Klemperer *et al.*, 1986). Seismic reflection profiles of the inner zone of the Northern Apennines, Italy, where the lithosphere has been affected by extensional deformation since the Middle Miocene, show the occurrence of a sharp reflector, which is probably located at the top of a shear zone acting as the décollement horizon for the latest (Pliocene-Quaternary) normal faults (Liott a and Ranalli, 1999). Depth of a sharp reflector changes from a shallower locus (3～6 km) in geothermal areas to a deeper one (9～12 km) elsewhere.

Collision zones have thick crustal roots that are approximately in isostatic equilibrium with their high topography. Many collision zones have been investigated in the past two decades. Early seismic reflection studies were concentrated on the continent-continent collision of the Cenozoic collision belts found in the Himalayas and the Alps. The Eastern Alps traverse shows a representative seismic reflection image of continent-continent collision belts. The lower crust dips toward the center of orogenic belts from both sides and shows seismic laminations terminating approximately at the Moho (Pfiffner *et al.*, 1988; Frei *et al.*, 1989; Blundell *et al.*, 1992). A deep seismic profile across the Pyrenees shows a similar configuration to the Alps traverse. The crust within the axial zone of the collision belt has been thickened by at least 15 km by the thick-skinned stacking of whole-crustal flakes (ECORS Pyrenees Team, 1988; Choukroune and the ECORS Team, 1989). Reflectivity can be followed to greater than 20 s two-way-taravel-time (more than 60 km), which suggests that lower crustal delamination also occurred beneath this orogen.

A recent example of a deep seismic reflective profile across the collision zone was provided by the Tibet-Himalayas transect (Hauck *et al.*, 1998). A band of bright spot reflections has been imaged by Project INDEPTH (International Deep Profiling of Tibet and the Himalayas) at about a 15 km depth along 150 km of the northern Yadong-Gulu rift, southern Tibet. Makovsky and Klemperer (1999) suggests that free aqueous fluids on the order of 10% volume in the Tibetan middle crust produces the observed bright spot reflections based on reflection-amplitude variation with offset (AVO) modeling. If such relatively large quantities of free aqueous fluids exist at the mid-crustal level, origin of the bright spots would be explained as a result of transient flow of aqueous fluids in the southern Tibetan crust (Makovsky and Klemperer, 1999).

Deep seismic profiling of the arc-arc collision belt in the Hidaka Mountains in Japan, shows quite similar features to those of the continent-continent collision belts (Arita *et al.*, 1998; Tsumura *et al.*, 1999). The important features obtained by them is that upper and lower crusts of the Hidaka arc separated into two portions due to the collision of the Kuril forearc. The lithospheric mantle beneath the Northeastern Japan Arc shows a wedge structure. This structure was explained as a result of lithospheric delamination of the Kuril arc 23 km deep in the lower crust. A reflective lower crust in the arc crust can be observed beneath a depth of 18 km.

Large-scale seismic reflective structures across the magmatic arc including both the island arc and the continental arc have not been conducted. However, some local studies suggest the presence of a reflective lower crust. Early work at subduction zone (Hyndman, 1988) suggests that the geophysical data acquired in the LITHOPROBE program that take place across the northern Cascadia subduction zone at Vancover Island have shown dipping reflectors at a depth of about 30 km with about 5 km thickness. Wide-angle seismic fan data of the Aleutian island arc shows well-defined reflectors toward the back arc and the volcanic chain (Fliedner and Klemperer, 2000). The 'bright spot' was observed adjacent to Mount Cayley, a large Quaternary volcanic center associated with the Cascadia subduction zone magmatic arc (Hammer and Clowes, 1996). Although universal validity of the presence of a reflective lower crust remains as an unknown problem, the accumulation of current studies confirms the existence of a bright reflective layer in the accretional complex and the magmatic arc.

2. 3. Linkage between electrical and seismic structures

Gough (1986), Klemperer (1987) and Ádám (1987) gave a general view that seismically reflective zones and electrically conductive zones are restricted to the lower crust. The depth to the upper bounds of these zones are 10–15 km, and the temperature estimated from surface heat flow suggests that the tops of most reflective bands and conductive layers are about 350–400℃ corresponding to the greenschist facies metamorphic condition. Based on the compiled data from geophysical observations at sporadic areas, Hyndman and Shearer (1989) suggested that the temperature condition corresponds to a depth where hydration reaction or precipitation of silica occurs. However, a correlation between seismic and electric structures should not be justified by comparison with geophysical data obtained at different places.

Comparison with onset depth of these anomalies obtained from a same location would help to understand whether these anomalies have the same origin or not. There are some examples to allow the evaluation of the correlations. There is an example in the case of stable craton. The Kapskasing Uplift in Ontario representing an exposure of rocks from deep levels within the crust, is an important condition for evaluating correlations between tectonic structures in association with the exhumation of deep crustal rocks and geophysical anomalies. The Kapskasing Uplift is the oblique exposed crustal cross section, that has been uplifted in association with thin-skinned tectonics of the early Proterozoic age (Percival and Card, 1985; Percival *et al.*, 1989). Seismic reflection studies suggest that reflectors are abundant in the upper 8 s (< ~25 km), diminishing downwards to ~13 s (40 km near Moho) (Percival *et al.*, 1989). Although the kilometer-scale, laminated, and deep crustal reflections, are confirmed from the profiles, the tops of the reflective zone seem to be relatively shallow (6–10 km). Images of high-resolution EM surveys at the Kapskasing Uplift show that conductivity rises at depths of 15–20 km (Bailey *et al.*, 1989). The characteristic results show lateral homogeneity and a higher resistive nature of the uplifted lower crustal section located at the upper crustal depth (Woods and Allard, 1986; Kurtz *et al.*, 1989; Bailey *et al.*, 1989). Although the most significant results of the electrical structure is that the resistivity depends on intrinsic conditions such as pressure and temperature rather than lithological boundaries or thrust faults, there is a little correlation between seismic reflections and conductive anomalies.

In contrast, the rift zone represents a strong correlation between the conductive and reflective zones. The Valhalla complex, a Cordilleran metamorphic core complex, is a domal culmination made up of gently dipping interlayered sheets of igneous and supracrustal rock body that was deformed and metamorphosed in the Middle Jurassic and Late Cretaceous periods, and exhumed by extensional faults in the Eocene (Schaubs and Carr, 1998). The results of the multidisciplinary studies, as part of the LITHOPROBE, classified this crustal section into three parts; (1) moderately reflective, highly resistive (thousands of Ωm), and cold (<450℃) upper crust, (2) highly reflective, moderately resistive (hundreds of Ωm), and warm (450–730℃) middle crust, and (3) non-reflective, highly conductive (~5 Ωm), and hot (> 730℃) lower crust (Jones, 1992). Three stratifications are interpreted as a dry upper crust, a wet middle crust and a partially molten lower crust, respectively (Lewis *et al.*, 1992).

Results from some subduction zones also show a strong correlation between conductive and seismic structures. The results from crustal structures beneath Vancouver island show a dipping conductive zone (30 Ωm) within an accretionary wedge above the top of a subducting plate, which correlates with a

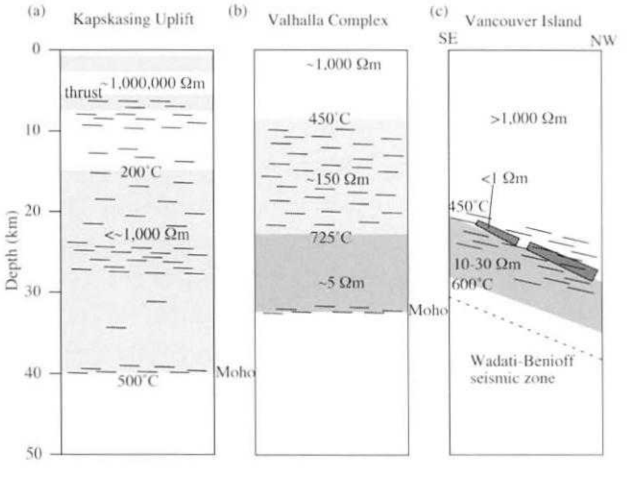

Fig. 1. Schematic cross-sections based on the combination of seismic and electrical data for three tectonic environments (stable craton, rift zone and convergent zone). Light and dark shaded regions represent slightly and extremely higher conductive regions, respectively. (a) Geophysical structures of Kapskasing Uplift as an example of stable craton (Percival *et al.*, 1989; Bailey, 1990), showing a cold, highly resistive, reflective upper crust to 15 km except for the thrust zone, a cold, somewhat conductive, reflective mid-crust to 30 km, and a warm, somewhat conductive, transparent lower crust to the Moho at ~40 km. (b) Geophysical structures of the Valhalla complex as an example of rift zone (Lewis *et al.*, 1992), showing a cold, highly resistive, unreflective upper crust to ~10 km, a warm, somewhat conductive, reflective mid-crust to ~22 km, and a hot, highly conductive, transparent lower crust to the Moho at ~33 km. (c) Geophysical structures of Vacouver Island as an example of a subduction zone (Hyndman, 1988), showing a dipping highly conductive zone from the model of Kurz *et al.* (1986) and dipping reflectors at the top of the downgoing oceanic slab.

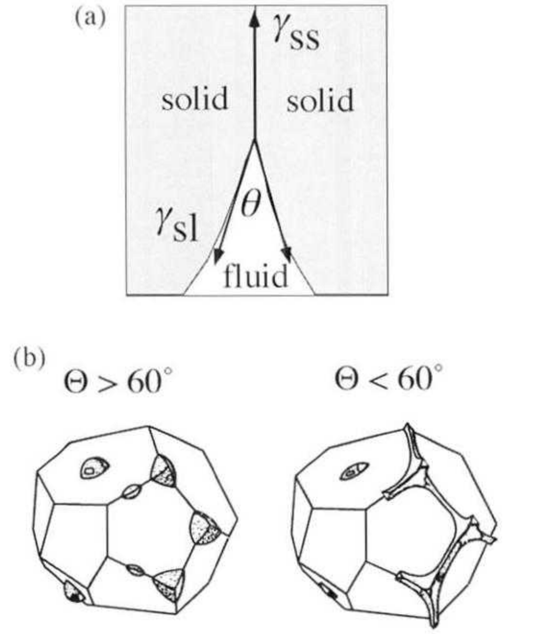

Fig. 2. (a) The balancing of interfacial energies at the pore corner in textural equilibrium. γ_{ss} is the interfacial energy per unit area of the solid-solid interface, and γ_{sl} is the interfacial energy per unit area of the solid-liquid interface. (b) Perspective drawing illustrating the distribution of fluid around a single grain in a rock as a function of the dihedral angle. The fluid phase has a three-dimensional continuity for $\theta < 60^\circ$, but occupies isolated pores when $\theta > 60^\circ$ (After Watson and Brenan, 1987).

strong reflection zone. Hyndman (1988) interpreted the conductive zone as the presence of free saline water generated by the dehydration of the downgoing oceanic plate. The crustal structure of the Kii peninsula, in the southwestern Japan forearc shows that the mid-crustal seismicity clusters near the bottom of resistive body and a conductive layer in the lower crust is consistent with seismic reflectors at 15-25 km depth (Fujita *et al.*, 1997).

Figure 1 shows schematic crustal sections for electrical and reflective structures in variable tectonic environments. Inconsistency of distribution between conductive and reflective zones in Proterozic or Archean terrains suggests a possibility that the geophysical anomalies shown in these old cratons have different origins. Presence of good correlations between seismic reflectivity and conductive anomalies would be only relevant to young tectonically active sites.

3. Pore geometry of crustal materials

The fluid-filled pore plays an important role on the physical properties of crystalline rocks. Because the seismic characteristics of the fluid phase are quite different from those of solid mineral phases, the increment in volume fraction of the fluid phase reduces the velocity of the bulk rock system. Indeed reduction of seismic velocities of both longitudinal waves and shear waves in crustal rocks with saturated-pore fluid has been observed experimentally under low temperature condition. Electrical conductivity is also influenced by small amounts of conductive fluid phase. As saline fluid has $\sim 10^7$ times greater conductivity with dry silicate rocks, electrical conductivities in fluid-saturated rocks increase with increasing porosity (ϕ). Based on the experimental measurements, an empirical relationship known as Archie's law has been widely used in sedimentary rocks and applied to many crystalline rocks at low temperatures. However, the physical properties of fluid-bearing rocks are not only related to porosity but also pore geometry. Especially, electrical conductivity is more sensitive to fluid connectivity controlled by pore geometry.

The importance of pore geometry to seismic velocity or conductivity is predicted from theoretical models (e.g., O'Connell and Budiansky, 1974; Mavko, 1980; Schmeling, 1985; 1986; von Bargen and Waff, 1986, Takei, 1998). These models provide a good chance to recognize pore geometry and porosity in crystalline rocks based on the studies of seismic tomography structures combined with attenuation of both longitudinal and shear waves against the seismic structure references. Technical development of seismic tomography will enable us to distinguish fluid from melt using the difference in physical parameters between melt and fluid and to predict pore geometry (Zhao, 2001).

The degree of pore interconnection critically affects the electrical conductivity of rocks with pore fluids or partial melts. Although lower crustal temperatures change for various tectonic regimes from the tectonically active zone to stable craton, conductive anomaly can be recognized at temperatures greater than 350℃ (Hyndman and Shearer, 1989), corresponding to the brittle-ductile transition. At temperatures higher than 350℃, textural equilibrium should be achieved between fluids and solids. Considering the pore geometry in the lower crust, the state of textural equilibrium may be an important factor to determine fluid connectivity.

At low fluid fraction ($\phi < \sim 0.01$) corresponding to the lower crustal condition, behavior of fluid in texturally equilibrated materials strongly depends on wetting properties between solids and fluids (e.g. von Bargen and Waff, 1986; Watson and Brenan, 1987). The grain-scale distribution and connectivity of the fluid phase in isotropic rocks is controlled by the dihedral angle (θ), formed at the intersection of a grain boundary with two solid/fluid interfaces. It has been shown both theoretically and experimentally that for a value of θ less than 60°, fluid interconnects along the grain edges and for a value of θ over 60°, fluid exists as isolated pores at grain corners (Fig. 2). Therefore determination as to whether the dihedral angle is over 60° or not provides a guide to assess pore connection at low fluid fraction.

Experimentally determined value for the dihedral angle in geological systems fall in the range 40–100° for H_2O-CO_2-NaCl fluids in the monomineralic aggregates such as quartz, calcite, halite, olivine and clinopyroxene under variable *P-T* conditions (review from Holness, 1997a; b). The observed dihedral angles range across 60°. There are general indications for dihedral angles of silicate minerals obtained from the hydrostatic experiments (e.g., Watson and Brenan, 1987; Watson *et al.*, 1991; Watson and Lupulescu, 1993; Holness, 1993; Mibe *et al.*, 1998; 1999; Yoshino *et al.*, in press) as follows. (1) Mafic minerals such as clinopyroxene and olivine show larger dihedral angles in comparison with felsic minerals such as quartz and plagioclase at the same *P-T* condition. (2) The dihedral angle for CO_2-rich fluids is generally higher than that for aqueous fluid without CO_2 except for clinopyroxene. (3) Brine in quartz aggregate possess a lower dihedral angle, and the dihedral angle decreases with increasing salinity in aqueous fluid. However, this property has not been observed for other minerals. (4) Increase in temperature or pressure generally leads to a reduction of the dihedral angles.

These observations suggest that the dihedral angle is regarded as a function of temperature, pressure and fluid composition. Reduction of the dihedral angle is observed in systems where either the fluid phase contains highly surface-active species and/or the fluid phase has a structure and composition close to that of the solid. The former cases are confirmed by observations of brine in quartz aggregates at a lower dihedral angle (Watson and Brenan, 1987; Holness, 1993). The latter cases are suggested by some observations that increase in mineral solubility to aqueous fluid with increasing temperature and pressure leads to the reduction of the dihedral angles (Holness, 1993; Yoshino *et al.*, in press), and that silicate melts having similar structure to silicate minerals show lower dihedral angles (20–60°) than fluids (Holness, 1997 a; b).

For common crustal rocks such as granitic and gabbroic, quartz and plagioclase can be considered as the most significant mineral controlling the pore connection. First, wetting behavior of quartz aggregate is the best investigated example as well as that of olivine aggregate, which is significant in the uppermost mantle. Watson and Brenan (1987) concluded that in the continental crust, saline fluids showing a low dihedral angle (less than 60°) in quartz aggregate contributes to high electrical conductivity and allows intergranular flow. These experiments were conducted on the 1.0 GPa, corresponding to the depth of Moho for a typical continental crust. Holness (1993) experimentally clarified the pressure dependence of the dihedral angle for quartz aggregates, suggesting that the dihedral angle is over 90° in a lower pressure condition (~0.6 GPa at 800℃), corresponding to the lower crustal condition. Thus it is unlikely to form an interconnected network along the grain edge in the most of the lower crust, even if the salinity is high enough. There is evidence for reduction in dihedral angles in terms of addition of other minerals to quartz aggregates. The dihedral angle of aqueous fluid in quartz aggregates decreases drastically with the addition of trace amounts of alkali feldspar under a narrow temperature range just below the melting point, where the magnitude of the reduction can reach 30°, that is below 60° (Holness, 1995). However, the effect of an additional mineral phase on the wetting properties of quartz aggregates is probably negligible at lower temperature conditions (< 550℃) far below the melting point.

The fluid connectivity of the lower crustal rocks should be approximated by the wetting behavior of plagioclase rather than quartz. Seismic velocity and Poisson's ratio structures in lower crust from magmatic arc or Proterozoic stable cratons have documented overall mafic composition and existence of slightly thick (*c.* > 10 km) mafic crust (e.g. Christiansen and Mooney, 1995; Rudnick and Fountain, 1995; Zandt and Ammon, 1995). The corresponding mafic lithologies are amphibolite to granulite facies rocks derived from basalts, gabbros and anorthosites. Tak-

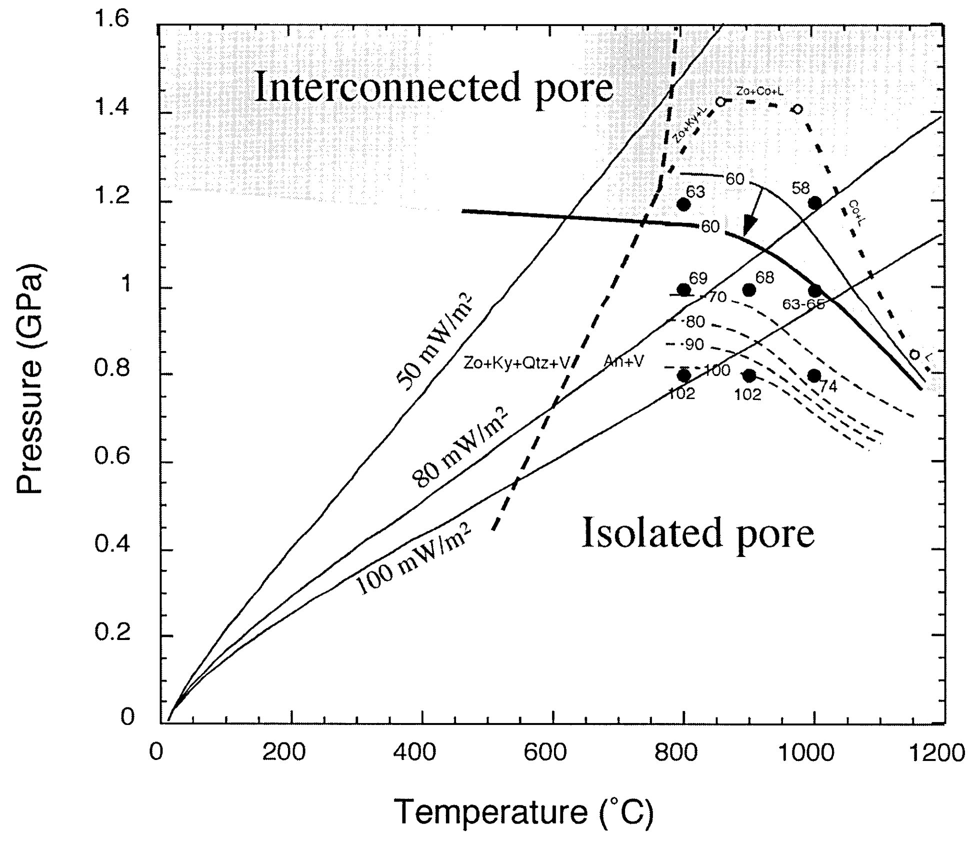

Fig. 3. *P-T-θ* plot for pure anorthite-H_2O (after Yoshino, 2000). The thick dashed line denoting the wet melting curve and phase boundaries from Boettcher (1970). The thin lines show some typical continental geotherms with value of surface heat flow. Arrow indicates shift of the contour of the critical dihedral angle (60°) due to the effect of the albite component on the dihedral angle. Abbreviations; An: anorthite, Qtz: quartz, Ky: kyanite, Zo: clinozoisite, V: vapor phase, L: melt, Co: corrundum.

ing into consideration the abundance and mineral assemblage under these metamorphic grades, plagioclase is the most significant mineral controlling the fluid connectivity in the lower crust.

Experimentally determined dihedral angles for C-H-O fluids of anorthite, which is the Ca end member of a plagioclase solid solution, are over 60° under conditions of 800–1,000℃ (<1 GPa) (Yoshino *et al.*, in press) (Fig. 3). Although the addition of albite (Na) component leads to a reduction of the dihedral angles for aqueous fluid, the absolute values are also over 60° at the same *P-T* range except for higher-pressure conditions (>1 GPa). Aqueous fluid can connect along the grain edge at only the lowermost part of the Tibetan and Andean crusts with the anomalous thickness (>40 km). Dihedral angles of anorthite for CO_2 fluid are over 120°. These observed high dihedral angles suggest non-wetting properties of the mafic rocks for C-H-O fluids under the lower crustal conditions. If fluid connectivity in the lower crust is controlled by the wetting behavior of plagioclase, estimation of fluid residence time in crust derived from porous flow through the crustal column based on the wetting properties of quartz is incorrect (e.g., Bailey, 1990; Frost and Bucher, 1994).

Non-connectivity of CO_2 fluid in silicate minerals-CO_2 system under even higher pressure and temperature conditions may provide some indications to the explanations for observed high conductivities in the lower continental crust. Frost *et al.* (1989) reported graphite on grain boundaries in plagioclase-rich rocks from the Laramie Anorthosite Complex by using Auger electron spectroscopy. It had been considered that graphite, which is an excellent conductor, precipitates on grain surfaces from a CO_2-rich fluid during cooling (Lamb and Valley, 1985), although the stability of graphite strongly depends on the oxygen fugacity. Given typical oxygen fugacity estimated from the most of granulite terranes, graphite formation on the grain boundary is petrologically possible. However, Frost *et al.* (1989) noted that the fine graphite film along the grain boundaries shows a disconnected nature. This observation is consistent with laboratory data on CO_2-rich fluids indicating large wetting angles (θ>60°) such that the fluids form isolated grain boundary corners (Watson and Brenan, 1987; Gibert *et al.*, 1998; Yoshino, 2000). This suggests that graphite precipitated from CO_2-rich fluids is primary disconnected. In addition, the presence of intergranular thin graphite film cannot produce seismic anomalies. Although graphite hypothesis is attractive for origin of the high electrical conductivity of the Precambrian crust with cold geotherm, it would not explain the excellent correlations between conductive zone and reflective zone in the tectonically active zone.

The non-wetting properties imply that a lack of connectivity of the fluid phase in crustal rocks under hydrostatic condition is a clear contradiction in the face of the observed high electrical conductivity (Fig. 3). However, it should be noted that these experiments have been carried out under hydrostatic conditions. Wetting behavior of rocks suffering from tectonic stress may be different from that of rocks under hydrostatic conditions because of the anisotropy of crystal configuration under a differential stress field. Both conductive anomalies and seismic reflections should represent some instabilities in the continental lower crust.

4. Water distribution and pathways: examples in the Kohistan Arc's lower crust

The geological study of exposed lower crustal sections is one of the available ways to constrain the fluid distribution of the lower crust. Characteristic distributions of metamorphic minerals provide clues to fluid distribution and fluid transport mechanisms, and lead to an understanding of the systematic fluid transport mechanism throughout the lower crustal column. The Kohistan Arc, in the NW Himalayas, is an ideal terrain for studies relating to the deep crust, where deep crustal metagabbros are extensively exposed (Yoshino *et al.*, 1998). The Kohistan Arc has been recognized as a Cretaceous intra-oceanic island arc sandwiched between the Asian and Indian plates and exposes approximately the whole crustal section (Tahirkheli *et al.*, 1979; Coward *et al.*, 1986). Although the host rocks of the lower crust of the Kohistan Arc is basically composed of anhydrous mineral assemblage, hydrous assemblage such as amphibolites occasionally develop within the lower crustal column. Therefore, amphibolites are regarded as the fossils of water passages. The occurrence of amphibolite is strongly restricted to the heterogeneous structures such as veins or ductile shear zones.

4. 1. Ductile shear zone

Anastomosing ductile shear zones are widely developed in mafic granulites derived from gabbronorites intruding within or beneath the Kohistan crust. The ductile shear zones with a width of a few millimeters to several meters are strongly localized and clearly cut the gabbronorites with granulite assemblage (Fig. 4 a; b). Modal composition of hydrous minerals such as amphibole (in particular hornblende) and epidote in the ductile shear zone is distinctly higher than in that of the protolith. Difference in mineral assemblage between the host rock and the shear zone suggests that retrogression of metabasic rocks underwent hydration reactions like pyroxene + plagioclase + H_2O = hornblende + quartz. The increment of hydrous minerals in the ductile shear zones would require the infiltration of aqueous fluid, because the surrounding rocks preserve anhydrous assemblage. The preferential hydration within the ductile shear zone promotes ductile deformation, and can produce overwhelmingly higher permeability than host rocks that were undergone relatively hydrostatic conditions (Dipple and Ferry, 1990).

The estimated temperatures (600–750℃) of the amphibolites in the ductile shear zones are systematically lower than those of the mafic granulites (~800℃), and the estimated pressures (0.9–1.0 GPa) of the amphibolites are similar to those of the mafic granulites (Yoshino, 1998; Yoshino *et al.*, 1998). It indicates that the shearing event occurred during cooling under nearly constant pressure conditions, *i. e.* isobaric cooling after peak metamorphism of the mafic granulites. Therefore, the shear zones occurred during the intra-arc stage. Consequently, the localized, hydrated ductile shear zone has a possibility to generate seismic reflections and/or electrical conductivity anomalies in the lower crust in a tectonically active environment.

4. 2. Cracks

An alternative significant fluid pathway is a crack. There are abundant veins filled with mineral deposits, composed of mostly quartz and plagioclase,

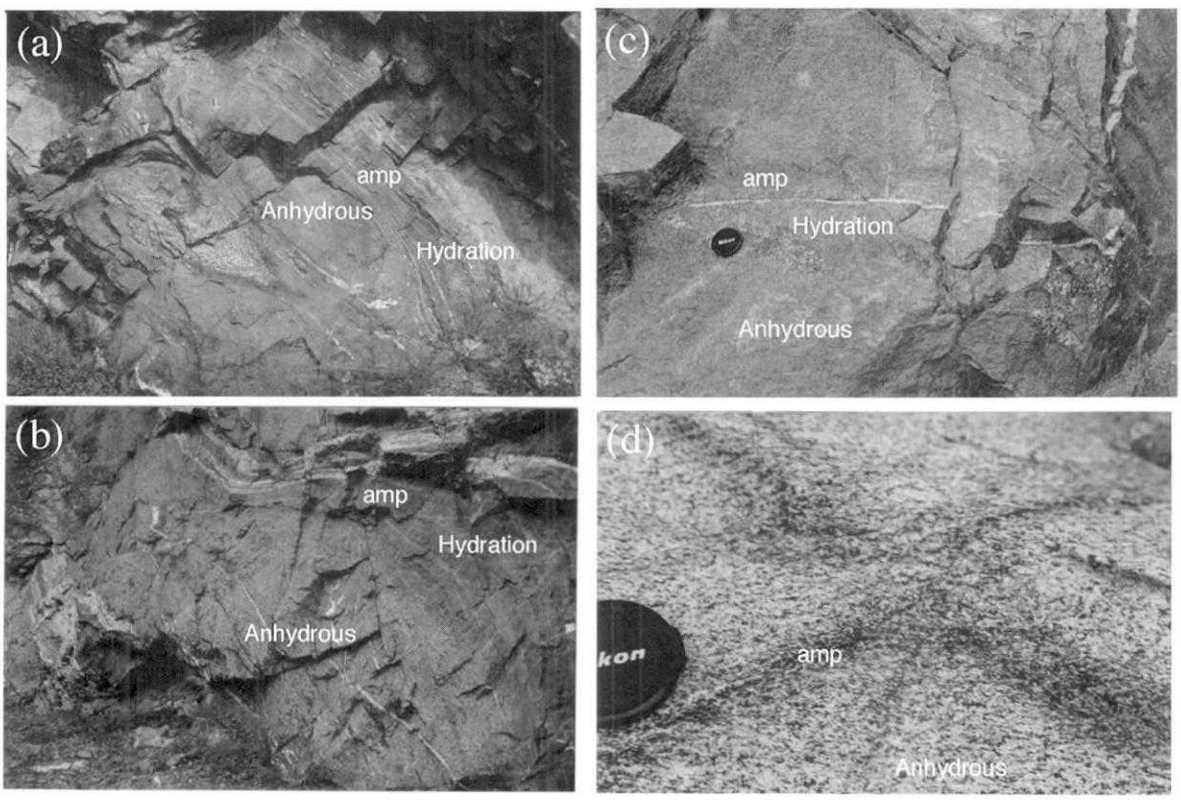

Fig. 4. Traces of water passage in the lower crust of the Kohistan Arc, NW Himalayas. (a & b) Field photograph showing that ductile shear zones consisting of amphibolites cut anhydrous gabbroic rocks. (c) Field photograph showing that hydrous reaction zones (amphibolite) occur along a quartz-feldspathic vein in anhydrous granulites. (d) Field photograph showing that hydrous reaction zones in anhydrous granulites occur without a quartz-feldspathic vein. Abbreviations: Amp, amphibolite.

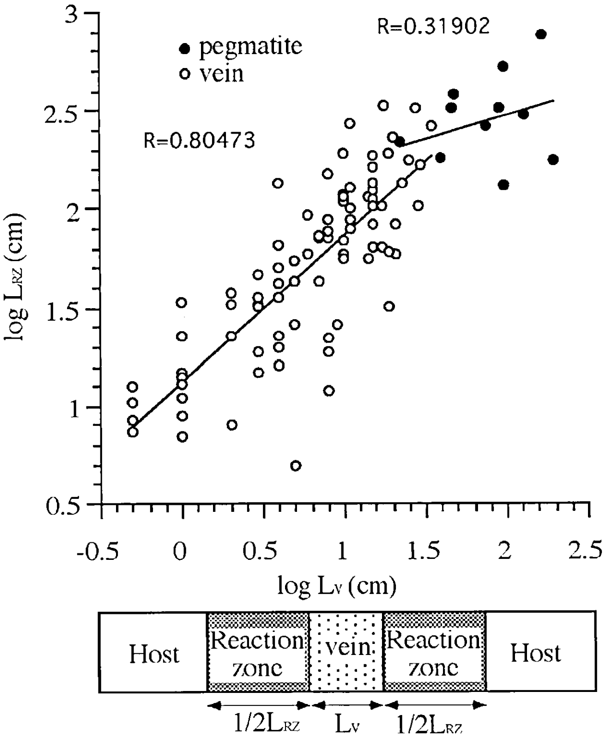

Fig. 5. Geometrical relationships between the amphibole reaction zone and the quartzo-feldspathic vein. Log total width of reaction zone vs. log width of vein. Note the linear correlation between them except for the pegmatite pair.

occasionally garnet and hornblende, in the lower crustal section of the Kohistan Arc. Anhydrous metagabbros are clearly cut by the veins in association with amphibole-bearing hydrous zone. The hydrous zone develops symmetrically on both side of the veins (Fig. 4 c). Width of the vein varies from a few millimeters to over 1 meter. The hydrous zone developed along the vein also have variable thickness and its width is proportional to the thickness of the veins (Fig. 5). Width of a vein increases with the increase of the reaction zone. In weakly hydrated areas, the network of the subplanar amphibole-bearing hydrous zone about 1-3 mm wide develops without a vein (Fig. 4 d).

In the hydrous zone, pyroxene is partially replaced by hornblende and quartz aggregates. Both clinopyroxene and orthopyroxene are commonly altered to fine Al-poor amphibole laths with quartz blebs clouded with magnetite dust changing to the brownish hornblende toward the adjacent plagioclase grains. The textures signify that hornblende associated with quartz are produced by the hydration reaction of pyroxenes and plagioclase. Formation of hydrous minerals would require the infiltration of aqueous fluid as well like in the case of the ductile shear zones.

The hydrous zones would be formed by the migration of aqueous fluid from a crack filled with fluid to the host rocks. Positive correlation of the width between the reaction zone and the vein filled with mineral deposits suggests that a thicker crack has a longer residence time. Thickness of reaction zones could be formed during geologically short time intervals (less than a year) (Etheridge *et al.*, 1984). The time interval is consistent with the time scale where isolated fluid-filled cracks could propagate quickly upward due to buoyancy forces (Nakashima, 1995). Therefore, cracks cannot hold out to trap water for geological time scale. If a cracking event produces the geophysical anomalies, it requires a large steady amount of aqueous fluid from depth to preclude crack healing due to the decrease of fluid pressure.

5. Discussion: Role of water-saturated ductile shear zones in geophysical anomalies

Aqueous fluid seems to be a strong candidate for generating both seismic reflections and high electrical conductivity in the lower crust. However, thermodynamic analyses of lower crustal rocks demonstrate that water fugacity has a value of tens to hundreds of bars (Lamb and Valley, 1988). This implies a fluid-absent condition or the presence of H_2O-poor fluid for typical igneous or high-grade metamorphic rocks now occurring in the lower crust. The petrological considerations are based on observations where predominant lithology of the exposed lower crustal column or xenolith is anhydrous granulite. Alternatively, effects of the localized hydrous zone on geophysical anomalies has been underestimated or neglected. On the other hand, seismic lamination in the lower crust corresponds to a heterogeneous structure showing different physical properties from the surrounding anhydrous rocks. In addition the resolution of electrical conductivity measurements cannot distinguish between small-scale conductive structure like narrow hydrated shear zones or cracks and large-scale resistive structures. Therefore the presence of a free aqueous fluid phase in the lower crust should be reconsidered in terms of the localized heterogeneity of the crust.

The present paper will propose a hypothesis that water-saturated ductile shear zones are the cause of conductive anomalies and seismic reflectors in the lower crust. The following three main observations support the hypothesis. Firstly, tectonically active zones where large fluid flux and ductile deformation can be expected, represent strong correlations between conductive anomalies and seismic reflection. Secondary, a non-hydrostatic condition is needed to establish connectivity of aqueous fluids to produce conductive anomalies, since wetting properties of plagioclase aggregate against C-H-O fluid phase show high dihedral angles over 60° under lower crustal *P-T* conditions (Yoshino *et al.*, in press). Finally, the preferential hydration in ductile shear zone is commonly observed in exposed lower crustal sections (e.g., Beach, 1980; Rutter and Brodie, 1985) as well as the Kohistan Arc.

5. 1. Stability in the hydrous phase

Preferential hydration within the ductile shear zone occurs under the sufficiently high activity of H_2O (a_{H_2O}) in comparison with conditions forming granulite assemblage. Figure 6 shows stability of the hydrous assemblage in simple two end-members CaO-MgO-Al_2O_3-SiO_2-H_2O (CMASH) and CaO-FeO-Al_2O_3-SiO_2-H_2O (CFASH) systems analogous to basic rocks. Assuming that metamorphic *P-T* conditions between rocks in ductile shear zones and surrounding rocks are not very different, a_{H_2O} is only an important parameter to form hydrous assemblage. If hydration occurred at 600℃ and 800 MPa, amphibole in the CMASH system becomes stable when log a_{H_2O} exceeds -2 (Fig. 6 c). In addition to the ferrous component of the system (Fig. 6 d), stability of the hydrous assemblage further restricts in the higher log a_{H_2O}. If H_2O-rich fluids in intergranular isolated pores were present, both rocks in the ductile shear zone and the surrounding anhydrous rocks should undergo a hydration reaction. Therefore, water in the ductile shear zone would not be derived from isolated fluid in intergranular pores, but from the external source. Thus fluid infiltration along the ductile shear zone is required. The preservation of anhydrous assemblage in the surrounding undeformed rocks implies that the undeformed rocks have an extremely lower permeability and little aqueous fluid phase before shearing. The infiltrating fluid was not totally consumed because pyroxene, in most of the hydrous ductile shear zone, is completely consumed. Hence the H_2O-buffer capacity in the ductile shear zones was insufficient. The remaining aqueous fluid can move through the ductile shear zone to the shallower depth due to buoyancy force.

5. 2. Salinity enhancement of upwelling fluids

The primary solute in aqueous fluid is generally NaCl at lower crustal conditions. Hypersaline fluid is an attractive candidate as a principal cause of deep conductivity. The average concentration equal to that of seawater (0.5 M) has been assumed to estimate porosity in the analysis of measured electrical conductivity (Hyndman and Shearer, 1989). However, conductivity of brine increases with increasing salinity (Quist and Marshall, 1968), and it is needed to constrain the change of salinity during metamorphic reactions and absolute concentrations of NaCl for lower crustal fluids. There are two mechanisms to change salinity in the deep fluids.

Firstly initial fluid composition evolves due to the progress of hydration. Markl and Bucher (1998) reported a solid salt solution (NaCl-KCl) together with chlorine-rich amphibole and biotite in lower crustal granulites at the Lofoten Islands in northern Norway. The Cl content of the amphiboles steadily increases toward the margin of grain. They interpreted that initial Cl-poor fluids evolve during their reaction with the granulite facies mineral assemblage to fluids of high salinity within the halite stability field. Then the fluid will successively become more Cl-rich until it finally reaches saturation with the halite (NaCl)-sylvite (KCl) solid solution because of the preferential incorporation of H_2O compared to Cl into hydrous minerals such as amphibole and biotite (Fig. 7 a). If ductile shear zones have insufficient capacity for H_2O, salinity in the upwelling fluid phase would be enhanced sufficiently without the precipitation of a solid salt solution.

Secondary salinity in fluid can change due to the fluid immiscibility as pressure and temperature change. Fluid inclusions in some of granulites are hypersaline brines, coexisting with dense CO_2-rich fluid inclusions of the same generation (e.g., Touret, 1971). Phase equilibria in the system CO_2-H_2O-NaCl predicts fluid immiscibility, with the coexistence of brines and CO_2-rich fluids, to high temperatures in salt-rich system in the deep crust (Skippen and Trommsdorff, 1986; Schmulnovich and Graham,

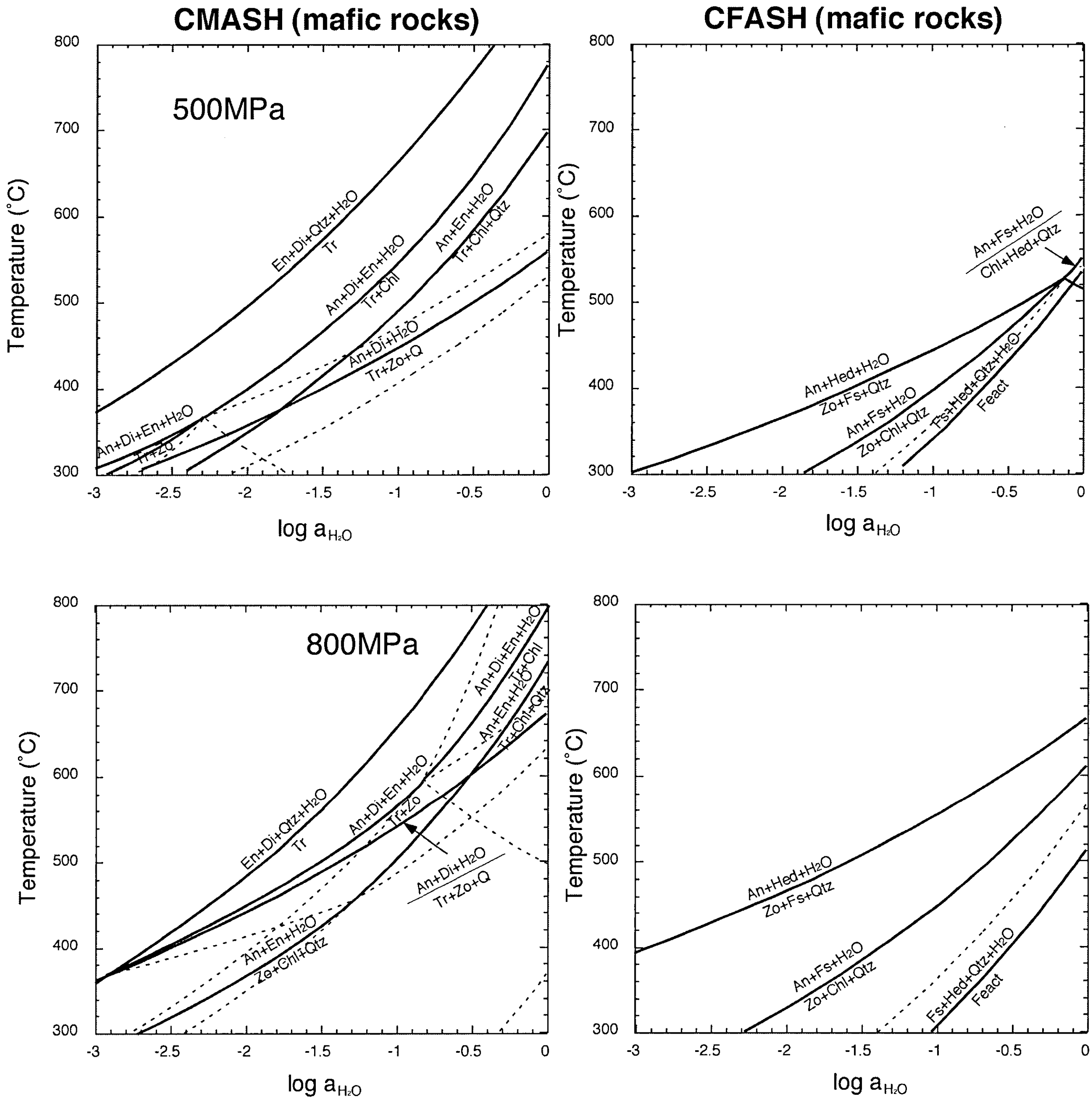

Fig. 6. Phase diagrams for water activity buffered by coexisting minerals of varying degrees of hydration as a function of temperature for constant pressure based on thermodynamic calculations (Holland and Powell, 1998). Idealized Mg end member of mafic rock system CaO-MgO-Al_2O_3-SiO_2-H_2O at 0.5 GPa (a) and 0.8 GPa (c). Idealized Fe end member system CaO-FeO-Al_2O_3-SiO_2-H_2O at 0.5 GPa (b) and 0.8 GPa (d). They provide an extreme case of the effect of iron on the buffering of water activity. Thick lines represent hydration reaction curve and dashed lines show reactions where both products and reactants include hydrous minerals. Shaded regions show typical temperature range for tectonic active areas. Abbreviations are An, anorthite; Chl, chlorite; Di, diopside; En, enstatite; Feact, Fe-actinolite; Fs, ferrosilite; Hed, hedenbergite; Ky, kyanite; Qtz, quartz; Tr, tremolite; Zo, clinozoisite.

1999). The immiscibility field in the system CO_2-H_2O-$NaCl$ decreases with increasing pressure and temperature. Therefore upwelling fluid that undergoes the decrease in temperature and pressure along the hydrated shear zone would reach the immiscible region. Then salinity in the aqueous fluids drastically increases compared with the initial miscible fluids or higher *P-T* unmixing fluids because NaCl concentrates in the aqueous fluids (Fig. 7 b). Even if infiltrated fluids did not react with the mineral phases,

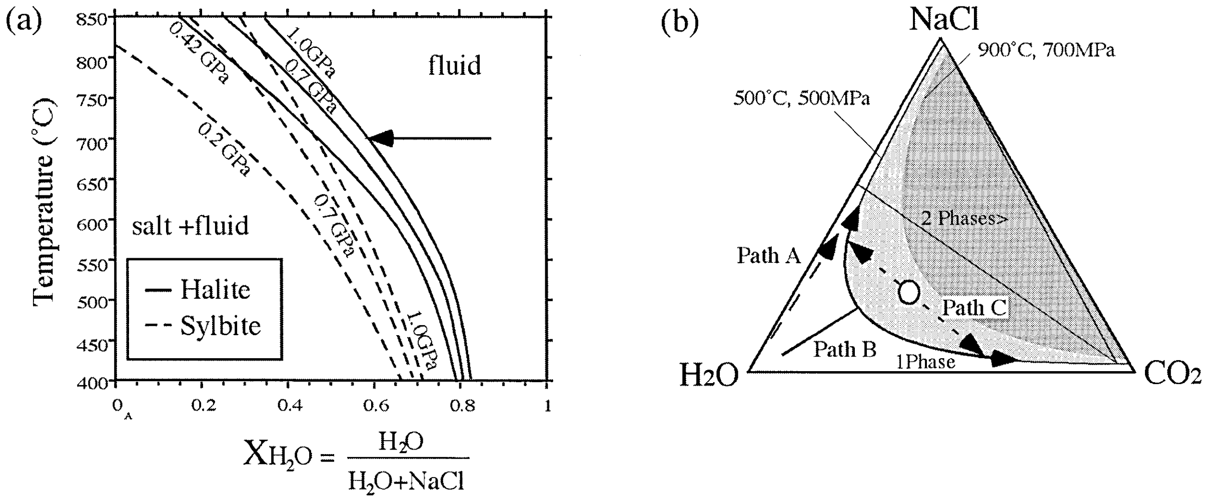

Fig. 7. (a) Temperature vs X_{H_2O} diagram of H_2O-NaCl fluids (modified after Markl and Bucher, 1998). Solid salt forms with decreasing X_{H_2O}, suggesting that the fluid from initially aqueous compositions forms a stable field of salt. Arrow shows isothermal path in association with salinity enhancement due to hydration. (b) Salinity change due to removal of H_2O in the system H_2O-CO_2-NaCl. Light and dark shaded regions are immiscible fields including more than 2 phases at 500℃ and 500 MPa and at 900℃ and 700 MPa, respectively (Shmulovich and Plyasunova, 1993; Gibert *et al.*, 1998). Path A: The X_{CO_2}/X_{NaCl} ratio of the initial fluid phase is relatively low. The fluid does not unmix during H_2O-buffered reaction, but evolves toward NaCl-rich compositions. Path B: The X_{CO_2}/X_{H_2O} ratio of the initial fluid phase is relatively high. When X_{H_2O} is sufficiently lowered, the fluid will separate to a NaCl-rich hydrous fluid phase and a CO_2-rich fluid phase. Path C: The initial fluid phase is miscible under higher *P-T* condition. During decompression and cooling, the fluid develops into an immiscible field and salinity in the hydrous fluid phase increases.

salinity in the aqueous fluids increases, while the very contrasting wetting behavior of the two immiscible fluids may lead to the selective removal of the brine (Gibert *et al.*, 1998).

In summary, increase of salinity due to hydration and the expansion of the immiscible field should lead to the increase of electrical conductivity at the mid-crustal level. If infiltrated fluids were derived from depth, electrical conductivity should increase from the Moho to the middle crust due to the increase of salinity in the infiltrated fluid. High electrical conductivity obtained from the middle crust does not contradict the model where deep fluids rise to the middle crust with increasing salinity throughout the lower crustal column.

5. 3. Permeability of the ductile shear zone

Ductile shear zones deformed under high temperature-pressure conditions are often selective sites of metamorphic transformations involving hydration and metasomatism (Rutter and Brodie, 1985). In contrast, the surrounding, less or non-deformed rocks are unaltered. Therefore, selective passage of water through the ductile shear zone requires higher permeability than the surrounding anhydrous granulites. Permeabilities of crystalline lower crustal rocks to H_2O has been measured in laboratory studies (Brace, 1980) and was found to lie within the range 10^{-18} to $10^{-23}\,m^2$ at room temperature. However, very little is known about the *in-situ* permeability of medium- to high-grade metamorphic rocks because of difficulty of the experiments. Therefore permeability of the lower crustal rocks has been estimated by fossil natural fluxes. For example, time integrated fluid fluxes have been calculated from the observed extent of progress seen in specific metamorphic reactions, especially in hydration or dehydration, for a wide range of lithologies (Ferry and Dipple, 1991). Time-integrated fluid fluxes can be combined with estimates of flow duration to constrain average flow rates and average *in-situ* permeabilities. For a fluid viscosity of $10^{-4}\,Pa\cdot s$ and a pressure gradient of $-1.5\,Pa/km$, estimated permeabilities of rocks in ductile shear zone are much higher ($10^{-15}\sim10^{-17}\,m^2$) than that of undeformed rocks during regional metamorphism, as calculated similarly from the progress of devolatilization reactions (Dipple and Ferry, 1992). These values correspond to a volume flux of $2.2\times10^4\,m^3/m^2$. The high permeability and fluid flux imply that the ductile shear zones behave as an effective water pathway in the lower crust.

The formation mechanism for the hydrous ductile shear zone is a complicated process. If fluid flow occurs synchronously with deformation in ductile shear zones, then infiltration has a significant effect on the rheology of altered shear zones (e.g. Beach, 1980; Rutter and Brodie, 1985). When a small amount of water is incorporated into the rocks, weakening of the rocks referred as hydrolytic weakening generally occurs (Griggs and Blacic, 1965). Thus deformation is likely to concentrate in the water-saturated ductile shear zones. On the other hand, metamorphic reactions are thought to enhance deformation rates by creating small strain-free grains, a process commonly referred to as reaction-enhanced ductility or reaction softening (e.g. Rubie, 1983). The shear strength of metasomatized rock would be weakened by grain-size reduction due to the nucleation or dynamic recrystallization. Further porosity should be created by the progress of the reactions during the infiltration of aqueous fluids since most hydration reactions have a positive Clapayron curve under lower crustal conditions. Increase of porosity in ductile shear zones would lead to the establishment of fluid connectivity and an increase of permeability. Although the rate of pore volume change is a trade-off process between porosity creation, due to the reaction, and porosity destruction, due to the plastic flow of constituent minerals, a sufficient fluid flux rate may maintain constant porosity and the connected pore geometry. These combinations can enhance the strain localization to generate the laminated seismic reflectors and conductivity anomalies in the lower crust.

Aqueous fluids migrate upward through the ductile shear zone with high permeability until reduced temperature results in hydration reactions and mineral precipitation, forming an impermeable layer that traps the fluid horizon below (Etheridge *et al.*, 1983). Such upward fluid migration with precipitation and impermeable layer formation in the middle crust would constrain the depth to the top of reflective and conductive zone (Frost and Bucher, 1994). The stored water beneath the impermeable layer can move episodically toward the upper crust when fluid pressure exceeds the fracture toughness of the brittle upper crustal rocks. This process may cause the seismic activity at the cut-off depth of earthquakes.

5.4. Fluid sources

If the ductile shear zones in the lower crust behave as pathways for water, the free fluid flux must be steadily supplied to keep the geophysical anomalies. Where does the free aqueous fluid in the hydrous shear zones come from? The tectonically active region showing a strong correlation between conductive anomalies and reflectors is the most favorable site where large fluid production can occur.

Two main sources of free aqueous fluid can be considered. One is water of magmatic origin. Crystallization of igneous melts may release a large volume of fluid that can penetrate the host rock. Magmatic underplating driven by basaltic magma derived from the mantle has been considered as an important process of growth of continental crust and granulite formation (Bohlen and Mezger, 1989). The basaltic magma intrudes at the lower crust or the Moho by gravitational instability due to the density contrast between the crustal rocks and basaltic magma (Herzberg *et al.*, 1983). This process can provide a large amount of aqueous fluid to the base of the crust, especially at subduction zone where the source region of basaltic magma is probably accompanied by water derived from subducting slab. This source can be available in the back-arc area from the volcanic front.

The other is water of dehydration origin directly derived from the subducting slab. The main dipping reflective and conductive zone beneath the northern Cascadia subduction zone at Vancouver Island is a typical example (Hyndman, 1988). The subducted oceanic crust, which initially contains ~6 wt.% water mainly in chlorite, lawsonite and amphibole, undergoes continuous dehydration and produces aqueous fluids of a total ~3 wt.% at the depth of 50 km (Iwamori, 1998). Thus large amounts of free water generated by dehydration reactions in the downgoing oceanic slab and sediments on it should be supplied to the above continental crust or accretion wedge. Although the fluid flux rate depends on the subduction rate, steady state fluid flux is assured as long as subduction continues.

6. Conclusion and future research

The present paper considers the relationship between electrical conductivity and seismic anomalies in the lower crust in terms of the presence of aqueous fluids. It will be clear from this review that we are far from a complete understanding of the interrelationships between those anomalies. Nonetheless, we can find an important feature. Only at the tectonically active zone such as convergent and divergent plate boundaries, results from magnetotelluric and multichannel seismic reflection studies show some correlations between conductive anomalies and seismic reflections in the continental lower crust, whereas the correlations are equivocal in the Precambrian crusts. This indicates that only in active areas of metamorphism or magmatism, fluids play a significant role to generate geophysical anomalies.

Experimentally determined wetting properties of plagioclase aggregate against C-H-O fluid phase show high dihedral angles over 60° under hydrostatic, lower crustal *P-T* conditions. The dihedral angles preclude formation of the three dimensional network of fluid phase in plagioclase-rich lower crustal rocks. Distribution of the aqueous fluid observed in the exposed lower crustal section of the Kohistan Arc, NW Himalayas is strongly restricted in heterogeneous structures related to hydration reactions such as vein or ductile shear zone. These observations suggest that fluid flow in the lower crust is channelized under non-hydrostatic conditions.

The hydration of a localized ductile shear zone is a strong candidate for the cause of electrical and seismic anomalies in the lower crust. This hypothesis is consistent with both petrological and geophysical observations. The presence of a hydrous ductile shear zone can produce the geophysical anomalies even if anhydrous rocks were dominant in the lower crust. General features of retrogressive ductile shear zones in the high-grade terrains suggest that physical anomalies in the lower crust for tectonically active zones, where aqueous fluids are steadily supplied, results from fluid distribution restricted in a narrow zone.

The effect of a water-saturated ductile shear zone on geophysical anomalies remains an outstanding problem for experimental investigations and a target for future incorporation into theoretical models of fluid-bearing rock systems in the lower crust. The water-saturated rock properties under lower crustal conditions have been undetermined yet, because of experimental difficulties of fluid-bearing systems. It is essential to investigate experimentally connectivity and pore topology of the fluid phase in silicate matrix under non-hydrostatic conditions. Moreover *in situ* measurements of electrical conductivity and permeability during high temperature creep can provide powerful constrains for the fluid flow and fluid connectivity in the ductile shear zone of the lower crust. For these purposes, we must establish firstly the measurement method of permeabilities and conductivities of fluid-bearing rocks under high pressure. Especially, electrical conductivity measurements in the laboratory under high pressure-temperature conditions have some difficulties such as the limits of sample size, control of oxygen fugacity and conductance of pressure media even if experiments were conducted in a fluid-free system (Fujita, 1999). In a fluid-saturated system, a sample must be sealed with a rare conductive metal jacket or capsule since water escapes toward the outside of the sample during experimentation. The most significant problem is insulation between the electrode and the capsule. Although Glover *et al.* (1990) used a guard-ring electrode arrangement to remove leaking electrical currents caused by the sleeving, the method is inevitable to perform removal together with the electric current through the sample. Therefore, development of the cell to seal the metal sleeve is an urgent need to measure bulk conductivity in a fluid-saturated system.

Geological observations at the exposed lower crustal section may provide information on fluid flow or storage of lower crust and provide a new insight into the generation mechanism of earthquakes. Although time-integrated fluid flux in ductile shear zones has been calculated by stable isotope studies and degree of reaction progress (Dipple and Ferry, 1992), the problems as to whether the fluid flow is continuous or episodic remains because of a lack of time constraint. It affects overwhelmingly the estimated permeability in the ductile shear zones. The time scales of deformation in localized shear zones are poorly known, are those of the associated strain rates. Muller *et al.* (2000) has determined the longevity and rates of deformation using rubidium-strontium (Rb-Sr) microsampling the dating of increments of fibrous strain fringes from a Pyrenean shear zone. Such microsampling dating of amphibole or mica forming the foliations may be a powerful tool to constrain the fluid flow rate and *in situ* permeability during deformation.

In the geophysical studies, it is necessary to carry out the combined analysis of magnetotelluric measurements, tomographic velocity models and deep seismic reflection images at the specified area. Then we can more accurately constrain the presence of fluids in the lower crust. In particular, the recent development of seismic tomographic studies is astonishing. For example, using the seismic tomographic method, Zhao *et al.* (1996) showed existence of a low seismic velocity (−5%) and a high Poisson's ratio (+6%) anomaly covering about 300 square kilometers at the hypocenter of the 17 January 1995, magnitude 7.2 Kobe earthquake in Japan. This anomaly was interpreted as an over-pressurized, fluid-filled, fractured rock matrix that contributed to the initiation of the Kobe earthquake. High-resolution seismic tomographic studies including P- and S-wave velocities will help to provide quantitative estimates of fluid volume and of pore topology in the lower crust.

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