Water transportation from the subducting slab into the mantle transition zone

Shigenori Maruyama a,⁎, Kazuaki Okamoto b

a Department of Earth and Planetary Sciences, Tokyo Institute of Technology, O-okayama 2-12-1, Meguroku, Tokyo 152, Japan
b Institute of Earth Sciences, Academia Sinica, P.O. Box 120, Academic Road Sec. 2, Nankang, Taipei, Taiwan, ROC

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Abstract

Using a recently developed petrogenetic grid for MORB+H2O, we propose a new model for the transportation of water from the subducting slab into the mantle transition zone. Depending on the geothermal gradient, two contrasting water-transportation mechanisms operate at depth in a subduction zone. If the geothermal gradient is low, lawsonite carries H2O into mantle depths of 300 km; with further subduction down to the mantle transition depth (approximately 400 km) lawsonite is no longer stable and thereafter H2O is once migrated upward to the mantle wedge then again carried down to the transition zone due to the induced convection. At this depth, hydrous β-phase olivine is stable and plays a role as a huge water reservoir. In contrast, if the geothermal gradient is high, the subducted slab may melt at 700–900 °C at depths shallower than 80 km to form felsic melt, into which water is dissolved. In this case, H2O cannot be transported into the mantle below 80 km. Between these two end-member mechanisms, two intermediate types are present. In the high-pressure intermediate type, the hydrous phase A plays an important role to carry water into the mantle transition zone. Water liberated by the lawsonite-consuming continuous reaction moves upward to form hydrous phase A in the hanging wall, which transports water into deeper mantle. This is due to a unique character of the reaction, because Phase A can become stable through the hydration reaction of olivine. In the case of low-pressure intermediate type, the presence of a dry mantle wedge below 100 km acts as a barrier to prevent H2O from entering into deeper mantle.

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1. Introduction

The role of water in the genesis of magma in subduction zone environments has been widely recognized (e.g. Wyllie, 1988). The dehydration has been attributed either to direct melting of the slab (e.g. Nicolls and Ringwood, 1973; Wyllie and Sekine, 1982; Brophy and Marsh, 1986) or to the release of fluids to decrease in the melting temperature of the mantle wedge (e.g. Ringwood, 1975; Delany and Helgeson, 1978; Tatsumi et al., 1986).

Dehydration due to isobaric amphibole breakdown, traditionally taken as the amphibolite–eclogite transformation reaction, is believed to provide the H2O necessary to trigger partial melting of the overlying mantle wedge. However, several high-pressure hydrous minerals have been documented to be stable at coesite and diamond isograds in many ultrahigh-pressure metamorphic terranes (see reviews by e.g., Liou et al., 1994; Chopin and Sobolev, 1995; Schreyer, 1995). Recent experimental studies (Schreyer, 1988; Poli, 1993; Poli and Schmidt, 1995; Okamoto and Maruyama, 1999) show that hydrous phases such as lawsonite, chloritoid, staurolite, phengite and zoisite–clinozoisite are passive transporters of considerable amounts of water in subducting oceanic crust to depths greater than that of amphibole stability.

It is important to determine the precise dehydration depth in the subducting oceanic plate and the mechanism of water transportation into the deep mantle, because several recent experiments suggest that H2O can be stored in several hydrous phases (A, E, hydrous-β, and -γ) in the mantle transition zone (e.g. Inoue et al., 1995; Kawamoto et al., 1996; Smyth and Kawamoto, 1997). The aim of this study is to propose H2O transportation mechanisms into the mantle transition zone, combining the experimental results for phase diagrams of MORB+H2O and peridotite+H2O,
together with an additional data set on the $T$-distribution on the slab surface.

2. Phase diagram of MORB+$H_2O$ system

Okamoto and Maruyama (1999) defined the high-pressure stability limit of lawsonite in the MORB+$H_2O$ system (Fig. 1). The lawsonite–eclogite facies is bounded by the high-pressure stability limit of a lawsonite–involving a continuous reaction defined by the compositional enlargement of garnet towards a grossular-rich end member with increasing pressure and/or temperature. Lawsonite in metabasite is stable up to 10 GPa at temperatures lower than 900 °C.

The so-called eclogite facies is subdivided into five facies. The dry eclogite facies has the largest $P$–$T$ space and is bounded by the jadeite eclogite and garnetite facies at $P$ higher than 16 GPa (Iriume et al., 1986; Okamoto and Maruyama, 2004) and the dry solidus on the high-$T$ side. The low-pressure limit of lawsonite-bearing eclogite facies is bounded by the zoisite eclogite subfacies involving a reaction with a gentle positive slope at 2–3 GPa (Poli and Schmidt, 1995). The low-pressure and low-temperature boundary is separated by a gentle reaction curve with a negative Clapeyron slope from the blueschist facies (Maruyama et al., 1996; Maruyama, 1997). The amphibole–eclogite facies is stable below the zoisite–eclogite facies, down to 1.0 GPa. Both the zoisite– and amphibole–eclogite facies are

Fig. 1. A phase diagram of the MORB+$H_2O$ system (Modified after Okamoto and Maruyama, 1999, 2004). Thin solid lines delineate polymorphic transformation of (1) quartz/coesite (Bohlen and Boettcher, 1982), (2) coesite/stishovite (Zhang et al., 1996), and (3) diamond/graphite (Bundy, 1980), and (4) an univariant reaction, albite=jadeite+quartz (modified from Holland, 1983, see Maruyama et al., 1996). The pale-colored shadow area delineates the stability field of lawsonite eclogite in the MORB+$H_2O$ system (Poli and Schmidt, 1995; Okamoto and Maruyama, 1999). Dark-colored shadow area delineates the slab melting region in the MORB+$H_2O$ system. Facies boundaries below 4 GPa are after Maruyama et al. (1996). Abbreviations; PA=pumpellyite–actinolite facies, GS=greenschist facies, EA=epidote–amphibolite facies, AM=amphibolite facies, GR=granulite facies, HGR=high-pressure granulite facies, BS=blueschist facies, Am Ec=amphibole eclogite facies, Zo Ec=zoisite eclogite facies.
terminated by wet solidus at \( T \) higher than 650 °C. The limits of metamorphic facies below 2.0 GPa are after Maruyama et al. (1996) and Maruyama (1997).

3. Fate of water from mid-oceanic ridge to subduction zone

3.1. The average \( H_2O \) content of hydrated oceanic crust

New oceanic crust is born at mid-oceanic ridges, and suffers hydrothermal fluid-alteration (“ocean-floor metamorphism” of Miyashiro, 1973). The crust generated at a mid-oceanic ridge is about 7 km thick, and the top 800 m is hydrated by reacting with the circulating hydrothermal water (Fig. 2). Beneath the 800 m depth, the crust is essentially dry except along fracture zones. The \( H_2O \) added to the oceanic crust by hydrothermal metamorphism is difficult to estimate precisely, simply because it depends on the degree of development of hydrothermal fracture systems at the mid-oceanic ridge, which we cannot directly observe. The \( H_2O \) is fixed as loosely bounded water in sheet silicates or pore space and structural water in the crystalline phases. The measured structural water from the DSDP hole 504B (Alt et al., 1986) is approximately 1 to 2 wt.% from the top of the pillow lava down to 600 m. However, it increases to 4–6 wt.% from 600 down to 800 m depth (the transition from the pillow to dike zone). The structural water of the dredged basaltic samples by multiple regression analysis also ranges from 1 to 8 wt.% (Hart, 1976). Staudigel et al. (1995) estimated the structural water at about 2.7 wt.% and the interstitial water is at least 2 wt.%, based on the void volume of the crust (approximately 5 vol.% by Johnson, 1979). The present study uses 5–6 wt.% \( H_2O^- \) which is stored in hydrous minerals due to ocean-floor hydrothermal metamorphism.

Dredged gabbros and ultramafic rocks from fracture zones are highly altered and suffered high-grade greenschist to amphibolite facies metamorphism. These rocks contain several kinds of hydrous minerals. However, such hydrated rocks are restricted to fracture zones (Fig. 2).

During transport of oceanic crust from the mid-oceanic ridge to the trench, no significant metamorphic and dynamic events occur between mid-oceanic ridge and subduction zone if OIB volcanism does not affect the MORB crust (Maruyama, 1991). Fluid trapped in the pore space of the subducting oceanic crust and overlying sediments is expelled within the 10–15 km depth range (e.g. Moore, 1992). Loosely bounded water (\( H_2O^- \)) is also released at shallow depths between 15 to 20 km. Subsequently, the subducted oceanic crust suffers subduction-zone

\[ \text{seawater addition mechanism} \]

Fig. 2. Water circulation system and plate tectonics.
metamorphism and continues to release H$_2$O, by dehydration reactions, to the overlying mantle wedge (Fig. 2). The change in H$_2$O content by dehydration reactions due to subduction zone metamorphism will be reviewed below.

3.2. The maximum H$_2$O contents in the MORB + H$_2$O system

The maximum H$_2$O content in the hydrated subducting oceanic crust can be calculated, based on the modal variation of hydrous minerals in metabasalt for given metamorphic facies (e.g. Peacock, 1993). The mineralogical changes in metabasalt as a function of $P$ and $T$ are best illustrated using the concept of metamorphic facies (Eskola, 1920), because most hydrous minerals in metabasalt break down by a complex series of continuous reactions. Peacock (1993) calculated the maximum H$_2$O content in each metamorphic facies, except for the eclogite facies, in the NCMASH system. He calculated a metamorphic mineral norm (that is similar to the CIPW norms) in order to estimate the maximum H$_2$O content in each of the metamorphic facies. He treated the eclogite facies as completely dry, and fate of hydrous minerals.

However, to estimate the amount of H$_2$O transported from the surface to the mantle by a subducting oceanic plate, both the blueschist–eclogite and the epidote amphibolite–eclogite transformations are of critical importance. The change in the amount of H$_2$O due to the blueschist – lawsonite–eclogite transformation was estimated to be 3 wt.% by Poli and Schmidt (1995) and that of the amphibolite–eclogite facies transformation to be 1.5 wt.% by Poli (1993). They calculated the mineral proportions employing a mass balance program which includes a Monte Carlo error propagation. In this study modal wt.% of each phase was calculated with the least square method for major elements. The modal wt.% of lawsonite at 8 GPa, 700 °C was estimated to 4.0 which corresponds to 0.5 wt.% of water in the hydrated MORB crust (Okamoto and Maruyama, 1999). The maximum H$_2$O content of hydrated MORB crust has been estimated by Peacock (1993), Poli (1993) and Poli and Schmidt (1995) to be below 6 GPa, and those are compiled in Fig. 3.
Based on the previous experimental studies, most workers considered that most H$_2$O in the subducting oceanic crust is released by the amphibolite–eclogite transformation at 60 km depth, and less than 1 wt.% H$_2$O is retained beneath the volcanic front (at ca. 100–120 km depth), if the temperature is below 650 °C (within the stability field of lawsonite–eclogite facies). However, if the temperature is above 650 °C at 30–90 km depths, subducting oceanic crust reaches the solidus temperature condition, causing the partial melting of amphibole-bearing metamorphosed oceanic crust. The released H$_2$O dissolves in the melt; then the remaining subducting oceanic crust transforms to dry eclogite. If the eclogite becomes anhydrous, it will not melt due to the high-$T$ solidus over than 1500 °C at about 200 km depth. The H$_2$O content in the subducting oceanic crust depends on the variation of $P$–$T$ path along the slab surface.

4. $T$-distribution on the slab surface

The fate of water in subducting oceanic lithosphere, thus, strongly depends on the $T$ distribution in the subducting slab. For an accurate estimate, it is critical to apply a multi-disciplinary approach, because numerical simulations cannot give a unique solution, because there are too many independent parameters to control the thermal structure along the slab surface. In the following, we first discuss the approach by numerical simulation, then summarize the field constraints, and finally define the predominant parameters which control the thermal structure.

4.1. Numerical simulation

Models presented over the past 25 years demonstrate that the thermal structure of a subduction zone depends on numerous parameters, including: (1) the thermal structure (age) of the incoming lithosphere, (2) the convergence rate, (3) the geometry of subduction, specifically subduction dip angle, (4) the distribution of radiogenic heat-producing elements, (5) the rate of shear heating along the subduction shear zone, and (6) the geometry and vigor of induced convection in the overlying mantle wedge (Fig. 4) (e.g. Oxburgh and Turcotte, 1970; Hasebe et al., 1970; Toksöz et al., 1971; Turcotte and Schubert, 1973, Anderson et al., 1978; Hsui and Toksöz, 1980; Honda and Uyeda, 1983; Wang and Shi, 1984; Cloos, 1985; Honda, 1985; van den Beukel and Wortel, 1988; Toksöz and Hsui, 1989; Peacock, 1990, 1991; Davies and Stevenson, 1992; Staudigel and King, 1992; Furukawa, 1993a,b; Peacock et al., 1994). Moreover, although poorly known, (7) the role of fluid dehydrated from the downgoing slab (Anderson et al., 1976, 1978; Delany and Helgeson, 1978), and (8) the exothermic or endothermic reaction within it would affect the thermal structure of the top of the downgoing slab.

Proposed $P$–$T$ paths for the oceanic crust at the top of the subducted slab differ sharply in the various models shown in the above papers. The results of selected numerical models are summarized by Peacock (1996). Although temperature distributions within the mantle wedge and in the subducted slab are relatively similar in all of these models, dramatic differences occur at the interface between the mantle wedge and the subducted slab. In the absence of shear heating, estimates of the temperature in the subduction shear zone at 100 km depth ranges from 300 to 750 °C. The lower calculated temperatures result from experiments in which the mantle wedge is assumed to be rigid. Qualitatively, induced convection in the mantle wedge brings warm mantle material into close proximity to the subducting slab; induced convection warms the subducting slab at the expense of cooling of the adjacent mantle wedge. Most models suggest that induced mantle–wedge convection heats the top of the slab by several hundred degrees. In the Peacock et al. (1994) estimation, induced convection increases the temperatures of the slab surface at 100 km depth from 150 °C to 450 °C (Fig. 5a).

In the absence of shear heating, faster convergence rates result in cooler subduction shear zone temperatures (Fig. 5b). Below 70 km depth and above 100 km depth, temperatures of the faster subducting slabs are lower than that of the slower subducting slabs. At 100 km depth, differing convergence rates do not cause a large temperature difference on the slab surface; $T=450$ °C if the $V=100$ mm/yr and $T=550$ °C if the $V=10$ mm/yr.

The thermal structure (age) of the oceanic lithosphere prior to subduction strongly influences the thermal structure of the subduction zone (Fig. 5c). If the age is 50 Myr, the temperature of the slab surface is approximately 450 °C at 100 km depth. If the age is less than 2 Myr, although the convergence rate is as fast as 100 mm/yr, subducting slab causes partial melting.
Temperature at 100 km depth is 750–1050 °C (Fig. 5c). If the age of the oceanic plate is older than 50 Myr and shear stress is negligible, the top of the subducting oceanic plate suffers blueschist facies metamorphism at depth range from 15 km to 60–70 km, and transforms to lawsonite–eclogite thereafter. The average age of oceanic plate is approximately 100 Ma on the modern Earth, and the oldest plate dates back to 200 Ma.

Shear heating, caused by shear stresses of the order of 100 MPa, dramatically increases calculated temperatures in the subduction shear zone to values in excess of 1000 °C at 100 km depth (e.g., Toksöz et al., 1971; Turcotte and Schubert, 1973) (Fig. 5d). Subduction-zone shear stress is poorly known. Recent estimates of shear stresses in subduction zones range from about 100 MPa (Honda, 1985; Scholz, 1990; Molnar and England, 1990) to several tens of MPa (Bird, 1978; van den Beukel and Wortel, 1988; Peacock, 1992; Titchelaar and Ruff, 1993) to approximately zero (Hyndman and Wang, 1993), and is negligible at depths greater than 50–60 km, where the subducting crust and the overlying mantle are mechanically coupled (Furukawa, 1993b). Furthermore, shear heating may be
absorbed by dehydration reactions taking place in the subducting slab (Anderson et al., 1976, 1977; Delany and Helgeson, 1978). Presence of H₂O also lowers rock strengths and frictional dissipation even lowers shear stress at depth.

Additional variables that lead to significant uncertainties in calculating the thermal structure of subduction zones include hydrothermal circulation in the oceanic lithosphere, the variation of thermal conductivity with T and P, the amount of heat consumed and released by metamorphic reactions, and the effects of fluid flow at shallow depths. Metamorphic dehydration reactions in the subducting slab may consume significant amounts of heat (Anderson et al., 1976, 1978; Delany and Helgeson, 1978).

Similarly, hydration reactions in the overlying mantle wedge may release significant amounts of heat (Peacock, 1987). Thermal calculations by Peacock (1990) that assumed 2 wt.% bound H₂O in the subducting oceanic crust indicate that metamorphic reactions have only a minor effect on the calculated thermal structure (<5 °C).

Summarizing the numerical simulation, the critical factors to control the thermal structure of the subducting lithosphere are the age of the lithosphere, the convergence rate and the induced mantle convection. Among these factors, the age the lithosphere has systematic spatial variation in the present- and paleo-subduction zones in circum-Pacific region. Shear stress is difficult
to estimate, but several studies show that it is less than several tens of MPa, giving a negligible effect at depth larger than 50 km. Thus we further argue the $P$–$T$ conditions in various subduction zones reconstructed from the age of the lithosphere.

4.2. Field constraints

4.2.1. (1) Adakites from slab melting of MORB

The trace element and REE patterns of adakites and high-Al Archean TTG suites suggest that they are derived from partial melting of a garnet-bearing MORB source (Kay, 1978; Martin, 1986). Recently, several researchers have described adakites in modern-arc environments where $<25$ Ma oceanic crust is currently being subducted (Fig. 6a; Drummond and Defant, 1990). These recent volcanics crop out between the main calc–alkaline arc and the trench, do not have typical arc geochemical signatures, and may represent direct partial melts of subducting oceanic crust. If the source of these volcanics is direct partial melting of the downgoing slab, then these magmas provide a powerful constraint on the thermal structure of subduction zones. The slab surface of the subducting lithosphere younger than 25 Myr may reach temperatures over 650 °C at 50 km depth or further; that is, when it reaches the wet solidus in the high-$P$ amphibolite and granulite facies (Fig. 6b). Applying the convergence rate and the age of the subducted lithosphere beneath southermost Chile and Mt. St Helens suites, $P$–$T$ paths for the top of the subducted lithosphere were estimated (Peacock et al., 1994). The estimated $P$–$T$ paths (condition satisfying both induced mantle wedge convection- and shear heating-free) also reach the wet solidus.

4.2.2. (2) $T$-estimates based on the depth range of the seismogenic zone of subduction thrust faults

Subduction zone faults generate earthquakes over a limited depth range. Hyndman et al. (1997) demonstrated that the updip (upper limit) and the down dip (lower limit) of the seismogenic zone are thermally controlled by the slab surface. The subduction zone faults are aseismic in their seaward updip portions and landward down dip of a critical point (Fig. 7; see above). The seaward shallow aseismic zone, commonly beneath accreted sediments, may be a consequence of unconsolidated sediments, especially stable-sliding smectite clays. If the clays are dehydrated and transformed to illite, then the fault may become seismogenic when the temperature reaches 100–150 °C (e.g. Hower et al., 1976), at 5–15 km depth (Fig. 7; see above). The downdip seismogenic limit for subduction of young, hot oceanic lithosphere is also temperature-controlled. The maximum temperature for seismic behavior in crustal rocks is 350 °C (brittle–ductile transition). For subduction beneath thin island arc crust and beneath continental crust in some areas, the forearc mantle is attached to the subducting oceanic crust at temperatures higher than the 350 °C. In such a case, the effect of the serpentinization is avoided. Using the determined seismogenic zone from Cascadia and Southwest Japan, where very young and hot plates are subducting, and from Alaska and most of Chile, where the forearc mantle is reached in the crust, updip and downdip temperatures were determined (see Hyndman et al., 1997; Oleskevich et al., 1999) (Fig. 7, below).

For Cascadia and SW Japan, melting $P$–$T$ conditions of the subducting slab causing production of the adakite suites are also plotted. The $P$–$T$ path for SW Japan is higher $P$ and lower $T$ than that for Cascadia. This may be due to the greater age of the subducting lithosphere of the westward descending Pacific plate. The $P$–$T$ paths for Chile (North Chile, Taltal, Coquimbo, and Valparaiso) and for Southern Alaska are steeper than those of Cascadia and SW Japan, showing that they are due to the older the subducting slab, and the steeper the geotherm along the Benioff thrust. However, the slab descending beneath South Chile, with the fastest subducting speed (9 cm/yr) of the present examples, has the steeper gradient in spite of being younger age (5 Ma) than the Cascadia (6.5–8 Ma) and the SW Japan (0–15 Ma) analogue. The Valparaiso slab (9 cm/yr) also has the steeper gradient in spite of being younger (37 Ma) than the North Chile, Taltal, and Coquimbo (45–50 Ma) and the South Alaska (50 Ma) plates. This tendency may be partly due to a higher convergence rate (9 cm/yr vs. 4–6 cm/yr). Generally speaking, all of these data support the idea that the age of the subducting slab dominantly controls the thermal structure on the Benioff plane although high convergent rate is recognized in some young slabs. Normal subduction-zone geotherm passes through the lawsonite–eclogite facies.

4.2.3. (3) $P$–$T$ conditions of HP and UHP metamorphic belts

Fossil geotherms in paleo-subduction zones are well-preserved in HP and UHP metamorphic belts all over the world (Maruyama et al., 1996). Using these fossil geotherms, radiometric and fossil ages, reconstructed paleo-plate geometries and convergence rates, we can estimate the dominant factors controlling the thermal structure on the slab surface in a few well-constrained metamorphic belts.

The fossil geotherm and age of the subducted oceanic crust are well reconstructed for the Cretaceous high-$P$–$T$ Sanbagawa belt, SW Japan (Maruyama and Seno, 1986; Izozaki and Maruyama, 1991; Maruyama, 1997). The reconstructed $P$–$T$ path is shown in Fig. 8. The metamorphic field gradients have been determined precisely by Banno and his co-workers for the last three decades (e.g. Banno, 1964; Higashino, 1975; Banno et al., 1978; Banno and Sakai, 1989; Higashino, 1990; Otsuki and Banno, 1990, etc). The facies series ranges from the pumppelyte–actinolite to transitional blueschist–green schist, green schist, epidote–amphibolite, and amphibolite facies, and is classified of the HP intermediate type (Miyashiro, 1961). Eclogite–facies assemblages occur locally in Fe$^{2+}$-rich high est-grade metabasites (Banno et al., 1978). Thus, the metamorphic grade ranges from 0.4–0.5 GPa and 250–300 °C for the lowest grade (Maruyama and Liou, 1983) to 1 GPa and 550–600 °C for the highest grade (Banno and Sakai, 1989).

The youngest fossils from the blueschist–facies rocks in the Sanbagawa belt are latest Jurassic to earliest Cretaceous radiolarians (Iwasaki et al., 1984). Radiometric ages range from 120 Ma to 60 Ma, depending on the distance from the root zone and metamorphic grade (Ityaya and Takasugi, 1988; compilation by Izozaki and Maruyama, 1991). $^{40}$Ar/$^{39}$Ar dates yield
Fig. 7. Subduction zone $P$–$T$ conditions deduced from the depth and temperature of the updip and the downdip seismogenic zone (see text for details). The updip and downdip temperatures are estimated by the temperatures of smectite–illite transformation and of the brittle–ductile transition of the crustal material, respectively (see Hyndman et al., 1997; Oleskevich et al., 1999).
the same results (Takasu and Dallmeyer, 1990; Takasu et al., 1994; Dallmeyer et al., 1995). Recent U–Pb SHRIMP age of the eclogite ranges 120–110 Ma as peak metamorphic age (Okamoto et al., 2004). Thus the ages of formation of the accretionary complex of the Sanbagawa protoliths and of metamorphism are very close to each other, at ca. 145–130 Ma and 120–60 Ma respectively, with a clustered metamorphic age peak at 80 Ma. The plate geometry from the Jurassic to the present in the NW Pacific Ocean margin was reconstructed by Engebretson et al. (1985). According to them, the Kula/Pacific mid-oceanic ridge subducted under NE Asia at about 80 Ma. Information recorded in the accretionary fold belt of the Shimanto belt indicates that the Kula/Pacific mid-oceanic ridge subducted underneath Japan at ca. 75 Ma (Taira, 1985), correlating the exhumation of the Sanbagawa high P–T belt from the 30 km depth, with approach of the mid-oceanic ridge to the subduction zone (Maruyama, 1997). Convergence rate of the plate was estimated to be approximately 202 mm/yr (Engebretson et al., 1985).

Another example is from the Alps (Fig. 8). The fossil geotherm of the subduction of the Piemont Oceanic crust beneath the Apulian plate is well-preserved in the HP–UHP metamorphic belt of the Western Alps (e.g. Chopin, 1984; Kienast et al., 1991, etc.). The metamorphic facies range from the pumpellyite–actinolite to blueschist, amphibole–eclogite facies, and zoisite–eclogite facies (see Maruyama et al., 1996). The Piemont oceanic plate was born at ca. 180 Ma (Jurassic to Triassic age; Channel et al., 1979), resulting from the break-up of Pangea and the separation of Laurasia+Gondwana to initiate the opening of the central Atlantic Ocean in the Early Jurassic. The metamorphic age of the UHP rocks was determined as ca. 35 Ma by Gebauer et al. (1993, 1997) by ion-probe dating (SHRIMP) of UHP zircons. The age of the subducting Piemont oceanic plate has been assumed to be 145 Ma. The convergence rate was estimated to be very low, approximately 0.2 mm/yr, according to the plate reconstruction of Dewey et al. (1989).
In the above contrasting examples (Fig. 8), an old slab with a very slow convergence rate vs. a young slab with an extremely fast convergence rate, clearly supports the idea that the age of the subducting oceanic plate is a critical parameter in determining the $P-T$ condition of the slab surface of the subducting oceanic plate. Applying this approach to estimate the $P-T$ path of the slab surface, we can roughly estimate the path in each respective subduction zone. The $P-T$ path beneath NE Japan is much colder than that of the Alps, because the age of the subducting oceanic plate is about 100 Myr with much faster subduction speed than that of Alps. The convergence rate is approximately 100 mm/yr, which is 50 times higher than that of Alps, indicating a path within lawsonite–eclogite facies.

5. Phase diagram of peridotite + water

Hydrated oceanic crust releases H$_2$O to the overlying mantle wedge under various $P-T$ conditions. The released H$_2$O should be mainly retained in hydrous phases; some is stored in anhydrous minerals (several tens ppm in olivine, garnet and several hundreds ppm in pyroxenes; see Bell and Rossman, 1992), in aqueous fluid phase, and in the melt (Thompson, 1992). The fate of released water is controlled by the stability relations of hydrous phases in the overlying mantle wedge. Thus to discuss the transportation mechanism of water into the deeper mantle, we need a phase diagram for peridotite+H$_2$O (Fig. 9).

Stability relations of hydrous minerals for ultramafic compositions in the mantle wedge has been determined recently by ultrahigh-pressure experiments (e.g. Yamamoto and Akimoto, 1977; Bose and Ganguly, 1995; Ulmer and Trommsdorff, 1995; Kawamoto et al., 1996; Kawamoto and Holloway, 1997). Phase relations of serpentine, chlorite, amphibole and others are shown schematically in Fig. 9. Amphibole, chlorite and serpentine are stable at depths shallower than 200 km; amphibole is stable below 3 GPa, 1000 °C, chlorite below 5 GPa, 800 °C and serpentine (antigorite) at temperatures lower than 500–600 °C below 6.5 GPa. Thus, with increasing pressure up to 6.5 GPa, maximum temperatures of hydrous mineral stability decreases from 1000 °C to 500–600 °C. With a further increase in pressure, hydrous silicates expand their stability fields towards higher temperatures. The hydrous minerals (e.g. phase A, E and hydrous-β, -γ) are candidates for H$_2$O containers in the mantle wedge (peridotite+H$_2$O system) above 6.5 GPa. The high-$T$ maximum stability limit of those phases shifts from 600 °C at 6.5 GPa to 1500 °C at 13 GPa. The hydrous β-olivine coexists with melt at temperatures higher than $T$=1100 °C and $P$=12 GPa (Fig. 9).

Fig. 9. (a) A phase diagram of the MORB+H$_2$O system and subduction zone geotherms. Open area: hydrous minerals are unstable, shaded areas: hydrous minerals are stable. Broken lines: phengite stability limit in pelitic rocks (D&H by Domanik and Holloway, 1996; S by Schmidt, 1996). (b) A phase diagram of peridotite+H$_2$O system (Kawamoto et al., 1996, 1997; Bose and Ganguly, 1995; Ulmer and Trommsdorf, 1995). Open area: hydrous minerals are unstable, shaded areas: hydrous minerals are stable. A, B and C; representative $P-T$ paths of the slab surfaces.
The water released from the subducting lithosphere is trapped by the overlying mantle wedge in which hydrous silicates are stable, if the geotherm is cold enough to stabilize hydrous phases. However if the geotherm is hot, the released H$_2$O is not stable in the hydrous phases in the overlying mantle. Instead, the H$_2$O would move upwards as porous flow, and cannot be transported into the deeper mantle through dragging down by the lithospheric counter-flow.

As already discussed above, the $P$–$T$ path of the slab surface varies, and is strongly dependent on the age of the subducting slab, and the convergence rate. Temperature conditions of the overlying mantle wedge depend also on the $T$ of the slab surface.

6. Mechanism of water transportation into the mantle transition zone

Combining the two phase diagrams in Fig. 9 in addition to Fig. 3, the water transportation mechanism can be estimated as follows. First, the following four different $T$ path cases along the slab plane (Fig. 9) are discussed: (A) steepest geotherm generated by old slab; (D) most gentle geotherm produced by very young slab; and intermediate types (B) and (C).

In the case of subduction of an old oceanic plate along a low geothermal gradient, the subducting oceanic crust suffers progressive high $P$–$T$ metamorphism from blueschist to lawsonite–eclogite facies (Fig. 9). Lawsonite–eclogite facies assemblages are stable over a wide $P$–$T$ field extending from 2.5 GPa, $T$<500 °C, through 8 GPa, $T$<800 °C to 9.5 GPa, <700 °C. That is, lawsonite is stable at depths ranging from 75 km to 300 km. As described above, lawsonite is continuously dehydrated with increasing temperature in the coesite field, and with increasing pressure and temperature in the stishovite field. Subduction of old oceanic plate (Path A) releases 3 wt.% H$_2$O continuously from 75 km to 100 km depth by the blueschist–lawsonite–eclogite transformation and retains 1 wt.% H$_2$O in the subducting slabs. Therefore, the small wedge corner triangle (trenchward from aseismic front) must be hydrated extensively with successive subduction of the oceanic plate with time (Fig. 10a). Intermediate-depth intraslab earthquake has been...
H2O can be carried to greater depths due to the induced convection of the mantle wedge. Below 270 km, phase E becomes the main H2O container, and is stable in the mantle transition zone. At greater depth, phase E is not stable, and H2O is stored in hydrous β-phase of olivine containing up to 3 wt.% H2O (Fig. 10b).

From the 150 to 240 km depth, the H2O released from the subducting oceanic crust by lawsonite-consuming reaction, migrates upward through the overlying mantle wedge if the fluid is dominated by H2O. The ascending aqueous fluid causes partial melting of the mantle wedge, and produces island-arc magma. However, the fluid dissolves considerable amounts of Si, Mg and other elements under ultrahigh-pressure conditions (Fujii et al., 1997). The addition of aqueous fluid may produce hydrous phases such as K-richerite and Ti-clinohumite in the mantle wedge.

A moderately old subducting oceanic plate may follow path C of Fig. 9. The P–T trajectory passes from the blueschist to zoisite–eclogite facies at 70 km depth. The transformation from zoisite–eclogite to dry eclogite occurs at 110 km, within the coesite stability field (Fig. 9a). Most of the H2O is leached out at 110 km depth when the slab is heated up to 650–700 °C. H2O content of the subducting slab decreases from 0.5 wt.% at 75 km depth to nearly 0 at 110 km depth. Released H2O could be stored in serpentine and chlorite of the overlying mantle wedge along the Benioff thrust. However, no hydrous mineral is stable at depths below 130 km in the mantle wedge right above the slab surface. Stored H2O in these hydrous minerals at shallow levels is migrated upward in the convected mantle wedge at 130 km. The ascending aqueous fluids cause the partial melting of the highest temperature region of the mantle wedge, to produce island-arc magma (Fig. 10c).

In the case of subduction of young oceanic crust (path D in Fig. 9), slab melting occurs at depths 30–70 km and the released H2O is stored in the melt ascending to the mantle wedge. Extensive melting of hydrated MORB crust forms calc–alkaline magma, in which all H2O is incorporated. Hence, H2O cannot be transported into the mantle wedge below 70 km depth. Once H2O is leached out of the subducting oceanic crust, it is difficult to produce melt again, because the melting temperature defined by the dry solidus of the MORB is higher than 1220°C at 70 km depth. The hydrated SiO2-rich felsic melt would react with α-olivine during the magma ascent into the overlying mantle, to form orthopyroxene as an interface curtain wall in the melt pathway (Fig. 10d; Maruyama, 1997).

7. Role of serpentinitized peridotites along transform faults on the oceanic floor

Oxygen isotope analyses of ophiolites (e.g. Gregory and Taylor, 1981) and direct drilling of oceanic crust at DSDP site 504B (Alt et al., 1986) indicate that hydrothermal circulation at mid-oceanic ridges is shallower than 800 m based on boiling study of the seawater detected at 504B, or at least less than 5 km, estimated by water/rock ratio by oxygen isotope on ophiolite. Maede and Jeanloz (1991) estimated the maximum thickness of the hydrated zone to be 12 km, which is far thinner
compared to the thickness of slab 600–100 km if it is older than 50 Myr. If this is true, major parts of oceanic peridotite under the MORB crust are predominantly dry, and are unrelated to the transportation of surface water into the deeper mantle.

However, it is well-known that serpentinized peridotites are exposed along transform faults on the ocean-floor (Fig. 2). Therefore, some parts of oceanic peridotite are undoubtedly hydrated and may transport surface water into the mantle transition zone, although it is difficult to estimate its role quantitatively. If the lower plane of the double seismic zone is representing the dehydration of the hydrated peridotite, the hydrous peridotite in the slab is not only part hydrated at the transform faults. Further speculative hypothesis has been proposed by Seno and Yamanaka (1996), therefore we do not argue further here. Thus, the presence of those hydrated peridotites promotes the water transportation into deeper mantle.

8. Role of phengite in subducting pelitic rocks

Capping the top of the oceanic lithosphere, the hemipelagic sediments and trench turbidites may subduct together with oceanic crust in deeper parts of a subduction zone, although most would be offloaded to form an accretionary complex at shallow depths on the hanging wall of the overlying plate. It is not certain whether they could subduct below Moho depth into the deep upper mantle, lie above the transition zone. The stability relations of hydrous phases in a pelitic system have been investigated on a wide $P–T$ range (Schmidt, 1996; Domanik and Holloway, 1996). Phengite–muscovite as a most common hydrous phase in pelitic rocks has a maximum $P$-stability up to 8–10 GPa and up to 1060 °C. Other accessory hydrous phases are topaz–OH, Mg–pumpellyite, lawsonite, and K–cymrite (Domanik and Holloway, 1996). Phengitic mica has a stability field 100–200 °C higher than the maximum stability limit of lawsonite in metabasite at $P=3–10$ GPa. Therefore, some amounts of water may be transported by pelitic rocks down to 300 km depth only if the geothermal gradients are low enough including A, B, and C in Fig. 9. Thus, pelitic rocks could, possibly help the transportation of $H_2O$ to great mantle depth.

9. Discussion

Before the advent of plate tectonics in Earth Science, most earth scientists never discussed the fate of surface water in terms of global material circulation on a whole mantle scale, because of the lack of critical petrologic observations, experimental data sets, and geophysical data on mantle circulation. However, the presence of phlogopite and K–amphibole in mantle xenoliths clearly indicate that water is present as structural $OH^-$ in mantle minerals. When plate tectonics appeared, to interpret subduction zone magmatism, most volcanologists inferred that calc–alkaline magmas are derived from slab-melting of hydrated MORB or partial melting of the mantle wedge, to which the released seawater is added to lower the melting $T$. Some petrologists considered that almost all dehydration occurs to generates arc magma at depths shallower than 150–200 km (Ernst, 1999), hence no $H_2O$ penetrates into the mantle transition zone with time. Therefore, the Earth’s mantle transition zone (410–660 km), if hydrated when the magma ocean had consolidated during Early Earth at 4.55 Ga, must have gradually dehydrated by plume activity through geologic time (Kawamoto et al., 1996). In contrast, there is an opposing idea that subduction zones deliver six times more water into the mantle ($8.7 \times 10^{11}$ kg yr$^{-1}$) than is delivered to the surface by arc volcanism ($1.4 \times 10^{11}$ kg yr$^{-1}$) (Peacock, 1990), implying the on-going water transportation into the mantle transition zone. If this is true, most subducted water is being swallowed into the mantle transition zone or even into the lower mantle at present, implying that the total amount of seawater in the oceans is now being reduced. From the standpoint of mantle mineralogy and petrology, dense hydrous magnesium silicates (DHMS) which are stable under mantle conditions, count more than 28 species (e.g. review by Thompson, 1992). For example, for upper mantle conditions, after the pioneering work by Ringwood and Major (1967), Yamamoto and Akimoto (1977) investigated the system MgO–SiO$_2$–H$_2$O at $P=2.9–7.7$ GPa and $T=470–1225$ °C, and found DHMS such as brucite, phase A, phase D, clinohumite, serpentine, talc, 10 Å phase, and chondrodite. Since then, superhydrous B (Pacalo and Parise, 1992; Burnley and Navrotsky, 1996), phase A (Bose and Ganguly, 1995; Pawley and Wood, 1996), and phase E (Luth, 1995) have been synthesized and analyzed. Akaogi and Akimoto (1980), Liu (1986, 1987), Kanazaki (1991), and Gasparik (1993) have investigated the phase relations near and in the mantle transition zone. Under lower mantle conditions, phase D (Li and Jeanloz, 1991; Yang et al., 1997; Shieh et al., 1998), phase F (Ohtani et al., 1995), and phase G (Ohtani et al., 1997; Kushiro et al., 1997) have been described as stable hydrous silicates if we presume the presence of water in the lower mantle.

Thus, a variety of hydrous silicates has been demonstrated as possible stable phases within the whole depth range of upper mantle and topmost parts of the lower mantle.

In spite of wide and high-$T$ stability fields of those DHMS in the deeper mantle, surface water cannot be simply transported into the deeper mantle, because (1) the subduction zone geotherm that they assumed is too warm, resulting in dehydration of all DHMS at relatively shallow depth ranges (see Fig. 9b), (2) subducting oceanic lithosphere is capped by 7 km thick MORB, and MORB cannot carry water into mantle deeper than 300 km. Although the core is the coldest part of the subducting slab, hydrous parts of the slab are less than the topmost 12 km (Maede and Jeanloz, 1991), or even less (probably less than 800 m as discussed before), and (3) water transportation into the mantle transition zone depends strongly on the subduction-zone geotherm as explained below. If it is warm, no water can enter into the deeper mantle. If the presence of limited lateral arrangement of arc volcanoes usually 120–200 km right above the slab surface, means that all DHMS dehydrate below 200 km depth, then a suggestion arises that no water enters into mantle depths greater than 200 km (Kawamoto et al., 1996).

To solve the debate, phase relations of MORB+$H_2O$ and peridotite+$H_2O$ under the whole upper mantle conditions, are
crucial. The summarized phase relations in Fig. 9b provide a concept regarding the fate of water. Two critical points at $X_1$ (6.5 GPa, 550 °C) and $X_2$ (5.5 GPa, 600 °C), are marked on Fig. 9b. $X_1$ corresponds to the lowest $T$ at which dry mineral assemblages are stable in mantle peridotite. From this point toward higher-$P$ and $-T$, the stability field of anhydrous mineral assemblages is spread over a wide range. Therefore, if the Benioff geotherm passes the lower-$T$ side of $X_1$, dragged downflow of the mantle wedge can carry the water within hydrous phases into the deeper mantle. If the Benioff geotherm passes on the higher-$T$ side of $X_2$, water cannot be transported into deeper mantle not only by downflow of the wedge mantle but also by hydrated MORB crust. If the Benioff geotherm passes between $X_1$ and $X_2$, lawsonite can transport the crystalline water down to 240 km depth (point $X_3$) to be relayed to the overlying downflow of mantle in which phase A is stable. Note a unique character of the reaction to form phase A connecting $X_1$ with $X_3$, because it is a hydration reaction forsterite + water = phase A (Pawley and Wood, 1996). If no water is supplied by the downgoing oceanic crust, the mantle wedge cannot be hydrated. In the case of the Benioff geotherm passing on the higher-$T$ side of $X_2$, water cannot be transported into the deeper mantle at all. All hydrous phases in downgoing MORB crust and in serpentinitized peridotites along transform faults would be dehydrated by 200 km depth, thereafter no water moves into the transition zone (Fig. 9b). Thus a small triangular area defined by $X_1$, $X_2$, $X_3$ is critical to control the fate of water in a subduction zone. The significance of the role of triangular zone in the $P$–$T$ space is portrayed in Fig. 10b and c where, the presence of a dry hanging wall peridotite prevents dehydrated water moving down as hydrous phases along the slab surface. Instead, free-water moves upwards to trigger the partial melting of peridotite.

Thus, the geothermal gradients along the slab surface determine whether or not the surface water can be transported into the deeper mantle. The variable $P$–$T$ gradients along the slab surface are a key to understanding the fate of water in subduction zones. If subduction zones do not follow $P$–$T$ paths like A or B, then we must support the interpretation by Kawamoto et al. (1996). But, as discussed before, if subduction zone geotherms are very variable and depend highly on the age of the slab, and follow paths such as A or B, we must support the idea by Peacock (1990). If the slab is older than 50 Myr, it would pass into the lawsonite stability fields or even much colder ($X_3$) where, the presence of a dry hanging wall peridotite prevents dehydrated water moving down as hydrous phases along the slab surface. Instead, free-water moves upwards to trigger the partial melting of peridotite.

Additional supporting observations are derived from tectono-metamorphic aspects of the mantle. Earthquakes deeper than 100 km, down to 650 km in the Earth may easily be explained if dehydration is on-going at those depths by subduction (Maede and Jeanloz, 1991; Stein, 1995). At those depths rocks are expected to deform by ductile flow rather than brittle fracturing or frictional sliding on fault surfaces (Maede and Jeanloz, 1991).

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References


