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CONDITIONS FOR MELTING AND METASOMATISM IN THE EARTH'S MANTLE

(Figs. 6)

Abstract: If we know the compositions of mantle rocks at various depths, and the geotherm in different tectonic environments, then the conditions for melting are defined by experimentally determined solidus curves. The term "metasomatism" in crustal processes is defined as reaction by solution or vapors, not by melts or magmas, and the same definition should apply to mantle processes (reaction with magmas is "hybridization"). H_2O , CO_2 , or both are made available for metasomatic reaction at deep dissociation fronts, or by solidification of volatile-charged magmas. Regions eligible for metasomatism are limited by solidus curves above which the melts dissolve volatile components. Beneath the lithosphere, there can be no metasomatism between about 120 and 260 km, because melting intervenes. Solidification of kimberlitic magmas at the base of continental lithosphere is a source of metasomatic fluids. Mantle metasomatism is expected in several regions above subducted oceanic lithosphere, interspersed with magmatic events. Major differentiation of the Earth is accomplished by melting, but metasomatism may cause significant redistribution of some elements.

Резюме: Если известны составы покрывающих пород в разных глубинах и геотерма в разных тектонических средах, потом условия плавления определяются экспериментально назначенными кривыми солида. Термин „метасоматизм“ в процессах коры определяется как реакция проходящая при помощи растворения или пар, не расплавов или магм, и такую же дефиницию можно применить для процессов мантии (реакцией с магмами является „гибридизация“). H_2O , CO_2 или оба соединения пригодны для метасоматических реакций в глубоких фронтах диссоциации или затвердеванием летучих магм. Районы подходящие для метасоматизма ограничены кривыми солида, над которыми расплавы растворяют летучие компоненты. Под литосферой не встречается метасоматизм между около 120 и 260 км вследствие плавления. Затвердевание кимберлитовых магм в базе континентальной литосферы является источником метасоматических флюидов. Мантийный метасоматизм ожидается в нескольких районах над поглощенной океанической литосферой перемешанной с магматическими событиями. Основная дифференциация земли проведена посредством плавления, но метасоматизм может вызвать значительное перераспределение некоторых элементов.

Introduction

The chemical differentiation of the Earth into mantle and crust is accomplished by magmatism. Magmas are generated when bodies of rock are transported across the depth-temperature limits of melting curves by: 1) physical convection; 2) temperature increase arising from tectonic conditions; or 3) depression of melting boundaries to lower temperatures by influx of volatile components. The conditions for melting in the mantle would be easily defined if we knew: 1) the compositions of mantle rocks at various depths; 2) the ex-

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perimentally determined solidus curves; and 3) geotherm in different tectonic environments. However, there remain many uncertainties in the precise definition of these three items.

The abundance of major and trace elements in many magmatic rocks are difficult to explain in petrogenetic schemes involving a mantle peridotite source. The concentrations of trace elements in magmas appear to be decoupled from those of major elements. It has become common practice in recent years to alleviate this problem by appealing to metasomatism caused by unspecified fluids, to adjust the mantle composition in order to satisfy the geochemical data (Walker, 1983). There is evidence that the mantle is heterogeneous, and that fluids have migrated through mantle rocks. Some investigators describe these fluids as dense, volatile-rich solutions, and others refer to silicate melts, perhaps similar in composition to kimberlites.

It is important in petrological discussions to retain the distinction between melt (liquid), vapor (or dense gas, solution), and the descriptive term "fluid" which can refer to either phase. Liquid and dense vapor are two distinct phases with different compositions and different properties. At high pressures and temperatures, the compositions and properties of these two fluid phases approach each other, but all experimental data so far available indicate that for normal rock composition with H_2O and CO_2 in the upper mantle, liquid and vapor maintain their separate identities. They can co-exist with each other. Liquids can exist only at temperatures above the solidus, whereas a vapor can exist through a wide range of temperatures below the solidus. Liquid may solidify and exsolve vapor. Vapor may cross a solidus boundary, and dissolve in liquid.

Metasomatism was defined originally for crustal processes, in which the compositions of rocks undergoing metamorphism were changed through the action of a mobile pore fluid, usually rich in super-critical H_2O , with CO_2 and CH_4 becoming significant in calcareous rocks. Changes in metamorphic rock compositions caused by addition of magmas, or by expulsion of melt from a rock undergoing anatexis, have not been described as metasomatic. There is a large vocabulary, including the terms migmatite or hybrid, to describe a rock formed by impregnation of magmas. If the nature of a fluid changing the composition of mantle rock can be identified as a silicate magma, then this is not a metasomatic change, unless the current definition of metasomatism in crustal rocks is extended.

Metasomatism of mantle requires the presence of volatile components. H_2O , CO_2 , or both are made available for metasomatic reactions in the upper mantle by: 1) release at deep dissociation fronts; 2) solidification at depth of volatile-charged magmas; or 3) upward migration of primordial volatiles.

Solidus curves for mantle materials

Candidates for upper mantle rocks include lherzolite, harzburgite, and eclogite. Fig. 1 illustrates solidus curves determined experimentally for synthetic pyrolite and four natural peridotites. The more refractory peridotites should have higher $MgO/(MgO+FeO)$. The extent of depletion or enrichment of the rocks in incompatible trace elements has little effect on the phase relationships. At pressures to 30 kbar, the solidus for quartz-free eclogite is barely distin-

guishable (within experimental limits) from that of undepleted lherzolite; the liquidus for dry basalt is not much higher in temperature. Quartz eclogite (tholeiite) begins to melt at temperatures significantly lower. None of the solidus curves in Fig. 1 was closely defined by good experimental brackets.

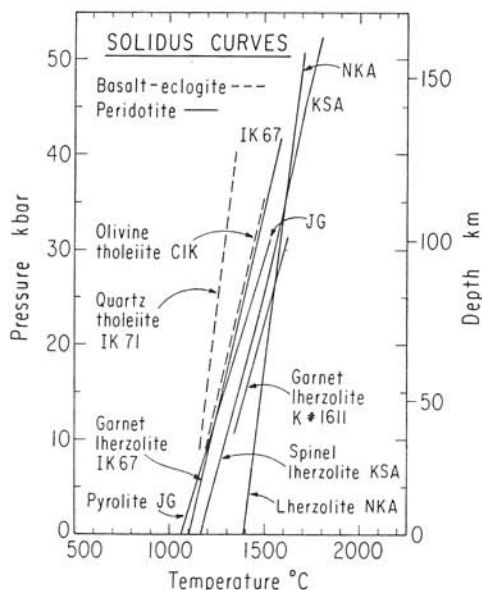


Fig. 1. Solidus curves measured for various peridotites, for olivine tholeiite (eclogite), and for quartz tholeiite (eclogite).

Abbreviations: IK, Ito—Kennedy, 1967; JG, Jaques—Green, 1980; K, Kushiro, 1973; KSA, Kushiro et al., 1968; NKA, Nishikawa et al., 1970; CIK, Cohen et al., 1967; IK, Ito—Kennedy, 1971.

Fig. 2 compares the original solidus of Kushiro et al. (1968) for a spinel lherzolite (KSA in Fig. 1) with a determination by Takahashi—Kushiro (1983). In the more detailed recent study, the solidus is lower by 70–100 °C. The results can be interpreted as representing three curves defining cusps at transitions from plagioclase lherzolite to aluminus-pyroxene lherzolite at about 11 kbar, and to garnet lherzolite at about 26 kbar, consistent with the results of Presnall et al. (1979) for model lherzolite in $\text{CaO—Mg—Al}_2\text{O}_3\text{—SiO}_2$.

The addition of H_2O and CO_2 to peridotite introduces hydrous minerals and carbonates, and lowers the solidus. Fig. 3A is a partly schematic phase diagram for peridotite with small amounts of CO_2 (5% by weight), H_2O (0.4%; 0.02%), and with $\text{CO}_2/(\text{H}_2\text{O}+\text{CO}_2) = 0.8$ (mole ratio). Subsolidus minerals represented are phlogopite, amphibole, dolomite, and magnesite. The solidus is the line PMQR, and the shaded area shows the region of partial melting. Along QR, the vapor phase is greatly enriched in $\text{H}_2\text{O}/\text{CO}_2$ with increasing pressure. At depths greater than about 100 km, the solidus for peridotite- H_2O is at temperature only slightly below that for peridotite- $\text{CO}_2\text{—H}_2\text{O}$.

Recent interpretations of mantle structure

The geochemistry of basalts, in particular the variations in isotopic compositions of Nd, Sr, and Pb, indicates that the mantle contains at least two separate reservoirs that have remained chemically distinct for a billion or more years. Mid-ocean ridge basalts have been derived from a depleted source, whereas oceanic island basalts and continental flood basalts appear to have been derived from undepleted sources. The data require heterogeneities in the mantle which may be achieved by layering, or by some less orderly structure.

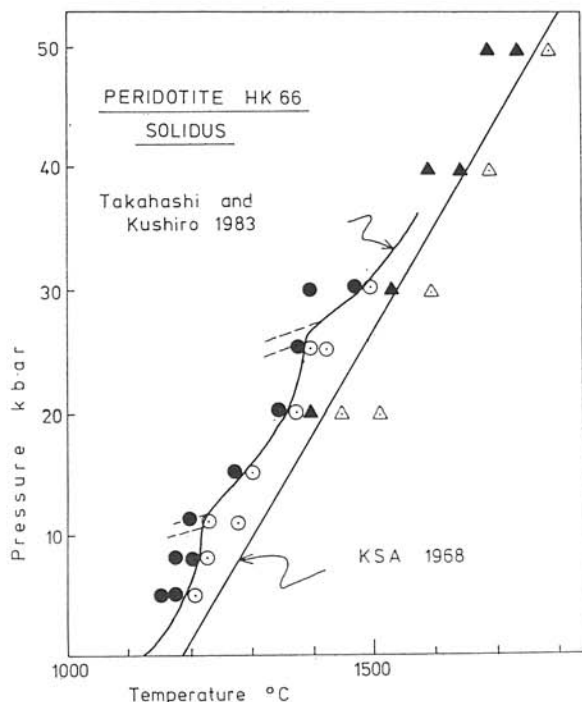


Fig. 2. Solidus curve for dry peridotite KH66, defined by runs shown as circles, after Takahashi — Kushiro (1983). Data from previous study (Kushiro et al., 1968, triangles and straight line) correspond to KSA in Fig. 1.

Jacobsen — Wasserburg (1981) presented a cartoon of the mechanism for crustal growth according to their Model I, with the continental crust and a thick layer of depleted mantle overlying undepleted mantle, from which the two overlying reservoirs had been derived by differentiation. This model can be represented in Fig. 4A by the layered mantle considered by Richter — McKenzie (1981). They assumed that the separate reservoirs, situated below the lithosphere (Layer 1), consisted of superimposed Layers 2 and 3, convecting without significant mass transfer across the boundary between them.

Fig. 4B illustrated Ringwood's (1982) modification of the pyrolite-basalt model (see Ringwood, 1975 for summary). The lithosphere is composed of harzburgite and lherzolite, depleted following extraction of basalt for oceanic crust at mid-oceanic ridges. The layered mantle structure is caused by phase

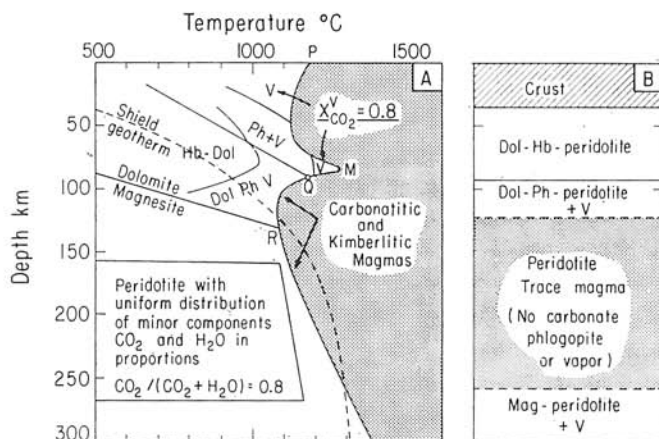


Fig. 3. (A) The phase boundaries for peridotite-H₂O-CO₂ on a depth-temperature diagram, assuming vapor with high CO₂/H₂O. (B) The petrology of a mantle cross-section for peridotite assuming uniform distribution of minor components CO₂ and H₂O, and a shield geotherm (Wyllie, 1980).

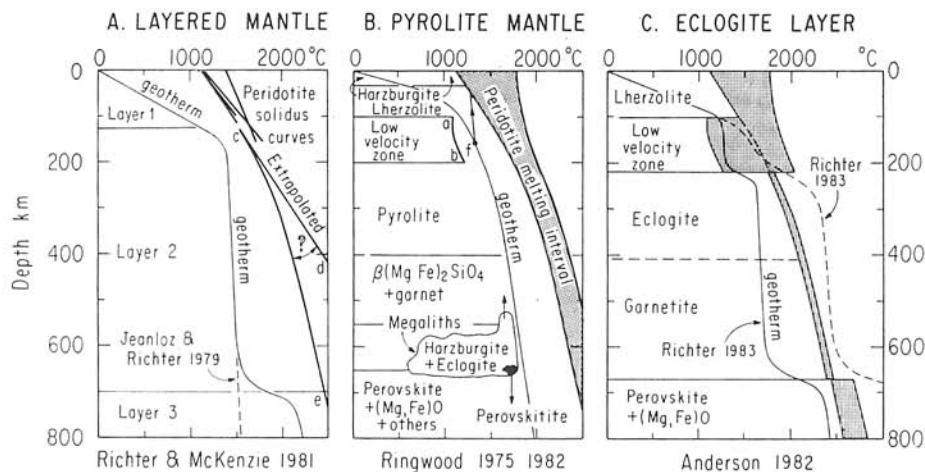


Fig. 4. Proposed mantle structures and geotherms compared with experimentally determined and extrapolated melting conditions for peridotite and eclogite (after Wyllie, 1984). (A) Peridotite solidus curves from Fig. 1, extrapolated along either *ce*, or *cd*. (B) Estimated peridotite melting interval, after Wyllie (1981). The curve *ab* is the solidus QR with buffered vapor, from Fig. 3A. (C) Eclogite melting interval extrapolated from 100 km, based on solidus from Fig. 1.

transitions in pyrolite composition (undepleted). The seismic low-velocity zone is attributed to incipient melting at temperatures above the solidus, *a-b*, for peridotite-H₂O-CO₂ (Fig. 3). Ringwood (1982) identified a second mantle reservoir composed of megaliths of depleted, subducted, cool lithosphere enclosing eclogite from oceanic crust, density-balanced at about 650 km depth.

Fig. 4C shows Anderson's (1982) model based on the geochemistry of basalts, kimberlites and peridotites, and consistent with seismic data. The upper mantle of peridotite composition contains a density-balanced layer of eclogite perched between 220 km and 670 km. The low-velocity zone is attributed to incipient melting, as in Fig. 4B. The eclogite layer is the depleted reservoir, and the upper peridotite layer is a reservoir enriched by kimberlite fluids.

Conditions for melting

Conditions for melting in the different reservoir materials are defined by the solidus for the material, and the local geotherm. It is commonly assumed that material is transferred from one level or layer to another, by upward movement of plumes or blobs in solid, partly molten, or liquid state.

Fig. 4A shows the solidus curves for different peridotites, transferred from Fig. 1. The cusps of Fig. 2 are barely distinguishable on the scale of Fig. 4. A linear extrapolation along *c-d* provides very high solidus temperatures at depth. The calculated Simon-type fit along *c-e* (Griggs, 1972) is probably a better representation of the peridotite solidus. This is paired with an estimated liquidus curve in Fig. 4B (Wyllie, 1981). The solidus *a-b* is for peridotite in the presence of H₂O+CO₂. The schematic, extrapolated melting interval for peridotite in Fig. 4B is replaced between 220 km and 670 km in Fig. 4C by an estimate of the melting interval for eclogite. The extrapolation preserves the narrow melting interval known to occur through 30 kbar, and the proximity of the solidus for quartz-free eclogite with that for peridotite (Fig. 1).

Ringwood (1975) used the calculated geotherm shown in Fig. 4B with diapiric uprise of mantle masses tracing adiabatic paths such as that rising from point *f*. Jeanloz—Richter (1979) presented an improved geotherm for a conducting lithosphere overlying a convecting mantle (Fig. 4A). Richter—McKenzie (1981) added the third Layer 3, convecting independently of Layer 2, and determined from laboratory and numerical experiments that there is a marked increase in temperature between the two convecting layers. Anderson's (1982) mantle in Fig. 4C has an additional convecting layer, compared with the model of Richter—McKenzie (1981) in Fig. 4A. This introduces another step into the geotherm. The solid line in Fig. 4C is a conservative estimate for the geotherm with this mantle model, but the dashed line is a more realistic estimate of what the additional convecting layer would do to the geotherm (Richter, 1983, personal communication).

For a mantle with separate convecting layers, as represented in Figs. 4A and 4C, the solidus curves provide constraints on the dimensions of the layers, because of the large step-like increases in temperature associated with each boundary. With the results used by Richter—McKenzie (1981) in Fig. 4A, the temperature at 700 km is not high enough to cause partial melting of

Layer 3. However, they pointed out that if the bottom of Layer 2 were shallower than about 500 km, then widespread melting is suggested by their calculation. Such an arrangement would be untenable. Therefore, it is not clear how the two-layer convecting mantle could have developed into what is presently a stable condition, if Layer 2 is the depleted mantle reservoir which has grown through time at the expense of undepleted source Layer 3, as in Model I of Jacobsen—Wasserburg (1981). The problem is exacerbated if mantle temperatures were even higher in early history.

Jacobsen—Wasserburg (1981) proposed that blobs of undepleted mantle rise from Layer 3 and intersect the solidus at shallower levels. No physical mechanism was proposed for escape of material from Layer 3.

Ringwood (1982) invoked delayed thermal buoyancy of sunken lithosphere as a mechanism for causing uprise of deep material. He suggested that the depleted lherzolite of the lower lithosphere is stripped off during subduction, and resorbed into the upper mantle providing a depleted source material. The deep, cool megaliths derived from the upper part of subducted oceanic lithosphere (Fig. 4B) reach thermal equilibrium with surrounding mantle at 650 km on a time scale of 1—2 billion years, which is followed by buoyant uprise of the harzburgite, rendered newly fertile by liquids from partial melting of entrained eclogite. The enriched harzburgite is incorporated into the lithosphere, yielding the alkaline basaltic suite by small degrees of partial melting. Parts of this model of multistage irreversible differentiation are inconsistent with the constraints. With the geotherms in Figs. 4A and 4B, the megalith at 650 km remains well below the extrapolated solidus temperature for eclogite (Fig. 4C), even in the presence of H_2O . Adiabatic uprise of buoyant megalith from the geotherm at 650 km in Fig. 4B would probably carry both harzburgite and entrained eclogite across their solidus curves in the depth interval 200—150 km. Adiabatic uprise of newly buoyant harzburgite from the geotherm in Fig. 4A at 650—700 km depth would carry it to a position below the lithosphere lid, Layer 1, without reaching the solidus curve of either harzburgite or entrained eclogite.

Anderson (1982) suggested that extensive partial melting of primitive mantle during accretion was followed by formation of a cumulate eclogite layer, as illustrated in Fig. 4C, with an overlying peridotite layer enriched by residual kimberlite fluids from below. Convection within the eclogite layer deforms the upper boundary, raising it above the solidus temperature. Partial melting produces a buoyant diapir with adiabatic ascent leading to extensive melting and eruption of mid-ocean ridge basalts. The eclogite layer is replenished by subduction of the oceanic crust derived directly from it some million years earlier. Using the more conservative of the two geotherms calculated by Richter, the temperature near the upper boundary of the eclogite layer is close to the solidus for both eclogite and peridotite. If the eclogite layer melts, one might expect the adjacent peridotite to melt as well. Clearly, we need better estimated of solidus curves for both rocks at 200 km.

The temperature of melting in the mantle is lowered by the presence of volatile components (Wyllie, 1979, for review) and potassium (Takahashi—Kushiro, 1983). If traces of these components are available, incipient melting occurs within mantle rocks in a depth interval of about 120 km to

260 km, as shown in Fig. 3B. Actual depths depend strongly on the nature of the geotherm (Fig. 4).

Conditions for metasomatism

Conditions for metasomatism are limited to depth-temperature regions between the phase boundaries for decomposition of hydrous minerals or carbonates, and the solidus curves for the beginning of melting of the mantle materials in the presence of H_2O , CO_2 , or other volatile components. These conditions depend on the composition of the mantle, the availability of volatile components, and the local geotherm. The geotherm depends upon the structure of the upper mantle, and the type of convection occurring. The models in Fig. 4 show a general increase in temperature as a function of depth. However, in other tectonic environments such as subduction, there are depth intervals within the temperature decreases with increasing depth.

Beneath stable plates, given volatile components, (Figs. 3B, 4B and 4C) show that a layer of incipient melting occurs at depths between about 120 km and 260 km. Therefore, metasomatism by dense solutions or vapors is limited to upper mantle shallower than 120 km, or deeper than 260 km.

Fig. 5 illustrates the conditions for metasomatism associated with the deep fractionation of kimberlites, or other low- SiO_2 , high- K_2O melts. Fig. 5A reproduces the solidus and geotherm from Fig. 3A. The solidus for volatile-free peridotite shows that the mantle cross-section in Fig. 5B remains solid unless volatile components become available. If volatile components rise along the geotherm, they induce partial melting in the peridotite as they cross the solidus at *a*. Partly melted diapirs may follow an adiabatic path (*a-f*), or a path such as *a-c* intermediate between the adiabat and the geotherm.

Fig. 5B includes a diagrammatic representation of kimberlite magma ponding into chambers near the base of the lithosphere, as the diapirs reach the solidus MQR (Fig. 5A). Solidification of the magmas would yield concentrated vapors. The vapors would cool along paths such as *f-h* or *e-h*, reacting with overlying peridotite to produce combinations of phlogopite, amphibole, dolomite, and other minerals. The high-temperature CO_2 -rich vapors become enriched in H_2O if the cooling path crosses the dolomite phase boundary (Wyllie, 1980).

According to this scheme, metasomatism is a consequence of kimberlite magmatism, rather than its precursory cause. If the term metasomatism is extended to include composition changes caused by the introduction of silicate magmas, then the layers in Figs. 3B (incipient melting) and 5B (traversed by partly melted diapirs) may be described as metasomatized mantle. The problem then is the source of the volatile components and other trace elements in the melt which give it the capability of enriching the mantle in the required geochemical fashion. Wyllie (1980) concluded that for kimberlite magmatism the most likely source was from deeper levels within the Earth, as illustrated in Fig. 5.

Conditions for melting and metasomatism in subduction zones

If the geotherm is lowered to temperatures below the solidus curve for peridotite- H_2O (Figs. 3A and 5A), then volatile-rich fluids, if present, can flow

without the formation of silicate melt. Release of vapors from deep dissociation fronts in subducted oceanic crusts leads to migration of fluids through large volumes of upper mantle. The geometry of the arrangement of metasomatic volumes, and volumes where the fluids dissolve in magmas, is strongly dependent on the geometry of subduction, the thermal history of the plate being subducted, and the thermal structure of the subduction zone, items which have not been well defined.

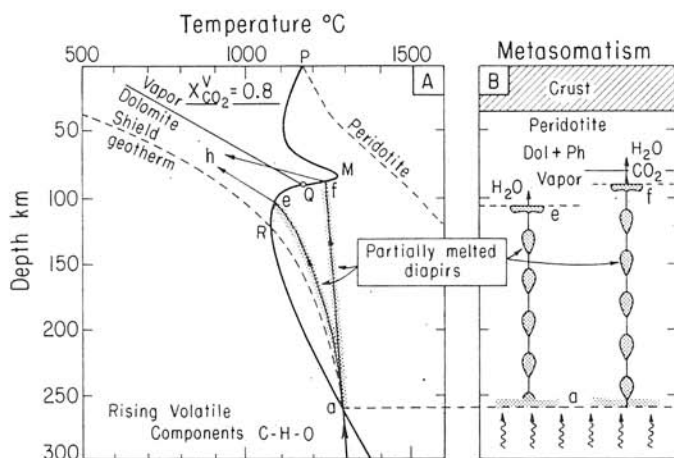


Fig. 5. Compare Fig. 3. In the absence of volatile components, the solidus curve PMQR extrapolated to 300 km has no significance. Volatile components from greater depth rise and cause partial melting at *a*. Partly melted diapirs rise, feeding magma chambers in the lower lithosphere at *e* and *f*, and solidification of the magmas releases volatile components for metasomatism along paths such as *e-h* and *f-h* (Wyllie, 1980).

The subducted oceanic crust may be relatively warm, or significantly cooled by endothermic dehydration reactions (Anderson et al., 1978; 1980). The mantle wedge may be relatively cool, chilled by the subducting slab, or it may be heated by induced convection (Toksöz—Hsui, 1978). Phase boundaries for dehydration and melting at a convergent plate boundary are shown in Fig. 6 for four thermal structures. The experimental basis for these diagrams was reviewed by Wyllie (1979; 1982), and Wyllie—Sekine (1982). The dehydration front for serpentine (D-G) approximates the greenschist facies boundary, and the dehydration front for amphibolite separates amphibole from eclogite in the subducted crust. Partial melting occurs if H₂O passes into the regions on the high-temperature sides of the dashed solidus boundaries for rock-H₂O.

Aqueous solutions rising from the dehydration front DG may precipitate metasomatic amphibole in the large shaded areas of mantle and continental crust, and generate magma in the continental crust (C). For a cool mantle-warm crust (Fig. 6A), most solutions enter the mantle along DE, but some generate magma from subducted crust at M. For a cool mantle-cool crust (Fig. 6B), the

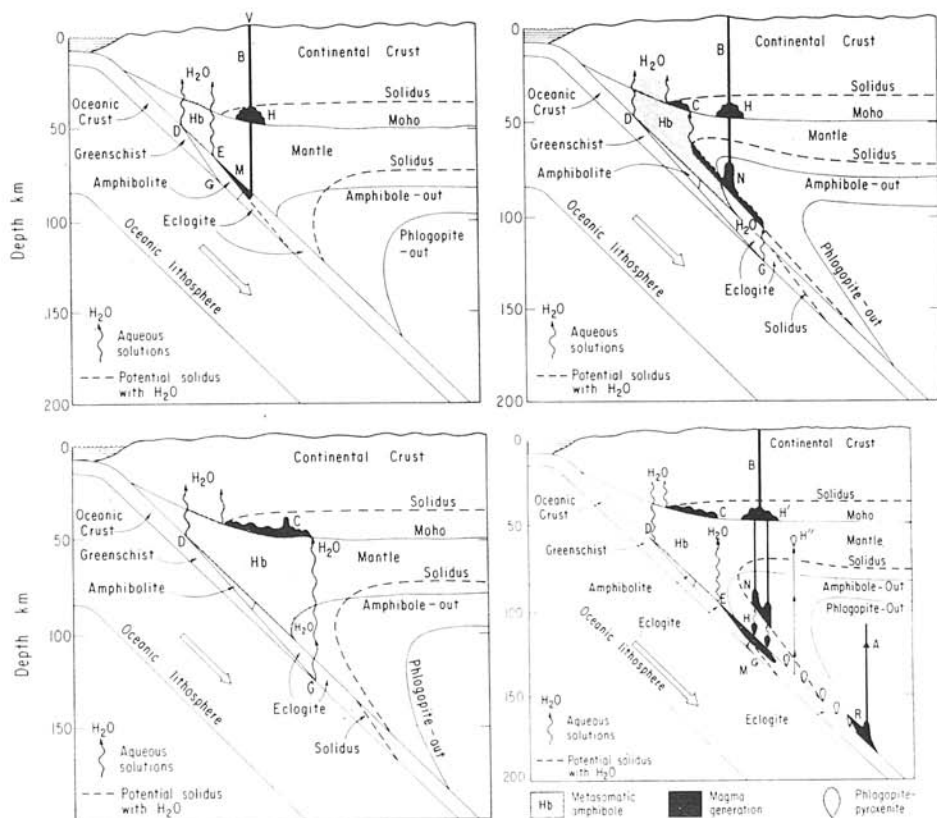


Fig. 6. Locations of dehydration, metasomatism, and melting in subduction zones from results of experimental petrology, according to four thermal structures obtained by combinations of warm or cool (by endothermic dehydration) subducted ocean crust, with cool or warm (by induced convection) mantle. (A) Cool mantle—warm crust. (B) Cool mantle—cool crust. (C) Warm mantle—cool crust. (D) Warm mantle—warm crust (Wyllie—Sekine, 1982).

metasomatic solutions do not enter the melting regions of either subducted crust or mantle peridotite. For a warm mantle—cool crust (Fig. 6C), induced convection brings the region for potential melting of the mantle above the dehydration fronts, and metasomatic fluids cause partial melting of peridotite. Fig. 6D represents an intermediate thermal structure with relatively warm mantle and relatively warm subducted crust. Magmas are generated simultaneously in all three source materials at M, N, C, and H'. Sekine—Wyllie (1982a; 1982b; 1983) and Wyllie—Sekine (1982) investigated the proposal of Nicholls—Ringwood (1973) that hydrous siliceous magma from M would react with overlying mantle, producing hybrid olivine pyroxenite. They concluded that a series of discrete or overlapping bodies of hybrid phlogopite-pyroxenite would be transported downwards with the slab to R, where

vapor-absent melting defines another possible site of magma generation. Solutions released during hybridization at H cause metasomatism, followed by partial melting in the shallower mantle at N.

Composition of magmas and metasomatic solutions

This topic requires the space of another paper. Recent studies of the compositions of melts from peridotite (Jacques — Green, 1980; Takahashi — Kushiro, 1983) confirm that near-solidus melts range from basalts at moderate pressures, to alkali picrite and komatiite above 25 Kbar. Wyllie (1984) reviewed available data and concluded that the generation of magmas more siliceous than basalts is limited to shallow suboceanic mantle with high H_2O contents; low- SiO_2 , high-alkali magmas are generated at depths greater than 100 km by concentration of CO_2 .

Wyllie — Sekine (1982) reviewed the limited experimental data on the composition of aqueous solutions or vapors. The solubility of components is sensitive to the mineralogy of the host rocks as well as to depth and temperature. Total dissolved solute for phlogopite-bearing assemblages may reach as high as 50 % (Ryabchikov — Boettcher, 1980). Mysen (1979, 1983) measured rare earth element partition coefficients between ($H_2O + CO_2$) vapors and silicates, and used the results to evaluate the processes of metasomatism and partial melting involved in the formation of alkali basalt from garnet lherzolite. Metasomatic vapors rising from greater to lower depths commonly follow paths of decreasing temperature (Figs. 4 and 5), but in subduction zones they may rise from lower temperature subducted oceanic slab into higher temperature mantle wedge (Fig. 6). Metasomatic vapors either scavenge peridotite for additional components, or change its composition by exchange or precipitation of components. We need much more experimental data.

Physical aspects

The physical properties of rocks, rock mushes, melts, and vapors are just as phase diagrams and geotherms in evaluating the conditions for melting and metasomatism, but less well known. There are three mechanism for transport of magmas and metasomatic vapors through the mantle: 1) by percolation or porous flow of fluid through a crystalline matrix; 2) by the motion of rock mushes or segregated melt in diapiric form; and 3) by the propagation of elastic cracks filled with fluid. The compositions of segregated melts depend strongly on the physical conditions for the escape of melt from rock mush. Walker (1983) recently reviewed these topics. Spera (1981) presented the fluid dynamics of ($H_2O + CO_2$)-rich fluids rising through cracks as a first step toward a more complete theory of the physical chemistry of mantle metasomatic processes.

It is the physics of rock masses and melts and vapors which eventually controls the chemical differentiation of the Earth. The major chemical differentiation is accomplished by melting and movement of magmas, but the distribution of minor and trace elements in the mantle and magmas is strongly influenced by the migration of fluids (melts or vapors) rich in volatile components.

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