Constraints Imposed by Experimental Petrology on Possible and Impossible Magma Sources and Products [and Discussion]

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Phil. Trans. R. Soc. Lond. A 1984 **310**, 439-456
doi: 10.1098/rsta.1984.0003

Phil. Trans. R. Soc. Lond. A **310**, 439–456 (1984)
Printed in Great Britain

Constraints imposed by experimental petrology on possible and impossible magma sources and products

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Experimental petrology can be used in forward and inverse approaches. The forward approach defines the compositions of liquids generated by partial melting of possible source rocks at various depths. The inverse approach determines conditions for multiple-mineral saturation at the liquidus of primitive magmas, correlates them with residual minerals of possible source rocks, and thus provides estimates of depths and temperatures required for their derivation. Review of a selection of forward and inverse results is followed by evaluation of petrological and geophysical processes in layered mantle and in subduction zones. Physical constraints imposed by solidus curves and geotherms present problems for models that derive basalts from deep mantle reservoirs, separated from overlying convecting layers. Magmas from mantle are limited to compositions less siliceous than basaltic andesite, with rare exceptions. Granite liquids cannot be generated from normal peridotite, nor from oceanic crust at mantle pressures in subduction zones. In continental crust, hydrous granite liquid is generated at depths of less than 30 km. Basaltic andesite and picritic basalt are parental magmas for the calc-alkaline series. Andesite is not primary from sub-continental depths, and can be generated as liquid in continental crust only if temperatures exceed about 1100 °C. Calc-alkaline magmas may contain components from mantle peridotite, subducted oceanic crust, and continental crust.

INTRODUCTION

The mantle, subducted oceanic crust, and continental crust constitute three chemically distinct magmatic hearths, which experience partial melting under different physical conditions.

The oceanic crust and continental crust were generated from the mantle by magmatic processes. The oceanic crust exists for no more than 200 million years before being returned to the mantle via subduction, except for that portion which is incorporated into the continental crust by tectonic or magmatic processes. The continental crust is made up largely of magmas derived from the mantle, from subducted oceanic crust, and from its own deeper components.

Understanding the history of mantle differentiation and continental growth requires identification of the source or sources of the magmas forming the crust, as well as estimates of the proportions of magmas derived from the various sources. Major efforts to identify the chemical characteristics of source rocks have been made by geochemical analysis of the observed magmatic products, and some possible source rocks. This approach has provided evidence for complex multi-stage processes involving source rocks and magmas from all three magmatic hearths. Experimental petrology provides additional constraints for these processes.

The results of experimental petrology should reveal all possible paths of liquid composition, equilibrium and fractional, together with the compositions of coexisting minerals and gas phases, for all bulk compositions relevant for the formation and development of the mantle–crust system. This includes the distribution of minor elements, trace elements, and isotopes among solids, liquids and gases as a function of pressure and temperature, with determination of the effect of the elusive volatile components such as H_2O, CO_2 and O_2 which may be influential in some environments. In fact, progress is made through a more selective approach.

FORWARD AND INVERSE APPROACHES IN SOURCE-MAGMA RELATIONS

Wyllie *et al.* (1981) discussed the 'forward' and 'inverse' approaches to the problem of source-magma relations, and reviewed in detail the inverse approach for basalts. The inverse approach is the standard geological method: geologists study the accessible rocks to infer processes that occurred at inaccessible depths, in earlier times. Interpretation of geochemical analysis of magmatic products in terms of the chemistry of source rocks is an inverse approach. Experimental petrology is used in both forward and inverse directions, and figure 1 is an outline of how they supplement each other.

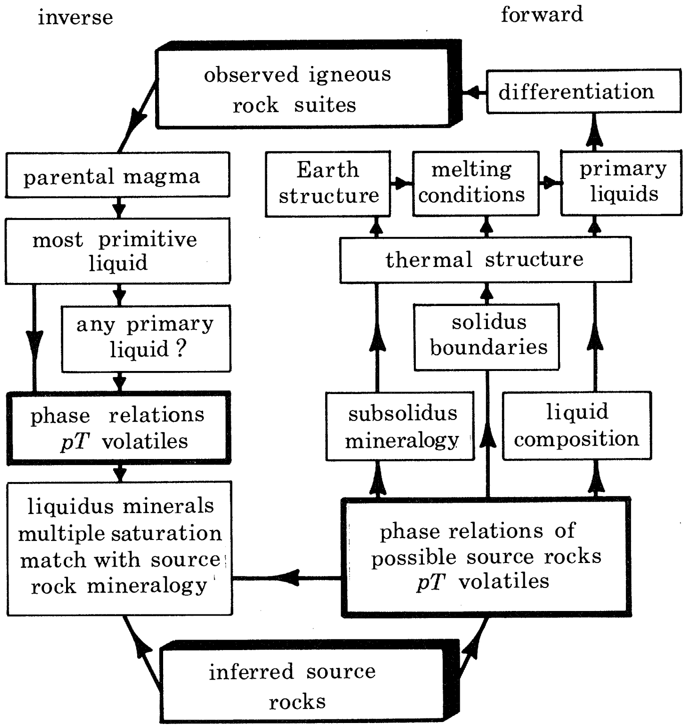

FIGURE 1. The inverse and forward approaches for testing the relations between observed igneous rocks and inferred source rocks as depth, using phase equilibrium results from experimental petrology.

Experimental data provide a set of constraints that can be used in the following ways: (1) to define the mineralogy of possible source rocks as a function of depth; (2) to define the compositions of liquids produced from possible source materials; (3) to define polybaric polythermal paths of crystallization, or differentiation; (4) to determine, in conjunction with geochemical data, which composition in an igneous suite represents the most primitive, or primary liquid; (5) to determine the conditions under which specific liquid compositions could coexist with residual minerals in a source rock. These approaches, in combination with the thermal structure, provide a means to test geophysical and petrological models in different tectonic environments.

THE FORWARD RESULTS: SOURCE ROCKS

The mantle, subducted oceanic crust, and continental crust contain several different rock types that may contribute to magmas. We should know for each rock: (1) the sub-solidus phase relations at least to the depths of magma generation, (2) the solidus curve, which defines the conditions for beginning of melting, (3) the percentage of melting as a function of temperature above the solidus, and (4) the changing composition of the liquid within the melting interval. For some environments, we should know (5) the effect of volatile components such as H_2O and CO_2 on the previous items. In order to translate these chemical parameters into magmatic processes, we need to know in addition (6) the physical characteristics of the partially molten rocks and their variation with pressure, temperature, percentage of melting, and liquid-mineral compositions. This information is required to establish the range of conditions through which a partly molten source may rise as a unit, the conditions through which a liquid may percolate upwards through the crystalline mush, maintaining equilibrium with its host, and the condition where the liquid may escape from its host and rise as an independent body of magma. The following figures summarize much of the information available for items (1)–(5), while demonstrating that we need more data on most possible source materials.

Mantle source rocks

Candidates for source rocks in the upper mantle include peridotite and eclogite. Peridotite may vary in mineralogy from lherzolite to harzburgite, according to the degree of depletion of major and minor elements caused by extraction of magmas. The more refractory peridotites have higher $MgO/(MgO+FeO)$. The extent of depletion or enrichment of the rocks in incompatible trace elements has little effect on the phase relations.

Solidus curves determined experimentally for a variety of mantle-derived lherzolites and synthetic pyrolite are compared in figure 2. The differences among the curves relate in part to different degrees of depletion of the peridotites, but in part to experimental problems such as those reviewed by Wyllie *et al.* (1981, §3.2). There have been few determinations of solidus curves for eclogites. Olivine eclogite (tholeiite) has solidus indistinguishable from the lowest-melting peridotite in figure 2; quartz eclogite (tholeiite) begins to melt at temperatures significantly lower. None of the published solidus curves has been closely defined by good experimental brackets.

The results of two attempts to measure the percentage of liquid produced above the solidus of peridotite are compared in figure 3*a*. Mysen & Kushiro (1977) reported near-isobaric invariant melting with production of much liquid just above the solidus at 20 kbar† and 35 kbar in a garnet peridotite, with distinct change in slope of the melting curve wherever one or more minerals disappeared. Jaques & Green (1980) claimed that their results for pyrolite demonstrated that the step-like melting characteristic of simple synthetic systems was smoothed out by solid solution effects involving Fe–Mg, Al–Si, Na–Ca, etc. Despite the different solidus curves for the two rock compositions, the melting results appear to converge for about 50% melting.

Figure 3*b* shows the subsolidus phase relations and estimated melting interval for garnet lherzolite, with the *p*–*T* ranges for liquids of various compositions within the melting interval. Liquid compositions correspond to those calculated by Jaques & Green (1980), with elaborate procedures to analyse by mass balance the results of experimental runs with two peridotite compositions between 2 kbar and 15 kbar. Jaques & Green (1979) had concluded that 'most,

† 1 bar = 10^5 Pa.

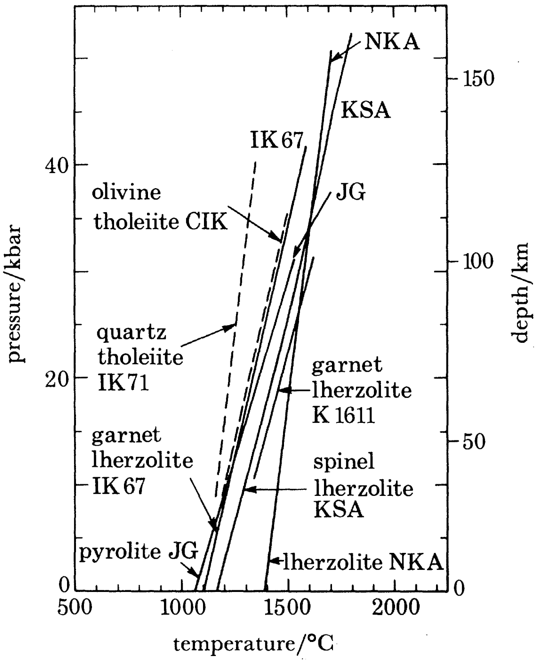

Figure 2. Solidus curves measured for various peridotites, for olivine tholeiite (eclogite), and for quartz tholeiite (eclogite), – – –, basalt-eclogite; ——, peridotite. Abbreviations: IK, Ito & Kennedy (1967) or (1971); JG, Jaques & Green (1980); K, Kushiro (1973); KSA, Kushiro *et al.* (1968); NKA, Nishikawa *et al.* (1970); CIK, Cohen *et al.* (1967). (1 bar = 10^5 Pa.)

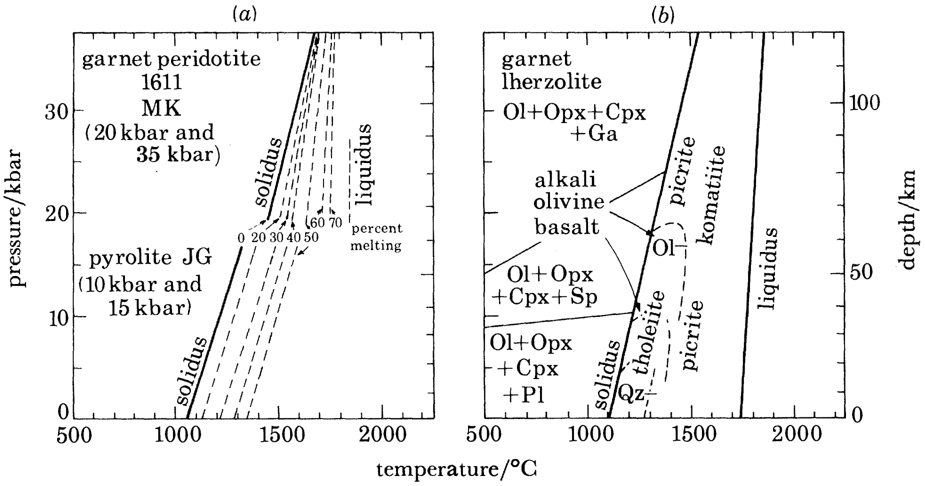

Figure 3. Phase relations for peridotites. (*a*) Limited data for percentage of liquid within melting interval above solidus (JG and K 1611 in figure 2); MK, Mysen & Kushiro (1977). (*b*) Compositions of liquids generated within melting interval after Jaques & Green (1980), with solidus curve IK 67 from figure 2.

perhaps all, previously published partial melt compositions obtained by direct analysis of quenched glasses' are in error, because of iron loss from sample (Nehru & Wyllie 1975) and growth of metastable crystals during quench (Green 1976).

Figure 4*a* reproduces the liquid compositions and solidus from figure 3*b*, and shows in addition the solidus curves for peridotite-H_2O, peridotite-CO_2, and the divariant solidus surface connecting these curves, for melting in the presence of $H_2O + CO_2$. Details of the

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phase relations and the geometry of isoplethal sections are given in Wyllie (1978*a*, 1979, 1980). There have been several experimental studies and many controversial reviews concerned with the compositions of near-solidus liquids in this system. Figure 4*a* is an attempted compromise that illustrates the major trends, with chemical variation shown by the lines representing changes in normative composition of the near-solidus liquids (Kushiro 1972; Nicholls 1974; Mysen & Boettcher 1975; Green 1976; Wyllie & Huang 1975; Eggler 1978; Wyllie 1978*a*; Wendlandt & Mysen 1980; Wyllie *et al.* 1981, pp. 553–556). The amount and composition of liquid developed at the solidus depends on the amount of (CO_2+H_2O), on CO_2/H_2O, and on the presence or absence of buffering minerals amphibole, dolomite, and phlogopite (see Eggler 1978; Wyllie 1978*a*, 1979). With progressive melting at high pressures, the liquids follow paths through wide temperature intervals towards the basaltic and picritic liquids near the solidus of volatile-free peridotite.

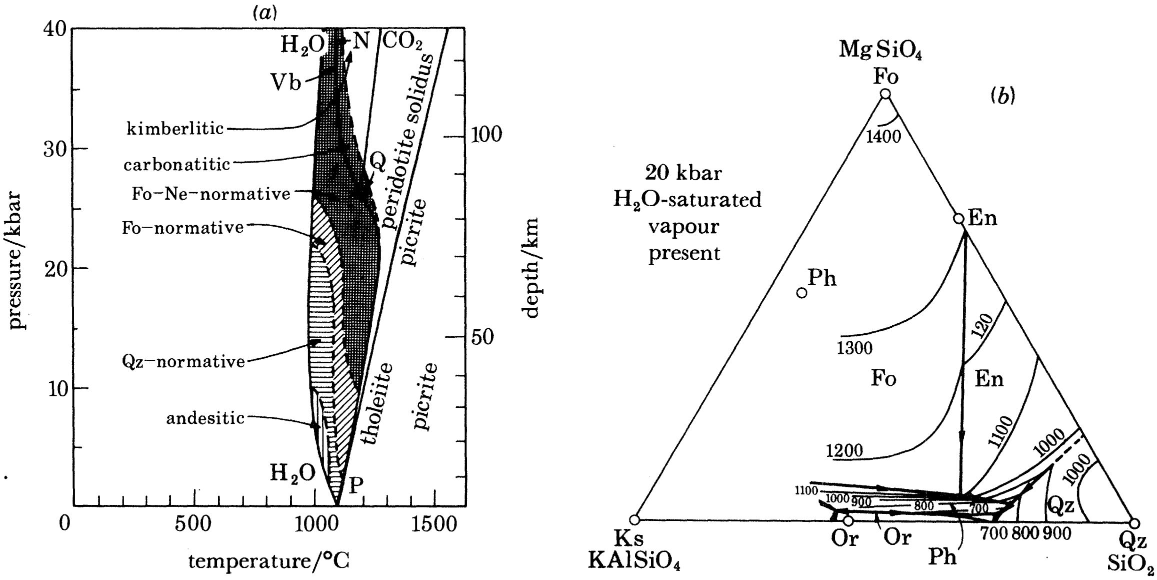

FIGURE 4. (*a*) Projection of the solidus surface for peridotite–CO_2–H_2O, with tentative boundaries showing the changes in composition of the near-solidus, vapour-saturated liquids (Wyllie 1979). Solidus with CO_2, P–Q–CO_2; with H_2O, P–H_2O; and for dolomite–peridotite with buffered vapour (Vb), Q–N. (*b*) H_2O-saturated liquidus for part of the system $KAlSiO_4$–Mg_2SiO_4–SiO_2–H_2O at 20 kbar, based on available experimental data, following Yoder (1976, p. 39). For isothermal sections, see Sekine & Wyllie (1982*a*).

Modreski & Boettcher (1973) reported microprobe analyses of glasses quenched from synthetic assemblages modelling phlogopite-peridotite. For enstatite-bearing assemblages with H_2O vapour between 10 kbar and 30 kbar, glasses averaged 72% SiO_2. High SiO_2 contents for H_2O-saturated liquids from phlogopite + forsterite + enstatite are illustrated by the peritectic point on the liquidus at 1130 °C in figure 4*b*. When phlogopite has melted, the liquid becomes depleted in SiO_2 and enriched in MgO as it follows the field boundary between the liquidus surfaces for forsterite and enstatite.

Liquid compositions in the narrow melting interval of eclogite at 27 kbar (compare solidus in figure 2 with dry liquidus in figure 5*a*) appear to be richer in Fe/Mg and in Ca/(Ca + Mg) than andesites of the calc-alkaline rock series. In the upper part of the melting interval, liquid compositions approach those of basaltic andesites more closely (Green & Ringwood 1968; Stern & Wyllie 1978).

Subducted oceanic crust

The main source rock in subducted oceanic crust is basalt or gabbro, metasomatized to uncertain extent by penetration of sea-water, and metamorphosed during subduction through greenschist, amphibolite, and eclogite facies. Subducted pelagic sediments may also contribute to the magmas (Karig & Kay 1981). Bodies of serpentinite represent a source of aqueous solutions. The subduction of volatile components is the feature that distinguishes this tectonic environment from all others, and the influence of these components after thermal dissociation produces magmas of distinctive composition.

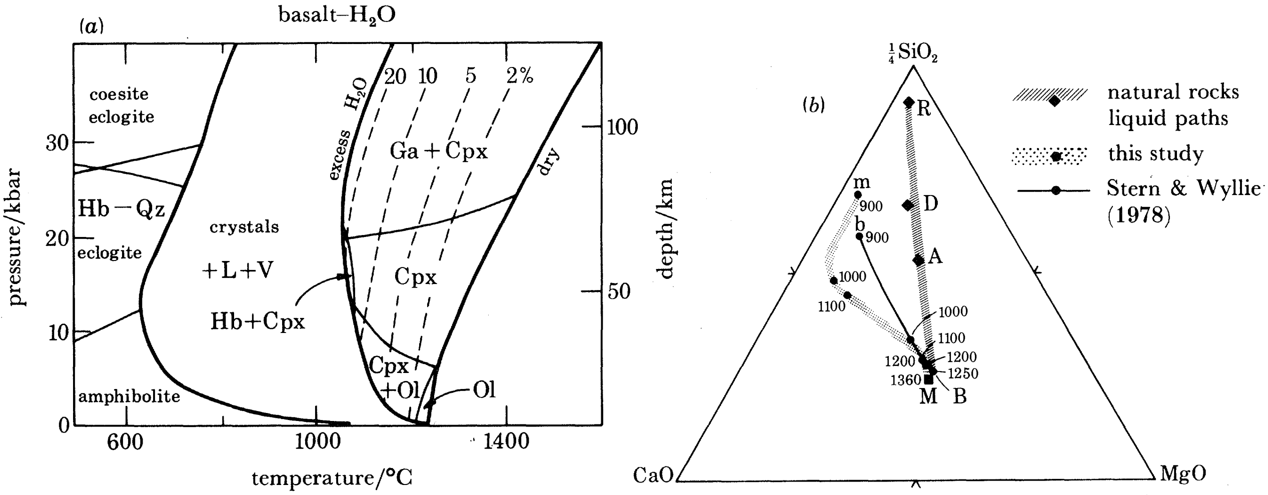

FIGURE 5. (*a*) Phase relations for basalt–H_2O, modified, after Stern *et al.* (1975) by Wyllie (1979). (*b*) The calc-alkaline rock sequence B–A–D–R compared with equilibrium paths for liquids generated experimentally in quartz eclogite at 30 kbar with 7.5% mole% H_2O, using natural basalt (bB, Stern & Wyllie 1978), and synthetic basalt in system $CaO–MgO–Al_2O_3–SiO_2–H_2O$ (mM); from Sekine *et al.* (1981).

There is little experimental information on the compositions of liquids generated in subducted sediments at mantle pressures. Stern & Wyllie (1973) presented data for a pelagic red clay, Huang & Wyllie (1973) evaluated the melting of micaceous quartzo-feldspathic sediments, and Huang *et al.* (1980) presented data applicable to the melting of siliceous limestones.

Figure 5*a* shows phase relations for basalt-H_2O. There is a very wide melting interval for quartz–coesite-eclogite with H_2O, extending from the solidus to the contour on the liquidus surface for the specific H_2O content. Figure 5*b* illustrates the results of attempts to determine the compositions of liquids produced during progressive fusion of quartz-eclogite with H_2O at 30 kbar. Stern & Wyllie (1978) determined liquid compositions ranging through intermediate SiO_2 contents, but with Ca/(Mg + Fe) higher than the average composition trend for calc-alkaline dacite–andesite–basalt. Results of Sekine *et al.* (1981) for a synthetic basalt composition in the system $CaO–MgO–Al_2O_3–SiO_2–H_2O$ reinforced the conclusion. From evaluation of these results and those of Green & Ringwood (1968), Green (1972), Allen *et al.* (1975) and Allen & Boettcher (1978), Wyllie (1978*b*, 1982) concluded that at 30 kbar, neither dry nor wet andesite liquids are produced during partial fusion of subducted oceanic crust. Temperatures in excess of 1400 °C at 100 km would be required to generate relatively dry basaltic andesites through very extensive melting of the crust (Maaløe & Petersen 1981).

Continental crust

The dominant rocks of the deep continental crust consist of the calc-alkaline plutonic rocks and their metamorphic equivalents, together with metamorphosed shales and greywackes. Wyllie (1977) presented a detailed review of experiments related to crustal anatexis.

Figure 6 shows the approximate p–T ranges for the generation of liquids of various compositions from continental gneisses, with a generous allowance of 2 % H_2O in pore fluids. The temperature interval for the persistence of H_2O-saturated granitic liquids is narrow. At lower pressures, H_2O-undersaturated granite (rhyolite) liquid coexists with residual minerals through a wide range of temperatures. Very high temperatures are required to reach liquid compositions corresponding to tonalites (andesites). Increasing pressure produces liquids with lower SiO_2, and rhyolite liquid cannot be produced at the base of the continental crust. For thickened crust, the near-solidus liquid approaches syenite composition (Huang & Wyllie 1981).

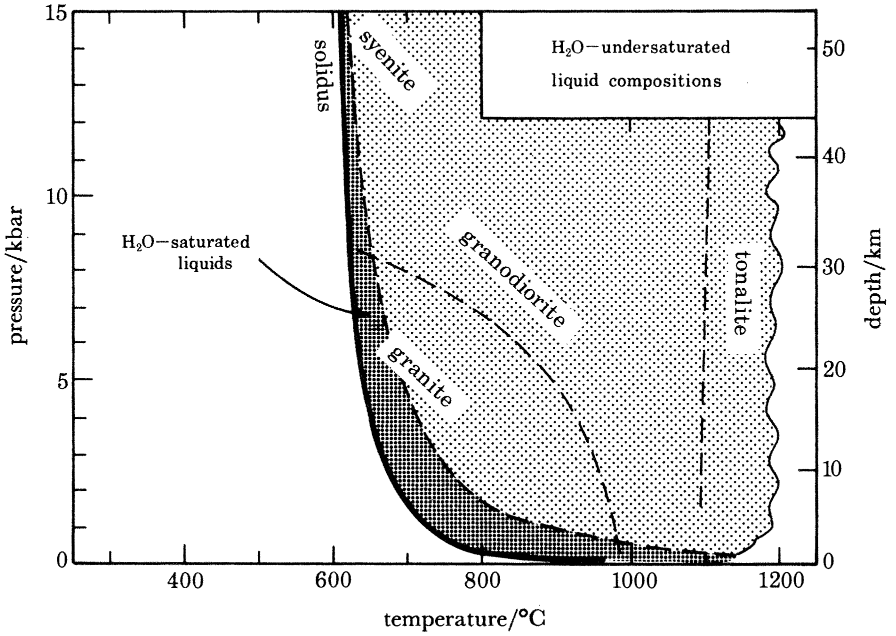

FIGURE 6. Compositions of liquids generated within the melting interval of rocks of continental crust in presence of H_2O (after Wyllie 1977).

THE INVERSE RESULTS: PRIMARY MAGMAS

A primary magma is a liquid whose composition has not changed since it was generated by partial melting in its source region. Figure 1 outlines the procedures in the inverse approach of deducing source rock chemistry and mineralogy from the phase relations of primary magmas. Wyllie *et al.* (1981) gave detailed accounts of the procedures, of a search for possible primary magmas, and a case study for oceanic tholeiites. They concluded that primary magmas were rare, most of them having experienced low-pressure crystal-liquid fractionation.

One critical test is to match multiple-mineral saturation on the liquidus of a primitive magma with the residual minerals of a multi-mineral source rock at the same p–T conditions. This inevitably involves some prior notion of possible source rocks; the inverse approach and the forward approach complement each other. Although an internally consistent solution may be constructed for source-magma relations, this is not necessarily unique. To demonstrate that a melt could be generated by partial melting of a particular source rock under specific conditions is no proof that this was, in fact, how the magma was derived. The approach may be more successful in demonstrating that a proposed petrogenetic scheme is wrong because it violates the constraints of experimental petrology.

The abundant magmas of most concern for the origin and evolution of the continental crust are komatiites, basalts, basaltic andesites and andesites, and dacites and rhyolites. Space limitations preclude a review of phase relations which might provide an inverse test for all of these magmas. We will consider next the effect of temperature and process on magmatic history, and then return to a compilation of conclusions about source-magma relations from combined forward and inverse approaches.

MANTLE TEMPERATURES AND PROCESSES

Figure 7 illustrates three recent interpretations of the structure and mineralogy of the upper mantle, taking into consideration the geochemical evidence for two reservoirs that have remained chemically distinct for a billion or more years. Mid-ocean ridge basalts are derived from a depleted reservoir, and oceanic island and continental basalts are derived from an enriched reservoir.

Figure 7*a* illustrates Ringwood's (1982) modification of the pyrolite-basalt model (see Ringwood 1975 for summary). The layered mantle structure is caused by phase transitions in pyrolite composition. The lithosphere is composed of harzburgite and lherzolite, depleted following extraction of basalt for oceanic crust at mid-oceanic ridges. The seismic low-velocity zone is attributed to incipient melting at temperature above the solidus, a–b, for peridotite-H_2O–CO_2 (figure 4*a*). 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.

Richter & McKenzie (1981) assumed that the separate reservoirs required by the geochemistry of basalts consisted of superimposed layers 2 and 3 in figure 7*b*, convecting without significant mass transfer across the boundary through a billion years or so. They evaluated the thermal consequences of this constraint. The mantle structure corresponds to the geochemical model of a layer (2), depleted in basaltic components, overlying deeper, fertile material (layer 3).

Figure 7*c* 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 figure 7*a*). The eclogite layer is the depleted reservoir, and the upper peridotite layer is a reservoir enriched by kimberlite fluids.

Figure 7*a* shows the variety of solidus curves for different peridotites (from figure 2), and the uncertainty of extrapolation through the upper mantle along c–d or c–e (Wyllie 1973). A schematic, extrapolated melting interval for peridotite is given in figure 7*b* (from figure 3*b*, and Wyllie 1981). In figure 7*c*, this melting interval is replaced between 220 km and 670 km by an estimate of the melting interval for eclogite. At depths to 100 km, the solidus for quartz-free eclogite is just below, but barely distinguishable (within experimental limits) from that for undepleted lherzotile (figure 2), and the liquidus for dry basalt (figure 5*a*) is not much higher. These relations have been preserved in the extrapolation.

The geotherm in figure 7*a* was calculated (Ringwood 1975), and diapiric uprise of mantle traces adiabatic paths such as that rising from point f. Jeanloz & Richter (1979) presented an improved geotherm for a conducting lithosphere overlying a convecting mantle (figure 7*b*). Richter & McKenzie (1981) added a third layer 3 in figure 7*b*, convecting independently of layer 2. From laboratory and numerical experiments, they determined that there is a marked increase in temperature between the two convecting layers. Anderson's (1982) mantle in figure 7*c* has an additional convecting layer, compared with that of Richter & McKenzie (1981) in figure 7*b*. According to Richter (1983, personal communication), this introduces another step in the geotherm. The solid line in figure 7*c* is Richter's most conservative estimate for the geotherm, but he considers the dashed line to be a more realistic estimate of what the additional convecting layer would do to the geotherm. Temperatures would become so high that the system could not be maintained in the proposed arrangement.

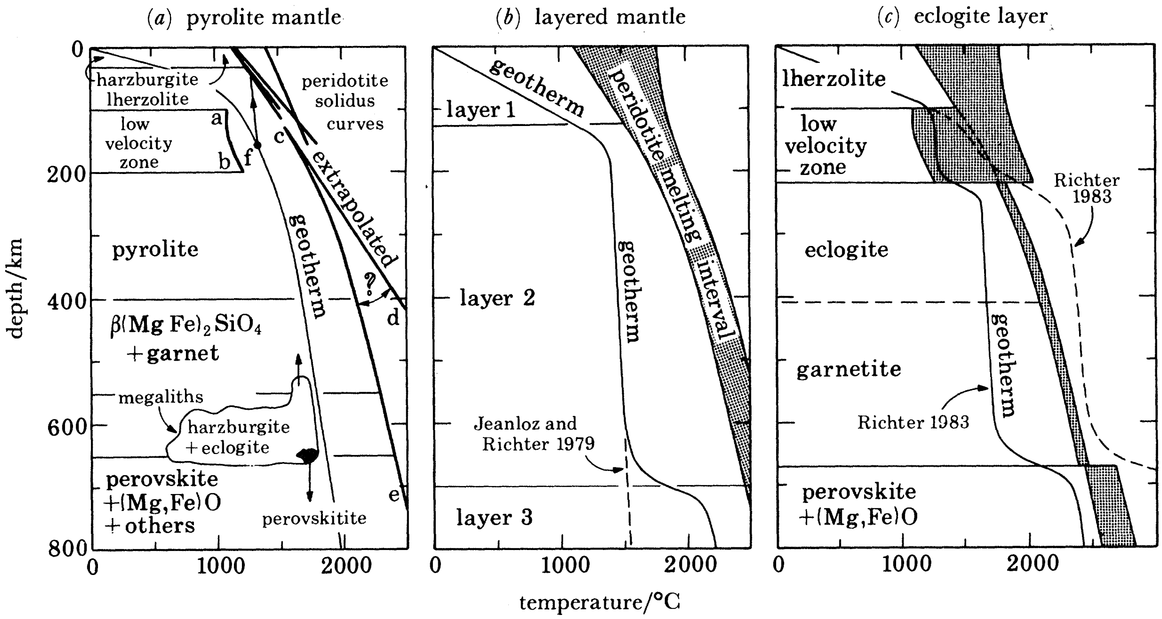

FIGURE 7. Proposed mantle structures and geotherms compared with experimentally determined and extrapolated melting conditions for peridotite and eclogite. (*a*) Peridotite solidus curves from figure 2, extrapolated along either ce, or cd. The curve ab is the solidus QN with buffered vapour, Vb (H_2O–CO_2), from figure 4*a*. (*b*) Estimated peridotite melting interval, after Wyllie (1981). (*c*) Eclogite melting interval extrapolated from 100 km, based on solidus from figure 2, and liquidus from figure 5*a*.

The relation between thermal structure and melting boundaries is obviously critical for interpretation of igneous processes. In order to generate magmas, mantle material in figure 7 must be raised from the existing geotherm to cross a solidus curve. This may be accomplished by adiabatic uprise, as from point f in figure 7*a*, or by elevation of the geotherm through some tectonic or deep-seated mechanism. If magmas are generated in deeper mantle reservoirs, such as layer 3 in figure 7*b*, their temperatures are very high. If partial melting is followed by diapiric uprise of crystal mush, the temperature may be constrained to remain within the melting interval by adsorption of latent heat, and thus lowered to more reasonable magmatic temperatures nearer the surface. However, if sufficient liquid is generated for escape and independent uprise, the liquid will enter the lithosphere in an extremely superheated condition. Hypotheses appealing to the generation of enriched magmas from deep mantle reservoirs must consider seriously the high melting temperatures, and devise processes either for bringing superheated melts from great depths and cooling them within the lithosphere, or for transferring solid material from layer 3 through layer 2 without mixing, followed by partial melting at shallower depths. Either process is difficult, given the evidence that convecting layers 2 and 3 have remained chemically separate through 1–2 billion years.

According to Ringwood's (1982) interpretation, the depleted pyrolite of the lower lithosphere generated by differentiation at mid-oceanic ridges is stripped off during subduction, and resorbed into the upper mantle, providing depleted source material for future mid-oceanic ridge basalts. The deep, cool megaliths derived from the upper part of subducted oceanic lithosphere (figure 7*a*) reach thermal equilibrium with the surrounding mantle at 650 km on a time scale of 1–2 billion years, followed by buoyant uprise of the harzburgite, which has been rendered newly fertile by liquids from partial melting of the 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 multi-stage irreversible differentiation of the mantle are inconsistent with the constraints provided by solidus curves. With the geotherms in figures 7*a* or 7*b*, the megalith at 650 km remains well below the extrapolated solidus temperature for eclogite (dry or wet, figure 5*a*). The harzburgite cannot be refertilized at this depth. Adiabatic uprise of buoyant megalith from the geotherm at 650 km in figure 7*a* 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 figure 7*b* at 650–700 km depth would carry the harzburgite to a position below the lithosphere lid, layer 1, without reaching the solidus curve of either harzburgite or entrained eclogite.

The presence of separate convecting layers in the mantle causes large step-like increases in temperature on the geotherm, and solidus curves thus provide constraints on the physical dimensions of such layers. With the results used by Richter & McKenzie in figure 7*b*, 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 the model calculation. Such an arrangement would be untenable, especially when one considers the consequences for earlier times, with even higher mantle temperatures.

According to Anderson (1982), extensive partial melting of primitive mantle during accretion was followed by the formation of a cumulate eclogite layer, as illustrated in figure 7*c*, with an overlying peridotite layer enriched by residual kimberlite fluids from the eclogite layer. Convection within the eclogite layer deforms the lower boundary, and raises 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. With the more conservative geotherm calculated by Richter, the temperature near the lower boundary of the eclogite layer is very close to the solidus for both eclogite and peridotite. If extensive melting of the lower part of the eclogite layer occurs, one might expect the adjacent peridotite layer to melt, as well. Clearly, we need better estimates of solidus curves for both rocks at 700 km to place reliable constraints on the model. However, the geometry of figure 7*c* suggests that adiabatic uprise of partly molten eclogite from a depth near 650 km would transport it through the liquidus, producing superheated eclogite liquid with an embarrassingly high temperature for melt entering the lithosphere.

The three models for mantle differentiation represented in figure 7 yield different interpretations for the evolution of the continental crust. Testing the models against proposed thermal structures and the constraints provided by solidus curves (greatly extrapolated from actual experimental measurements) indicates that all three models have problems to be resolved.

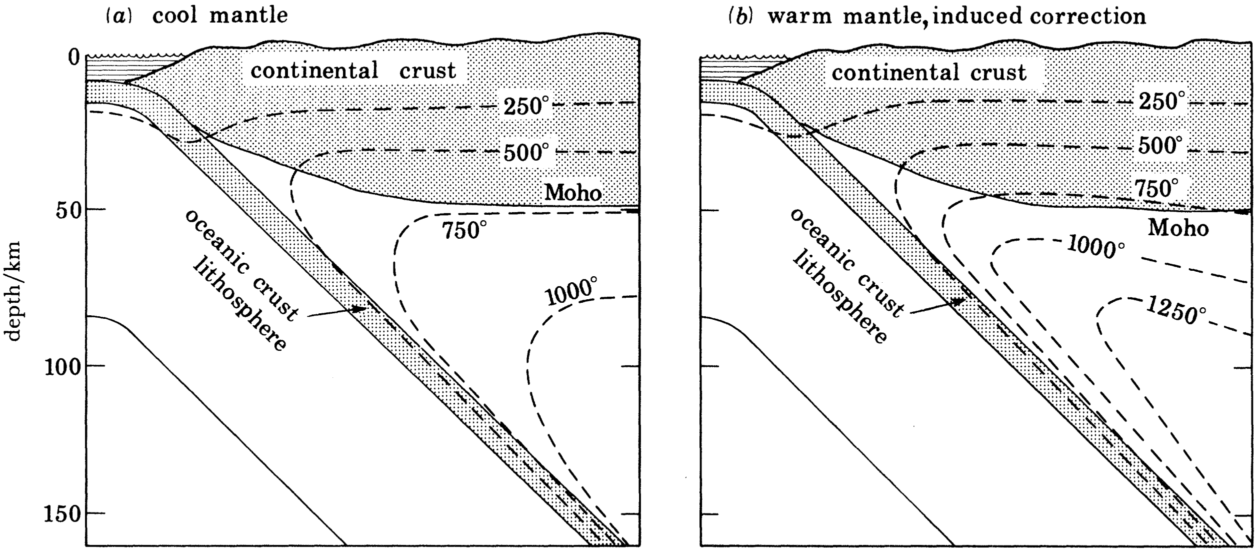

Figure 8. Schematic thermal structures in subduction zones. Mantle isotherms may be relatively low, as in (*a*), or relatively high through induced convection in the mantle wedge (Toksöz & Hsui 1978) as in (*b*).

Subduction zone temperatures

All three source materials, the subducted oceanic crust, mantle peridotite, and continental crust, occur in subduction zones. In order to relate the observed magmatic products to the source regions, the thermal structure and its variation with time are key factors.

The thermal structures of subduction zones have not been well defined. 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). Isotherms for these two conditions are illustrated in figure 8. Each can be modified for a cool or warm subducted crust, which provides a set of four thermal structures. For ocean–ocean convergent boundaries, Marsh (1979) and Maaløe & Petersen (1981) have appealed to extreme induced convection cells in the mantle wedge, raising the temperature of the subducted oceanic crust at 100 km depth to 1250 °C or 1450 °C. This provides a fifth thermal structure, the 'super-hot' mantle, and Maaløe & Petersen (1981) suggest that similar conditions may occur beneath an ocean–continent plate boundary, because they refer to the contamination of similar primary andesitic magmas by continents.

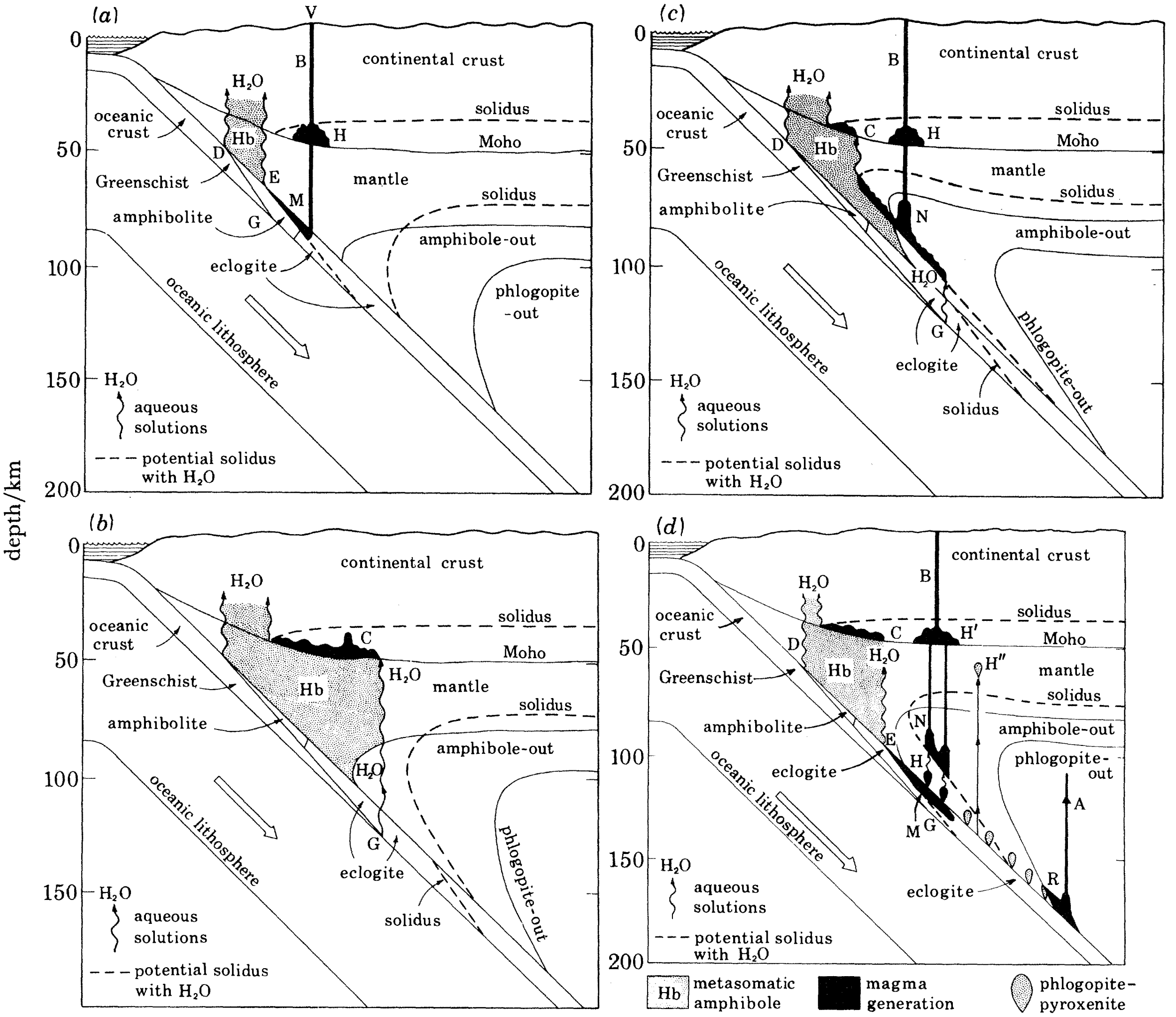

FIGURE 9. Locations of dehydration, metasomatism, and melting in subduction zones from results of experimental petrology, according to four thermal structures obtained by combinations of warm or cool (by endothermic dehydration) subducted ocean crust, with cool (figure 8*a*) or warm (by induced convection, figure 8*b*) mantle. (*a*) Cool mantle-warm crust. (*b*) Cool mantle-cool crust. (*c*) Warm mantle-cool crust. (*d*) Warm mantle-warm crust.

DEHYDRATION, METASOMATISM, AND MELTING IN SUBDUCTION ZONES

Phase boundaries for dehydration and melting in the three major structural units at an ocean-continent convergent plate boundary are shown in figure 9 for four thermal structures. The figures are designed to illustrate the kinds of processes and products that can occur and develop with various thermal structures. They are not intended to illustrate what actually happens in any particular subduction zone. The experimental basis for these diagrams was reviewed by Wyllie (1979, 1982), and Wyllie & Sekine (1982).

Dehydration occurs wherever rock masses are transported across the dehydration boundaries. The dehydration front for serpentine (D–G) approximates the greenschist facies boundary, and the dehydration front for amphibole separates amphibolite from eclogite in the subducted crust. Partial melting occurs if H_2O passes into the regions on the high-temperature sides of the dashed solidus boundaries for rock–H_2O.

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 (figure 9*a*), most solutions enter the mantle along DE, but some generate magma from subducted crust at M. With transportation downwards, the interstitial liquid may increase significantly at the depth where amphibolite melts incongruently, perhaps sufficiently to escape and rise through the mantle and into the continental crust. Influx of hydrous melt into the crust, or underplating of magmas beneath the continent, could generate additional magma from crustal rocks at H. Magmas could rise into the region B to form batholiths, with eruption represented by V. For a cool mantle-cool crust (figure 9*b*), the metasomatic solutions rising from the extended greenschist dehydration front (DG) do not enter the melting regions of either subducted crust or mantle peridotite. They can produce magmas only in the continental crust. For a warm mantle-cool crust (figure 9*c*), induced convection brings the region for potential melting of the mantle above the greenschist (DG) and amphibole dehydration fronts. The metasomatic fluids cause partial melting of mantle peridotite, with the amount of melting possibly enhanced in the region N by H_2O from dissociation of metasomatic amphibole in the mantle as it is transported downwards across the amphibole-out boundary. Figure 9*d* 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 (1982*a*, *b*, 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 vapour-absent melting defines another possible site of magma generation. Solutions released during hybridization at H cause partial melting in the shallower mantle at N.

The compositions of melts generated in the various locations illustrated in figure 9 can be estimated from the results given in figures 3–6. With magmas of different compositions rising from different sources at different depths, there are many opportunities for magma contamination, hybridization, and magma mixing in multi-stage processes. Wyllie (1982) and Wyllie & Sekine (1982) reviewed the conclusions to be drawn about magmatic products in subduction zones according to the constraints of experimental petrology.

Possible and impossible magma sources and products

Mantle peridotite yields komatiite or picrite at high pressures, and basalt at low pressures. The question of whether mid-ocean ridge basalts are derived from mantle peridotite or eclogite has profound implications for evolution of the mantle and crust. The available results from experimental petrology can accommodate either model, but physical constraints provided by solidus curves and geotherms present problems for the eclogite model, and reveal features remaining to be satisfied for models involving deep peridotite reservoirs.

Magmas from the mantle appear to be limited to compositions less siliceous than basaltic andesite, with two exceptions. In shallow, sub-oceanic mantle peridotite, high temperatures and high H_2O contents may yield magnesian-andesite or tonalite-like magmas. There is evidence from Na-free synthetic systems that under special circumstances, for peridotite with phlogopite and free H_2O, high-SiO_2 liquids with granitic affinities may be generated, but only at high temperatures, and only in amounts limited by the amount of phlogopite present. Granite and trondhjemite liquids cannot be generated from normal peridotite, nor from oceanic crust in subduction zones. However, trondhjemite and tonalite magmas of the Archean could probably be derived by partial fusion of sinking blocks of basic protocrust, with moderate H_2O content, at lower pressures and higher temperatures than those at the surface of subducted lithosphere slabs.

Hydrous granite or granodiorite liquid is the normal product of regional metamorphism of continental crust, and andesite liquid can be produced only if the crust is heated to 1100 °C or more. Granite liquid and restite can produce magmas of intermediate composition. Granite liquid cannot be produced at depths greater than about 30 km, and syenite melts may be generated in thickened continental crust.

The calc-alkaline magmatic series erupted and intruded at convergent plate boundaries make up the bulk of the rocks forming the continental crust. Basaltic andesite is a suitable candidate for primitive magmas in subduction zones, produced either by partial melting of hydrous mantle peridotite, or of relatively dry subducted oceanic crust (but only if the crust is heated to higher temperatures than those in most published thermal structures). Tholeiitic or high-Al_2O_3 basalts occurring in subduction zones (Kay & Kay 1982; Anderson, A. T. 1982) are probably derived from mantle peridotite, indicating that temperatures in the mantle wedge have approached the solidus for volatile-free peridotite. The basalts probably differentiated from picritic parents (compare figures 8, 9, and 4*a*).

Figure 9 shows that the calc-alkaline magmas at ocean–continent plate boundaries can involve components from all three magmatic hearths. Andesite is unlikely to be a primitive magma in this environment and, indeed, it has chemical and mineralogical characteristics of magmas evolved at low pressures (Gill 1981). Unravelling the details of actual processes between sites of magma generation and solidification requires a combined approach including not only geochemistry and experimental petrology, but also the physics of rheology and thermal history. Interpretation of trace element geochemistry and isotope systematics must satisfy the constraints imposed by the straightjacket of major element phase relations and phase boundaries.

This research was supported by the Earth Sciences Section of the National Science Foundation, N.S.F. grants EAR 8108626 and EAR 8206178. I thank F. M. Richter for calculating and explaining to me the geotherm in figure 7*c*.

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Discussion

M. F. Osmaston (*The White Cottage, Sendmarsh, Ripley, Woking, Surrey, GU23 6JT, U.K.*). I have two questions for Professor Wyllie. First, has he any opinion as to how the mantle water is held, or whether it is free, at depths greater than the stability field of amphibole? In the sub-oceanic upper mantle K_2O is too low for much water to be held as phlogopite.

Second, I think he does not need to be too concerned if present computations of regional isotherms around a Benioff zone fail to yield melting temperatures at the plane below the volcanic belt. The failure of the computations very likely stems mainly from inadequate modelling of the enormous (but still largely undetermined?) variation of effective viscosity near the solidus, and of a smaller effect arising from thermal conductivity variation, and not, I believe, from inadequacy of the basic simple subduction model. The observed seismicity gap or minimum below volcanic belts strongly suggests that mechanical heating at the Benioff plane is being buffered by the lubricating presence of melt. (I use the word in its widest sense of released fluid.) Local deficiency of this lubrication, whether due to melt migration or to insufficient heat, would sharply increase the local shear stress and consequent heating, possibly in a very thin zone, until lubrication is re-established. Final dissipation, in the zone, of the locally accumulated elastic strain energy could help to ensure this. Interstitial melt would also markedly lower thermal conductivity in the vicinity of the Benioff plane.

Does Professor Wyllie know of any work that bears on the effective viscosity question in particular? I suspect that a lot of very difficult work would need to be done before thermal computations could hope to answer the crucial question of whether the melt is a magma or mostly water.

P. J. Wyllie. The first question, about mantle water, has been the topic of many reviews and much speculation. I believe that most of the time, in most places, there is very little water in the mantle. Scattered phlogopite could contain essentially all of the water and potassium present in some regions of the mantle. At pressures above the normal amphibole stability limits, water may be stored in other hydrous minerals, such as phlogopite and dense hydrous magnesian silicates, or in minerals such as hydrogarnets, or simply dissolved in silicates; water may be present in fluid inclusions, which would migrate through deformation and recrystallization into intergranular fluid, or adsorbed on the surface layers of minerals; water may be dissolved in interstitial melt (e.g. in the seismic low velocity zone). I believe that water is irregularly distributed within the upper mantle, experiencing redistribution as a result of physical processes which transport mantle rocks across phase boundaries, or which cause migration of interstitial melt or other fluid phase.

The second question includes a commentary that needs comment. I am not aware of any experimental work on the effective viscosity of rocks at high pressures near the solidus, and I agree that data for this and other physical properties of rocks and partially melted rocks represent one of the great gaps in the information required to decipher processes in subduction zones, and elsewhere in the mantle. Early interpretations that partial melting in Benioff zones was caused by stress heating appear to have been abandoned, following claims that only small stress differentials could be maintained at depth in Benioff zones. The relations of magma generation at depth to localized heating through shear stress is a topic that has teased petrologists for years. The few theoretical appraisals published indicate that under suitable conditions, this relation may play a very important role in magma genesis. But because the problem is not one with which I can cope, I have concentrated on attempting to explore the possible locations of dehydration and melting reactions through a range of thermal structures encompassing the published computations. Correlating these possibilities with the structure and geochemistry of volcanic and plutonic arcs may provide limits for thermal structures, which are intimately related to rheology and motion of the rock masses in and near subducted slabs.

I must comment on the use of the word ‘melt’ to include any ‘released fluid’, and the statement about 'whether the melt is a magma or mostly water', because I think it is most important in petrological discussions to retain the distinction between melt (liquid), vapour (or dense gas), and fluid. Fluid can be used as a descriptive word to cover any material with fluid characteristics, but in the context of phase equilibria (and petrological processes), it has another, more restricted definition. Silicate rocks at high pressures and temperatures may melt to produce liquid (melt) with dissolved water and other volatile components. The liquid can coexist with an aqueous vapour (gas) with dissolved solid components and other volatile components. The liquid (melt) and the dense vapour (gas) are two separate phases with different compositions and different properties. At high pressures and temperatures, the compositions of the two phases approach each other, because of increasing solubilities of volatile components in the liquid, and of solid components in the vapour. Under some conditions, a single fluid phase which is neither liquid nor vapour, but which may be transformed into one or the other by suitable variations of composition, pressure, and temperature may exist. All experimental data so far available indicate that for normal rock compositions with H_2O and CO_2 in the upper mantle, liquid and vapour maintain their separate identities. They are both fluids in the descriptive sense, but they are not fluids related to critical phenomena. The melt can exist only at temperatures above the solidus, whereas a dense, concentrated vapour can exist through a wide range of temperatures below the solidus. The distinction among melt (liquid), vapour (gas), and fluid is important not only in crustal processes, and in subduction zone dehydration and melting, but also in mantle metasomatism, where appeals are commonly made to the effect of unspecified fluids. It is for this reason that I consider it justifiable to expend so many words in a lesson in phase equilibria, already familiar to many readers, but frequently overlooked when attempts are made to adjust the mantle composition by metasomatism in order to satisfy geochemical observations.

M. A. Morrison (*Department of Geological Sciences, University of Birmingham, P.O. Box* 363, *Birmingham B*15 2*TT, U.K.*). Current discussions of the phase transformations taking place in subducted oceanic crust and of the magma(s) that might be generated from it are usually based on the experimentally determined phase relations of basalt in the presence of variable amounts of water. Hydrothermal alteration of basic rocks by seawater involves the addition of not only water but also substantial amounts of magnesium together with extensive leaching of calcium and silica (Humphris & Thompson 1978). Geochemical and geophysical models of the oceanic crust suggest that a substantial proportion (50–20%) may have been modified in this way. Would Professor Wyllie care to speculate on the effect such changes might have on possible melt compositions generated within the downgoing slab?

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P. J. Wyllie. If we knew the composition of subducted oceanic crust at a depth of 75 km or so, we could determine the phase relations of that composition. In view of the uncertainty about the details and extent of metasomatism of oceanic crust by seawater, the best experimental approach is to start with experiments on oceanic basalt, and then to determine the effect of variations in composition. Determination of the melting relations of spilite is one of the many incomplete items on the shopping lists of experimental petrologists, along with similar relations for other possible products of hydrothermal alteration of basic rocks by seawater.

Speculations about the compositions of melts derived from subducted basalts become worthwhile only for specified bulk compositions that can be transposed into mineral assemblages, because liquid paths are controlled by mineral assemblages rather than by bulk compositions. A simple speculation for a basalt with composition modified as outlined by Dr Morrison would be that melt compositions generated at depth would follow paths similar with respect to SiO_2 and H_2O, but somewhat enriched in MgO/CaO compared with results for an unaltered basalt. This would bring the liquid compositions somewhat closer to the average calc-alkaline rock compositions. However, I believe that the major element characteristics of the calc-alkaline series are imposed, not by the subducted crust, but by processes in the overlying mantle wedge, and by shallow-level differentiation.

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