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Continental crust subducted to depths near 100 km: Implications for magma and fluid genesis in collision zones*

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Abstract—The metasedimentary pyrope–coesite–kyanite–talc–phengite–jadeite rocks of the Dora Maira Massif, Western Alps, were formed under metamorphic conditions of about 30 kbar, 700°–800°C and thus indicate a depth of subduction of some 100 km at low average geothermal gradients (7–8°C/km) along a possible geotherm of 40 mW/m^2. These pressure-temperature conditions lie above the eutectic melting curve for wet alkali granite and overlap with the melting curve of jadeite + coesite in the presence of excess H_2O. While there are no obvious signs of partial melting in the pyrope–coesite rocks, occasional interlayered bands of jadeite–kyanite–almandine–SiO_2 rock could possibly have gone through a liquid state. Because of thermal barriers such as phengite–talc or phengite–pyrope in the K_2O-MgO-Al_2O_3-SiO_2-H_2O system, it is unlikely that the pyrope-coesite rock itself represents a restite composition after extraction of a trachyte–like eutectic liquid. It is clear, however, that rocks with the hydrous mineralogy of the pyrope–coesite rock will have to melt partially when, subsequent to subduction, higher and more normal average geothermal gradients are being established at these mantle depths.

Experimental studies in the model system K_2O-MgO-Al_2O_3-SiO_2-H_2O indicate that the assemblage K feldspar–phlogopite becomes unstable, in the presence of excess water, in the pressure-temperature range 15–20 kbar and 400°–700°C and forms very MgSi–rich phengite and quartz plus a (supercritical?) hydrous, KMg–rich fluid with an estimated oxide ratio near K_2O:3MgO:7SiO_2. On this basis it can be predicted that common rocks of the upper continental crust such as granites and acid gneisses will, upon deep subduction, develop KMg–rich fluids at the expense of their biotite–K feldspar parageneses. When interacting with neighboring ultramafic mantle rocks these highly reactive fluids can cause the commonly observed mantle metasomatism producing phlogopite and K–richterite at the expense of olivine and clinopyroxene. Very K–rich igneous rocks such as those of the lamproite family, which are believed to have formed from partial melts developed in highly metasomatized mantle sources, could thus exhibit close spatial relationships with zones of preceding crustal subduction. In at least five cases (Spain, Colorado Plateau, Western Alps, Corsica, Karakorum) such relations exist. Moreover, the trace element and isotopic signature of lamproites is best explained by contamination of normal mantle with material from continental crust prior to magma formation. Magma genesis in continent/continent collision zones is severely influenced, or even governed, by the fluid and melt production within slabs of the continental crust that are subducted to considerable mantle depths.

INTRODUCTION

UNTIL RECENTLY the metamorphism of sediments, both continental and oceanic, and of other rocks of the continental crust such as granites, has been regarded as an essentially internal process within the crust of the earth attaining normally a maximum thickness of some 30–40 km. Thus, disregarding the materials of the deeper "mountain roots" below young fold belts, the maximum pressures of metamorphism endured by such rocks were believed to lie near 10 kbar, and indeed all the classical petrogenetic grids of metamorphism end at this limiting pressure (*e.g.*, WINKLER, 1974; ERNST, 1983).

Petrologic studies of metasediments from the Alpine/Mediterranean area (*e.g.*, CHOPIN and SCHREYER, 1983; SCHLIESTEDT, 1986), and especially the rediscovery and detailed examination of the exceptional pyrope–coesite rock of the Dora Maira Massif in the Western Alps (CHOPIN, 1984) have indicated, however, that this limit to depth and pressure of the metamorphism of crustal rocks

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* Dedicated to H. S. Yoder, Jr.

cannot be generally true. As a result, a newly modified and considerably expanded view of geodynamic processes involving continental crustal materials is developing. This view takes into account subduction of crustal rocks during continent/continent collisions to at least 100 km depths and perhaps even more (SCHREYER, 1985; CHOPIN, 1986a).

Within the new conceptual framework of very deep crustal subduction the problem of (partial) melting of crustal rocks at mantle depths and its possible bearing on magma genesis as a whole will have to be included. To open, or reopen, the discussion on this point using earlier experimental data is a major goal of this paper. In addition, new experimental results suggesting the evolution of highly potassic fluids from crustal rocks at mantle depths, that may cause the metasomatic changes within the ultrabasic mantle materials as known from xenoliths within kimberlites are evaluated.

THE DORA MAIRA PYROPE–COESITE ROCK

The petrology and origin of this—thus far—unique metamorphic rock was described by CHOPIN (1984), with a few additions by SCHREYER (1985). In the present context, only a brief summary can be presented, but it must be clear that the problem is not exhausted and that additional, still more detailed work will undoubtedly yield further insights.

Although the outcrops of fresh, non-retrograde pyrope–coesite rocks are limited to dimensions in the 100 m range, it is clear that the total area affected by this type of extreme metamorphism is at least 5 × 10 km (CHOPIN, 1986a). There are coarse varieties of the rock with pyrope single crystals up to 20 cm in diameter, that were initially identified as conglomerates (see VIALON, 1966), but there are also sugary quartzites with pale, pink pyropes of only centimeter size. It is in these smaller garnet crystals that inclusions of relic coesite were identified (CHOPIN, 1984). Because the typical feathery quartz aggregates formed after coesite were also found in portions of garnet that are open to the quartz matrix, Chopin concludes that all free SiO_2 of the inclusions as well as the quartz matrix had once been coesite.

In addition to pyrope and coesite (quartz) the rock contains kyanite, phengite, and talc as major constituents. Jadeite (with some 25 mol percent diopside component) is occasionally enriched in layers and irregular trails. Rutile is a common accessory mineral. The coexistence of jadeite and talc implies the instability of a partially CaMg-substituted glaucophane due to the water-conserving reacton glaucophane = jadeite + talc. However, Ca-bearing glaucophane with excess Mg is found as relic inclusions in some large pyrope crystals (CHOPIN, 1986b). Additional inclusions within the large pyrope crystals are Mg-chlorite, the new MgAlTi-silicate mineral ellenbergerite (CHOPIN *et al.*, 1986) and, locally enriched, relics of large crystals of the Mg-tourmaline dravite (see SCHREYER, 1985, Figure 14). The resulting compatibility relations are shown, in a simplified way, in the right-side portion of the AKF-plot of Figure 1. Due to the presence of small amounts of FeO in addition to MgO in the Dora Maira rock, four rather than three solid AKF-phases may coexist with coesite.

The overall chemistry of the Dora Maira rocks as deduced from their mineral contents is best explained by a sedimentary protolith such as an evaporitic clay, which had already been cited by SCHREYER (1977) to derive the talc–kyanite rocks (whiteschists), which are found elsewhere in the world in metamorphic sequences subjected to relatively high pressures. Because the whiteschist assemblage is related to the pyrope–coesite rock by the simple reaction

$$\text{talc} + \text{kyanite} = \text{pyrope} + \text{coesite} + H_2O, \quad (1)$$

it is believed that the Dora Maira material represents a higher-grade metamorphic equivalent of whiteschists.

The pressure-critical minerals and mineral assemblages making up the Dora Maira rock indicate, more or less independently, metamorphic conditions lying within the coesite stability field. Based on the assemblage pyrope–SiO_2–talc–phengite CHOPIN (1984) has derived a minimum pressure of

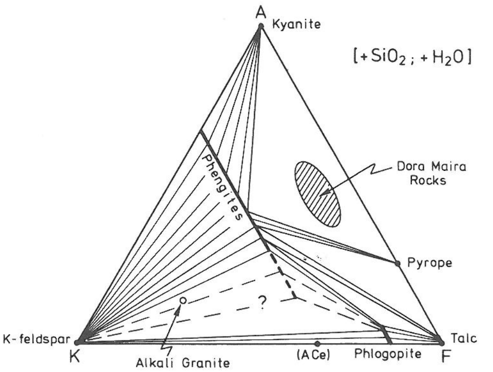

FIG. 1. Conventional AKF-diagram showing in the right portion the compatibility relations as found in the pyrope–coesite rocks of the Dora Maira Massif. The projection point for the composition of alkali granite was taken from WINKLER (1974). ACe stands for Al-celadonite, the theoretical end member of the phengite series with the formula $KMgAl[Si_4O_{10}](OH)_2$. The excess SiO_2-phase is coesite.

28 kbar and temperatures of 700°–800°C. For jadeite + kyanite, the high-pressure equivalent of paragonite, CHOPIN (1984) estimated a minimum pressure of 26 kbar. Preliminary experimental work on the stability field of the Ti-end member of the new mineral ellenbergerite (CHOPIN *et al.*, 1986) by SCHREYER, BALLER, and CHOPIN (unpublished data) indicate minimum water pressures for this phase to lie in the order of 25–30 kbar and maximum temperatures just below 800°C. This is in general agreement with the theoretical deductions by CHOPIN (1986b). Application of a new phengite barometer, for the limiting assemblage (see Figure 1) phengite + kyanite + a Mg-silicate + quartz/coesite (MASSONNE and SCHREYER, 1985) results in a water pressure of 35 kbar for the Dora Maira rock (see MASSONNE and SCHREYER, 1986b).

Conservative estimates of metamorphic conditions of the Dora Maira pyrope–coesite rock are shown in Figure 2. They are indicative of a thermal regime characterized by a very low geothermal gradient which—linearly extrapolated—is on the order of 7°/km. Such low gradients are prerequisite for blueschist facies metamorphism within subduction zones, although this type of metamorphism takes place at considerably lower pressures and temperatures (compare SCHREYER, 1985, Figure 19). Thus one may conclude that the Dora Maira metamorphism operating at some 100 kilometers depth is the higher grade equivalent of the blueschist facies. It is interesting to note, however, that the Dora Maira pressure-temperature conditions also fall close to the 40 mW/m^2 *continental* geotherm calculated by POLLACK and CHAPMAN (1977), along which all the low- as well as high-temperature lherzolite xenoliths from the Udachnaya, Siberia, kimberlite analyzed by BOYD (1984) lie.

PARTIAL MELTING IN THE DORA MAIRA ROCKS?

With the metamorphic conditions of the Dora Maira rocks located at some 30 kbar, 700°–800°C (Figure 2), the general question arises as to the possible melting behavior of crustal rocks under these conditions, and—specifically—whether or not there is any evidence of partial melting in the Dora Maira rocks. Because granites are believed to be those rocks of the continental crust that melt at the lowest temperatures, their behavior under mantle pressures is also of interest.

Pioneering experimental work on granite melting relations at pressures up to 35 kbar was performed

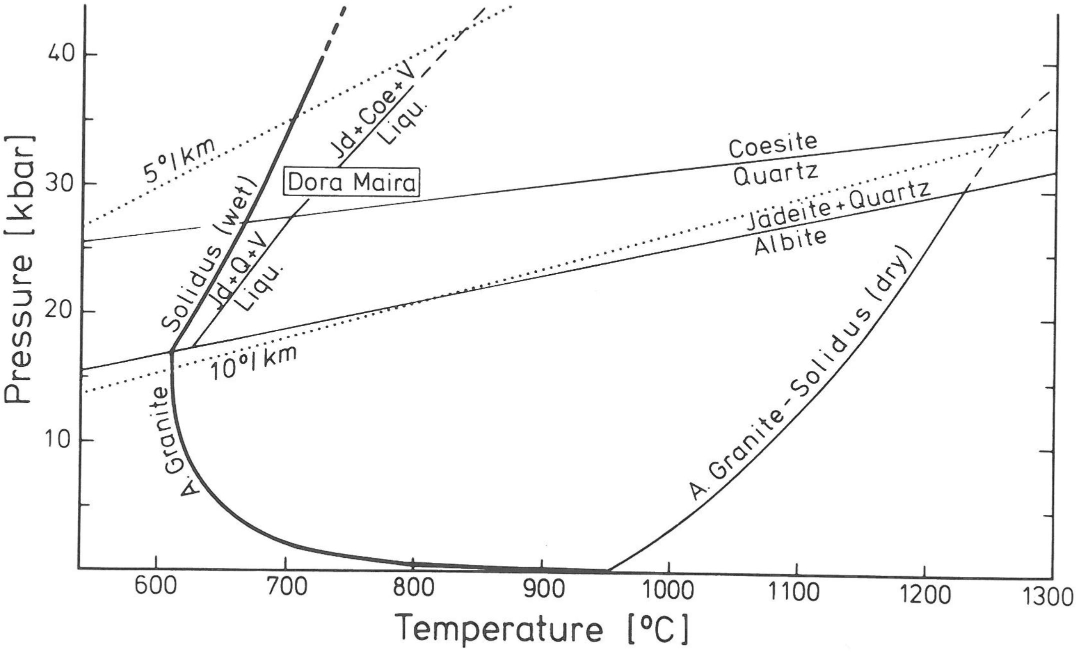

FIG. 2. Melting behavior of alkali granite as a function of pressure as simplified from the diagrams of HUANG and WYLLIE (1975, Figures 2–3). The box named Dora Maira indicates the metamorphic conditions of the pyrope–coesite rock discussed in the text. Dotted lines represent constant average geothermal gradients as labelled. Abbreviations: Coe = coesite; Jd = jadeite; Liqu. = hydrous melt; V = hydrous vapor.

during the last decade by the Wyllie school (*e.g.*, STERN and WYLLIE, 1973; HUANG and WYLLIE, 1973, 1975). Some salient features taken from the 1975 paper are reproduced in Figure 2. It can be seen from this figure that the Dora Maira pressure–temperature conditions are just above the solidus temperature of an alkali granite, provided that water pressure equalled total pressure. Because partial melting of the Dora Maira rock was not observed, CHOPIN (1984) assumed water activities well below 1.0.

An important aspect of the granite melting relations at high pressures is that granite is no longer of eutectic composition, because the eutectic melt moves toward albite composition with increasing pressure (HUANG and WYLLIE, 1975). Thus, a normal granite with a composition close to the low-pressure, crustal eutectic would—upon melting under the Dora Maira pressures—produce a more "trachytic" liquid with quartz remaining as a solid phase. Under these provisions the question arises as to whether or not the very SiO_2–rich pyrope–coesite rock of Dora Maira may in itself represent a *restite* after partial melting. A thorough examination of this possibility clearly requires more experimental melting data than presently available. Thus only a rather superficial treatment of this problem can be offered here.

The AKF-plot of Figure 1 shows the projection point of alkali granite, which lies rather close to the K (=K feldspar) corner of the diagram. Presumably, a "trachytic" liquid in the sense of HUANG and WYLLIE (1975), produced as a eutectic melt at the Dora Maira metamorphic conditions, would project into the same general area. Depending on the extension of the phengite solid solution (MASSONNE and SCHREYER, 1986a,b), it would fall either into the 3–phase field K feldspar + phlogopite + phengite, or into the 2–phase region K feldspar + phengite, and is thus clearly separated from the compositional range of the Dora Maira pyrope–coesite rocks. The phengite solid solution line, as well as the tie lines phengite–pyrope and phengite–talc, are stable during the high–pressure metamorphism and would act as thermal barriers between the pyrope–coesite rocks and the eutectic liquid. Hence, the two compositions cannot coexist; the solid assemblage cannot be the restite of "trachytic" eutectic melting. However, it must be borne in mind that this conclusion applies strictly only to compositions in the pure AKF-system K_2O-MgO-Al_2O_3-SiO_2-H_2O. Additional components such as Na_2O and FeO, which are present in subordinate amounts in the pyrope–coesite rocks, may change the situation, especially if they had been abundant initially in a hypothetical starting material.

The possibility was also considered that at least some of the numerous polymineralic inclusions within the pyrope megacrysts once represented liquid. However, this was ruled out because these inclusions are nearly free of K_2O and Na_2O, or at least strongly depleted in these elements relative to the matrix of the rock. In the Dora Maira country rocks that are more Fe–rich, no compelling textural evidence for partial melting has so far been found (CHOPIN, 1986a).

The data in Figure 2 also indicate that the estimated metamorphic conditions at Dora Maira overlap with the eutectic melting of the assemblage jadeite + coesite in the presence of excess water after HUANG and WYLLIE (1975). This may be of interest regarding the origin of the bluish jadeite–kyanite–almandine quartzite which forms "rare, decimeter-thick layers (or veins?)" within the pyrope–coesite rock (CHOPIN, 1984). Figure 3 presents a close-up view of the mutual contact of the two rock types, which—unfortunately—does not allow discrimination between igneous intrusion of the jadeite rock into the pyrope rock and subsolidus interbanding of the two different lithologies. Cross-cutting relationships were not observed elsewhere either. Nevertheless, it is possible that the jadeite–quartz rock had been partially liquid, but has subsequently recrystallized, together with the pyrope rock, to form quartz from coesite. The Fe–rich garnet composition as opposed to that of the country rock may be in favor of this melt hypothesis, but the crucial problem concerning the availability of excess water cannot be solved.

In summary, the specific question "Was there any partial melting in the Dora Maira rocks during the high–pressure metamorphism?" cannot be answered with present knowledge. More important, however, is the result from Figure 2 that the temperatures during this metamorphism were sufficiently high to produce melts from units of appropriate bulk composition. As mentioned in the introduction, just the existence of the Dora Maira pyrope–coesite rock implies that rocks of the continental crust may be subducted to depths of about 100 km. Both SCHREYER (1985) and CHOPIN (1986a) have emphasized that the geodynamic conditions required for such deep subduction are still much easier to understand than those necessary for the retrieval of the high–pressure rocks without significant mineralogical changes back to the surface. Therefore, the chances are indeed very high that the subducted continental materials remain at

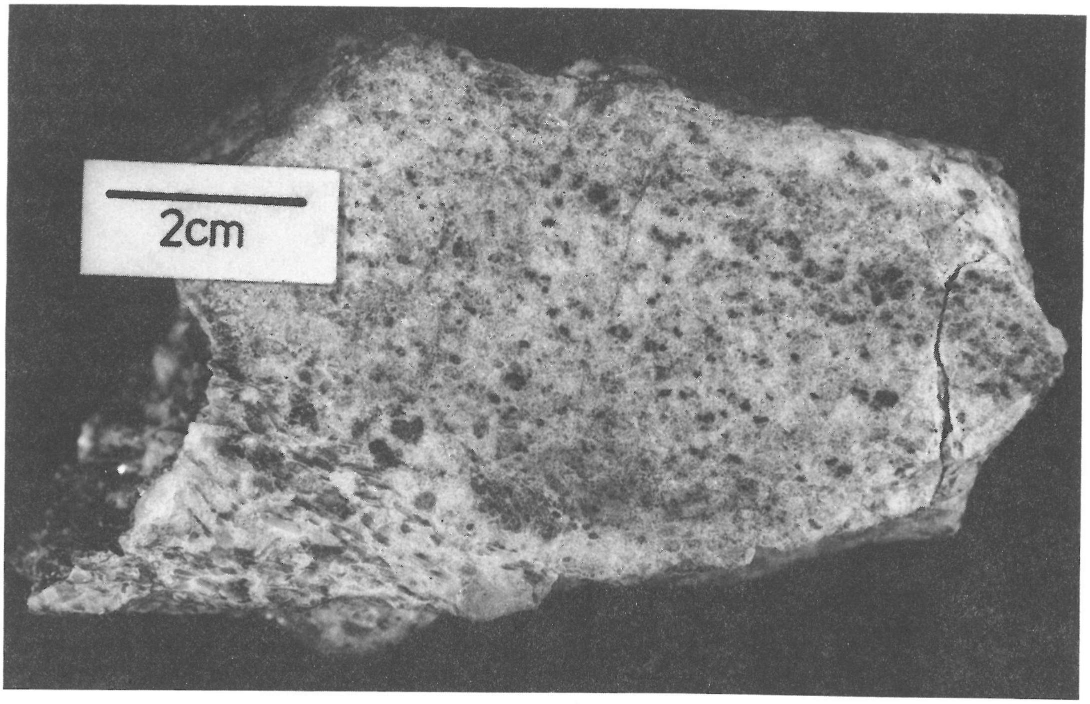

FIG. 3. Concordant contact between schistose, phengite–bearing pyrope rock (bottom) and a more massive layer of jadeite–kyanite–almandine quartzite (top). Parigi, Italy.

mantle depths sufficiently long for higher, more normal geotherms to develop (see Figure 2), which would then inevitably lead to increasing proportions of melt in these materials. These crust-derived, acid melts could then react with neighboring mantle peridotite, that is they "may experience hybridization" and produce phlogopite–bearing pyroxenities, similarly as proposed by SEKINE and WYLLIE (1982, 1985) for magmas above subducted oceanic crust.

At any rate, any model of magma genesis in continent/continent collision zones must necessarily take into account the contribution by melting of subducted continental crust. However, as we hope to show in the two final sections of this paper, there may be an additional contribution of material from continental rocks by virtue of a mechanism other than eutectic melting.

INSTABILITY OF THE K FELDSPAR–PHLOGOPITE ASSEMBLAGE AT HIGH FLUID PRESSURES

In their reports on experimental studies of phengite synthesis and stability in the system K_2O-MgO-Al_2O_3-SiO_2-H_2O MASSONNE and SCHREYER (1986a,b) emphasized that, for a fixed pressure and temperature, a critical phengite that is useful as a geobarometer must coexist with the limiting assemblage K feldspar–phlogopite–quartz. This is shown, in a simplified fashion, by the AKF diagram of Figure 1. In the course of determining the pressure–temperature stability range of this limiting assemblage the surprising discovery was made that it becomes unstable at the comparatively moderate water pressure of 15–20 kbar. Specifically, the tie line between phlogopite and K feldspar is broken, in the presence of excess water, by a reaction to form the solid phases phengite and quartz. Because the theoretical endmember of the dioctahedral mica series Al–celadonite, $KMgAl[Si_4O_{10}](OH)_2$, which lies on the join K feldspar–ideal phlogopite, $KMg_3[AlSi_3O_{10}](OH)_2$, (see Figure 1), cannot be synthesized (MASSONE and SCHREYER, 1986a), the phengite formed by the above reaction must contain less MgSi but nevertheless represent the critical composition. In order to account for the bulk composition along the K feldspar–phlogopite join, an additional phase must be present which, however, is not observed in the X-ray diffractograms of the quenched run products. For stoichiometric reasons this phase must be very poor in Al, or free from it, and consist essentially of K and Mg together with additional amounts of SiO_2 and H_2O. Earlier investigations of the system K_2O-MgO-SiO_2-H_2O by SEIFERT and SCHREYER (1966) at the much lower pressure of 1 kbar have indicated that hydrous ternary KMg silicate melts may appear in this system at surprisingly low temperatures (<500°C). It can be predicted that, with increasing fluid pressures, these melts will dissolve more and more H_2O and become low-viscosity, perhaps supercritical fluids with rather variable K:Mg:Si ratios. Therefore, it is most likely that the "missing components" in the experiments of MASSONE and SCHREYER (1986b) are contained, under run conditions, in such a K, Mg-rich fluid. Indeed, the water observed after opening the quenched run capsules was often found to yield a precipitate when drying up.

In Figure 4 the AKF plot has been extended to include the Al-free border system, in which the assumed K, Mg-rich fluids lie. It can be seen that instead of the assemblage K feldspar + phlogopite the limiting phengite coexists with a series of fluid phases that are presumed to be confined to the system K_2O-MgO-SiO_2-H_2O. According to MASSONNE (1986) the fluid in equilibrium with phengite for the bulk composition of Al-celadonite (see above) plus excess water has an estimated composition given on an anhydrous basis as $K_2O \cdot 3MgO \cdot 7SiO_2$. In increasingly more potassic bulk compositions K feldspar appears first as a second solid phase that coexists with a KMg-fluid of invariant K/Mg-ratio, and then phengite disappears with only K feldspar + fluids remaining.

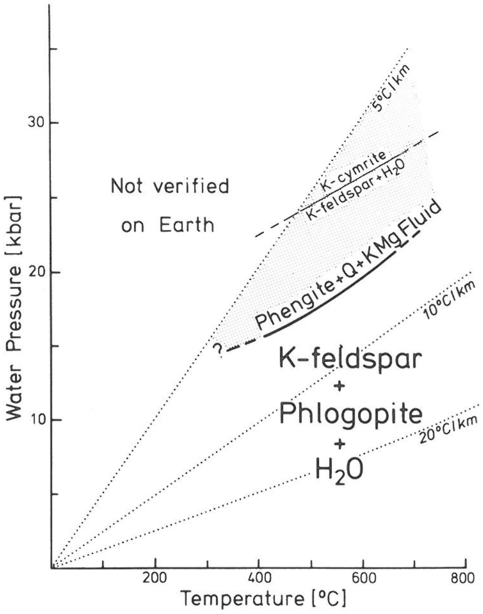

FIG. 5. Water pressure/temperature plot showing the approximate stability field of the assemblage K feldspar-phlogopite and the location of its high-pressure breakdown to phengite + quartz + KMg-rich fluid (see Figure 4). The hydration curve of K feldspar to form K-cymrite, $KAlSi_3O_8 \cdot H_2O$, is taken from HUANG and WYLLIE (1975). Dotted lines represent constant geothermal gradients as labelled. Thus the shaded field, in which the KMg-rich fluid exists, can only be reached in areas with a low thermal flux such active subduction zones.

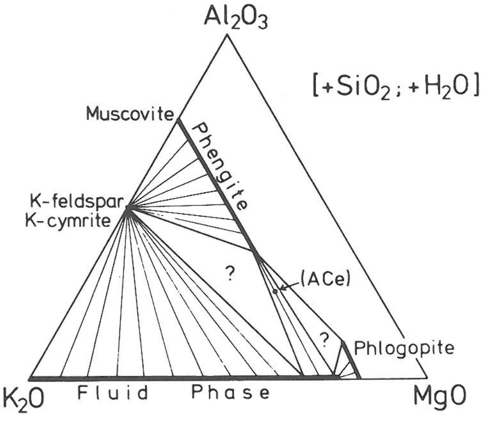

FIG. 4. Excess-SiO_2 and excess water plot of the system K_2O-MgO-Al_2O_3-SiO_2-H_2O showing compatability relations of solid and fluid phases at fluid pressures above 20 kbar and temperatures of about 450°–650°C. ACe is Al-celadonite as in Figure 1, K-cymrite is the phase $KAlSi_3O_8 \cdot H_2O$. For relevant reaction curves see Figure 5. To simplify the projection all solid solutions are shown as binary. The most aluminous phlogopite coexisting with phengite is less aluminous than ideal phlogopite, $KMg_3[AlSi_3O_{10}](OH)_2$ according to the data of MASSONE and SCHREYER (1986b). The compositional range of the KMg-rich fluid phase is hypothetical.

Figure 5 shows the location of the univariant curve for the reaction of K feldspar and phlogopite in the presence of excess water as determined by MASSONNE (1986) in the temperature range 450°–650°C. In the case of water undersaturation, K feldspar + phlogopite were found to coexist with phengite + KMg-fluid towards higher pressures. At still higher total pressures K feldspar itself reacts with H_2O to form the solid K feldspar hydrate phase named K-cymrite here (Figure 5). This phase is the potassium analogue of the mineral cymrite, $BaAl_2Si_2O_8 \cdot H_2O$, and has the composition $KAlSi_3O_8 \cdot H_2O$. SEKI and KENNEDY (1964) first synthesized this phase, but their reaction curve had to be modified on the basis of more recent work. (HUANG and WYLLIE, 1975; MASSONNE, unpublished data). Thus, at high fluid pressures on the order of 30 kbar, phengites near the critical composition may coexist with KMg-fluid, coesite, and K-cymrite (compare Figure 4).

CONSEQUENCES FOR MANTLE METASOMATISM AND RELATED LAMPROITIC MAGMAS

As a result of the phase relations depicted in Figures 4–5 a petrogenetic mechanism evolves that may be of considerable interest regarding the interaction of deeply subducted crustal rocks with mantle materials and the genesis of ultrapotassic lavas. The paragenesis of K feldspar with trioctahedral potassic micas (phlogopite, biotite) is widespread within the continental crust, because it occurs in such common upper crustal rocks as granites and many acid gneisses. Therefore, during deep subduction of crustal materials in collision zones, along relatively low geotherms (6–8°C/km), the reaction curve shown in Figure 5 for pure Mg-micas will be transgressed at depths of some 50–70 km, and—in the presence of a hydrous gas phase that may be formed during progressive subduction zone metamorphism—the KMg-rich fluid should develop. Presumably, the amount of fluid forming would be governed by the amount of interstitial H_2O gas present initially, and there is also an unknown influence of other components hitherto neglected such as Fe-oxides, TiO_2, and Na_2O.

Because of high reactivity of the KMg–fluids there is virtually no chance that they can ever reach the earth's surface, although their enrichment in K and Mg is reminiscent of lamproitic and perhaps K–rich kimberlitic igneous rocks. However, these rocks also contain other components such as Al_2O_3 and CaO, and all estimates and determinations of melting temperatures of lamproitic rocks yield much higher values (~800°–1000°C; see WAGNER and VELDE, 1986) than the temperatures of fluid formation (450°–650°C, Figure 5). Particularly interesting relations arise when one envisions that the KMg–fluids produced from subducted crustal rocks react with neighboring ultramafic mantle materials.

Petrological observations on mantle xenoliths from kimberlites have provided compelling evidence for mantle metasomatism (*e.g.*, DAWSON and SMITH, 1977; JONES *et al.*, 1982; BAILEY, 1982). Among the newly formed minerals that replace olivine and clinopyroxene the phases phlogopite, and K–richterite, ideally $KNaCaMg_5[Si_8O_{22}](OH)_2$, predominate. Although there is agreement that this metasomatism is due to infiltration of peridotites and harzburgites by fluids, the sources and origins of these fluids are still under debate (RYABCHIKOV *et al.*, 1982; WYLLIE and SEKINE, 1982; SCHNEIDER and EGGLER, 1986). It is clear from the present study that the KMg–fluids emerging from subducted crustal rocks containing biotite and K feldspar would represent almost ideal agents for the mantle metasomatism observed. The additional enrichment of the metasomatites in Ba, Ti, and other large–ion–lithophile–elements may be explained by their former presence in the biotite that reacted to form the metasomatizing fluid.

If mantle metasomatism was indeed, or at least partly, caused by fluids originating from subducted crustal rocks, a spatial relationship between the metasomatites in the mantle and relics of crustal rocks could be expected at least in some cases. An example may be provided by the xenolith suites from South African kimberlite pipes which comprise, on one hand, metasomatites and the MARID (mica–amphibole–rutile–ilmenite–diopside) "pegmatites" (DAWSON and SMITH, 1977) and, on the other hand, rocks that can only be explained through crustal origin. Regarding the latter ones, xenoliths of kyanite-bearing grospydites from the Roberts Victor Mine (SMYTH and HATTON, 1977) are particularly interesting, because they contain K feldspar and coesite as well. We wish to emphasize, however, that—to our knowledge—in none of the mantle xenoliths the assemblage K feldspar + phlogopite was found.

There seems to be general consent that the K–rich magmas of the lamproite family are products of partial melting of a large–ion–lithophile–element enriched mantle source such as the metasomatites discussed above (see JAQUES *et al.*, 1984; NIXON *et al.*, 1984; WAGNER and VELDE, 1986). The high temperatures of the lamproite magmas cited before require high temperatures of magma genesis in the mantle. Therefore, we conclude that the production of these melts occurred considerably later than the metasomatism which—if effected by the mechanism proposed here—was a low–temperature process along the initially low subduction zone geotherm. The lamproite melts may have formed when more normal, post–subduction geotherms of the mentle were established, or in fact even much later during a subsequent thermal event.

Disregarding the variable time difference between subduction-induced mantle metasomatism and lamproite melt formation, the mechanism proposed here should imply a close *spatial relationship* between lamproite occurrences and underlying former collision zones. Although this problem has not been tested in any systematic way, there are indications from the regional geology of some lamproite occurrences that such relationships may indeed exist.

At least some of the lamproitic and related rocks of the Western United States, including minettes and kimberlites, may overlie subduction zones, because metamorphic rocks typically formed along subduction geotherms are in fact found as xenoliths within them (HELMSTAEDT and SCHULZE, 1977). Still more importantly, such xenoliths contain phengites that exhibit the highest MgSi-contents found in nature (SMITH and ZIENTEK, 1979; HARLEY and GREEN, 1981) and thus represent critical phengites that may have coexisted, at some stage, with the KMg–fluid discussed. Similar relationships hold for the lamproites of southeastern Spain (WAGNER and VELDE, 1986), for which most models of the tectonic evolution involve young subduction (see NELSON *et al.*, 1986). Indeed, in the clearly subduction-zone metamorphosed crustal slab of the Sesia-Lanzo zone in the Western Alps, some 100 km north of the Dora Maira Massif, "K–rich lamprophyre" dykes (DAL PIAZ *et al.*, 1979) of Oligocene and thus postsubduction age are found, which have compositions practically identical to lamproites. DAL PIAZ *et al.* (1979) emphasize the similarity of their "lamprophyres" to postcollisional lamprophyres of Karakorum (VITERBO and ZANETTIN, 1959) as well as to a minette intruded into the previously subducted schists of Corsica (VELDE, 1967; WAGNER and VELDE, 1986). Thus there are at least five regions in the world, where a spatial relationship between lamproite occurrences and former collision zones exists. For the diamond–bearing lamproites of the Kimberley region, Western Australia, no such relationship can be identified at present, because the 20 million year old intrusions occur near a mobile belt of Proterozoic age (JAQUES *et al.*, 1984).

The sequence of events as proposed by the present model, that is 1) deep subduction of crustal rocks, 2) mantle metasomatism, 3) genesis of K–rich magmas, is also supported by geochemical data on lamproites, especially those from Spain. HERTOGEN *et al.* (1985) have successfully modeled the isotopic and peculiar trace element patterns of the mantle source region of these lamproites by contamination of normal mantle with material released from subducted continental sediments. Similarly, NELSON *et al.* (1986) conclude that the metasomatic component added to the mantle below southeastern Spain prior to lamproite formation "has the Sr, Nd, and Pb isotopic characteristics of continental crust or sediments derived from continental crust."

To summarize the implications of deep subduction of continental crust into the upper mantle in continent/continent collision zones it is clear that both the development of melts and of low–temperature fluids within the subducted crustal slabs must have profound effects on the genesis of magmas in these areas, and may even govern them.

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