Subduction of Continental Crust to Mantle Depths: Petrological Evidence

by Werner Schreyer

Metamorphism of rocks from the continental crust does not end at the conventional crust/mantle boundary, some 35 km deep with a pressure near 10 kbar. Because of deep subduction during continent/continent collisions, this process can continue to at least 30 kbar. This article reviews recent findings of high-pressure minerals and mineral assemblages in continental metasediments of the Western Alps, as well as some of the underlying experimental phase equilibria results. Guidelines are summarized for recognizing this new "ultrahigh-pressure metamorphism," which is only rarely preserved in nature because of the common subsequent re-equilibration (retrograde metamorphism) of these rocks during obduction to shallower levels.

Introduction

A classical concept for decades now has been that the continental crust acts as a sink for products of the differentiation processes that the Earth has undergone since it was formed through accretion as a planet some 4.5 Ga ago. It was thus believed that the composition of the average continental crust, which is dominated by granites, gneisses and (meta-) sediments and also some basaltic rocks, is really that of the sum of various solidified melts that were produced in and ejected from the mantle during the geological past. The enormous variety of sediment compositions, which are generally very distinct from any melt compositions, is due to weathering and other exogenic processes that may cause local enrichments of elements such as Si, Al, and Ca. Nevertheless, the average composition of all these sediments is very similar to that of the average of all acidic to ultrabasic igneous rocks (e.g. Sibley and Wilband, 1977), which are believed to have their ultimate source in the mantle or its chondritic predecessor.

If this simple picture is correct, the continental crust should have grown throughout the Earth's history, while the mantle became more and more depleted in those elements enriched in the crust. However, the now widely accepted plate tectonic concept, which implies that oceanic lithosphere is subducted to great depths in the mantle, may already call for modification of this simple model. While the depleted oceanic mantle and the overlying basaltic crust would simply be reincorporated into the mantle without causing much overall change in its composition, this can hardly be true for the incorporation of oceanic sediments. But these are, on a large scale, negligible quantities and, moreover, many geoscientists believe that such thin sedimentary skins on the downgoing oceanic slabs are scraped off at shallow depths within the subduction zones.

Thus, despite plate tectonics, the simple model of an essentially "one-way street" for materials from the mantle into the continental crust has been maintained in most

geochemical concepts. It also seemed to be corroborated until recently by petrological studies on the metamorphism of rocks of the continental crust. The classical pressure-temperature diagrams for the facies subdivision of metamorphic rocks reach maximum temperatures of about 1000°C and maximum pressures near 10 kbar, which are indeed the lithostatic pressures prevailing at the base of the continental crust with a normal thickness of 30-40 km (Fig. 1). Although greater crustal thicknesses up to 60 km were recognized long ago by geophysical methods as the so-called "roots" of young orogenic belts, no petrological evidence could be presented until recently that crustal metamorphic rocks now exposed at the surface had ever reached such great depths.

The simple "one-way street" model from mantle to continental crust has lately been questioned by two branches of the geosciences. On the one hand, some geochemists have reported "crustal fingerprints" in igneous rocks that were thought to have crystallized from "primary" mantle sources (Allégre and Turcotte, 1985; Jaques, 1987). However, this evidence may not always be compelling, especially in view of the controversial interpretation of isotope data. The second and independent line of reasoning is the petrological evidence that rocks of the continental crust were metamorphosed at depths near 100 km. This evidence of paramount importance is based on equilibrium phase relations of

Figure 1: Schematic profile through the upper 100 km of a continent.

rock-forming minerals that were calibrated by high-pressure experiments. Some salient features of these new discoveries are reviewed in the following.

Experimental and Metamorphic Petrology

Experimental mineralogists and petrologists have been studying the high-pressure, high-temperature behaviour of minerals and rocks for many years. The experimental PTX (pressure, temperature, composition) conditions, however, were generally chosen according to the accepted geologic concepts of rock distribution within the crust and the mantle. Thus, while ultrabasic and basic rocks were studied both under crustal and mantle pressure, typical crustal materials like sediments and acid gneisses seemed to require only work up to pressures near 10 kbar. One exception was the study of simple model systems imitating clay sediments that were subjected to higher pressures. These model systems contained only 4 or 5 oxide components instead of the 10-15 of natural rocks. Surprisingly, new solid phases and assemblages were synthesized that had not been recorded from any natural rocks, so these studies seemed to be purely of academic interest. But the surprise was even greater, when it turned out that such phases and assemblages do occur in metamorphic rocks after all.

A first example is illustrated in Figure 2. In the model system $\text{MgO-Al}_2\text{O}_3\text{-SiO}_2\text{-H}_2\text{O}$ the assemblage talc plus kyanite is seen to be stable only at high water pressures, above 4 kbar, over a considerable temperature range. It replaces the very common assemblage chlorite + quartz, which occurs at low PT conditions in young sediments as well as in many "normal" crustal metamorphic rocks of low to medium grades (e.g. greenschists). The previously unknown talc-kyanite assemblage was first synthesized in the laboratory and subsequently found to be rock-forming in many metapelites that had been subjected to high pressures, such as the whiteschists of Schreyer (1977, 1985).

Figure 2: Water PT stability field (heavy solid lines) of the mineral assemblage of talc with kyanite. The relevant chemical model system $\text{MgO-Al}_2\text{O}_3\text{-SiO}_2\text{-H}_2\text{O}$ is shown in a projection from water in the triangular inset. After Schreyer (1988).

Pyrope-Coesite Assemblage of the Dora Maira Massif

In 1984 Chopin described a spectacular rock from the Dora Maira Massif of the Western Alps containing talc plus kyanite and very large crystals (Fig. 3) of almost pure pyrope ($\text{Mg}_3\text{Al}_2\text{Si}_3\text{O}_{12}$), which coexist with a SiO_2 phase, generally quartz. Some pyropes contain relics of coesite,

the high-pressure phase of SiO_2 , partly transformed to quartz and surrounded by radial cracks due to the necessary volume expansion (Fig. 4).

Figure 2 shows that the assemblage pyrope + SiO_2 becomes stable at the expense of the talc-kyanite pair at temperatures near $750^\circ\text{--}800^\circ\text{C}$ and at water pressures above at least 25 kbar. In fact, a pyrope + coesite assemblage requires a minimum pressure near 30 kbar. Thus the Dora Maira rocks exhibit a progressive "ultrahigh-pressure" metamorphism that had produced talc-kyanite schists and just attained the reaction boundary of Figure 2 where this assemblage reacts to form pyrope + coesite. The calculated depth of metamorphism (based on an assumed rock density of 2.8 g/cm^3) is, therefore, about 100 km. The linear geotherms shown by thin solid lines in Figure 2 indicate that this type of metamorphism takes place along rather low geothermal gradients characteristic of subduction zones in general (Ernst, 1973).

Figure 3: Euhedral crystal of pyrope still attached to portion of its fine-grained, quartz-bearing country rock. Parigi, Italy.

For full appreciation of this sensational find it is important to discuss some basic geological features of the pyrope-coesite rock occurrence. Although clearly crystallized under PT-conditions of the upper mantle, these rocks are by no means xenoliths found in mantle-derived igneous rocks such as kimberlites. They occur over an area of at least $5 \times 10\text{ km}$ as one particular type of metamorphic rock in a sequence of regionally metamorphosed sediments and igneous rocks that belong to the so-called "polymetamorphic unit" of the Dora Maira Massif (Chopin, 1987). The age of the metamorphism is probably early Alpine, around 110 Ma (Goffé and Chopin, 1986). There can be no doubt from the chemistry of the pyrope-coesite rock that its protolith was of sedimentary origin of unknown age. The extremely high Mg/Fe ratio indicates that the best estimate is a magnesian clay of an evaporite environment, as discussed for the whiteschists by Schreyer (1977).

In view of current models for the origin of the Alps, it seems that this ultrahigh-pressure metamorphism took place during the early stages of the collision between the African and European plates, both of which carried continental crusts. According to Goffé and Chopin (1986) the European continental crust including the Dora Maira rocks was subducted deeply and steeply under the African plate. For the return of these rocks to the surface the authors envisage that those parts of the European plate distal to the subduc-

tion zone broke up into smaller units, which were, in turn, underthrust below the already subducted portion. This resulted in backthrusting and upward transport into shallower levels (i.e. obduction) of these earlier subducted plate units.

Other High-Pressure Phases

As regard the problem of deep subduction of the crust, experimental studies in the system $\text{MgO-Al}_2\text{O}_3\text{-SiO}_2\text{-H}_2\text{O}$ (MASH) have turned out to be particularly important for they show an unexpectedly large number of crystalline phases that are only stable at relatively high to very high pressures. Thus, starting with Mg-enriched clay protoliths, there must be drastic differences in the mineral contents of their metamorphic equivalents, depending on whether metamorphism took place at relatively low pressures within a normal, 35 km thick crust, or at high pressures within a continent/continent collision and subduction zone.

Figure 4: Inclusion of a fractured crystal of the coesite (center) within pyrope, which developed radial cracks around this inclusion. White areas between coesite and pyrope are occupied by quartz formed through partial conversion of coesite. Large transparent inclusion in right portion is kyanite, small black one rutile. Field is 1 mm wide. Parigi, Italy.

Figure 5 shows, in a schematic way, the PT stability fields of six of the high-pressure phases of the MASH system. It is interesting to note here that some of these phases were first discovered as new minerals in natural rocks, whereas others were first obtained in the synthesis experiments and may, or may not, have natural counterparts detected later in high-pressure rocks.

The mineral Mg-carpholite with a composition close to the ideal end-member, $\text{MgAl}_2\text{Si}_2\text{O}_6(\text{OH})_4$, was found in metabauxites of the French Alps (Goffé, 1980). It stimulated synthesis work, which was soon successful (Chopin and Goffé, 1981). Mg-carpholite stability (Fig. 5) begins near 6 kbar, and 400°C (Chopin and Schreyer, 1983) and ends near water pressures close to 50 kbar (Schreyer, 1988).

For the mineral chloritoid, $(\text{Mg}, \text{Fe}^{2+})\text{Al}_2\text{SiO}_5(\text{OH})_2$, only crystals with up to about 70 mole % of the Mg-chloritoid end-member are known from natural rocks. Nevertheless, the pure phase was synthesized in the MASH system, requiring a minimum water pressure of about 18 kbar (Chopin and Schreyer, 1983). Like Mg-carpholite, it persists up to about 50 kbar (Fig. 5).

Yoderite, $\text{Mg}_2\text{Al}_6\text{Si}_4\text{O}_{18}(\text{OH})_2$, was found in Eastern Africa as a spectacular rock-forming mineral (McKie, 1959), and it was later synthesized by Schreyer and Yoder (1968). However, the stability field (Fig. 4) seems to be limited to the small pressure range of about 9-19 kbar at relatively high temperatures (Schreyer, 1988).

Pyrope garnets, with approximately 70% of the ideal end-member $\text{Mg}_3\text{Al}_2\text{Si}_3\text{O}_{12}$, have been known for a long time as characteristic minerals of garnet peridotites of the upper mantle. The pyrope crystals of the Dora Maira rocks attain 98% of this end-member, thus most closely approaching the pure synthetic pyropes that were synthesized as early as 1955 by Coes. The stability field of pyrope shown in Figure 5 extends to very high pressures, and an upper pressure limit for this garnet will only be reached, when silicon changes its coordination against oxygen in solids from four to six such as in the $\text{Mg}^{[12]}\text{Si}^{[6]}\text{O}_3$ phase with a perovskite-type structure (figures in brackets indicate coordination numbers).

Mg-staurolite, about $\text{Mg}_4\text{Al}_8\text{Si}_8\text{O}_{46}(\text{OH})_2$, the pure Mg analogue of the common, but Fe-rich, rock-forming mineral staurolite was synthesized in the MASH system by Schreyer (1968), but it was only recently detected as a natural mineral (up to 95 mole % end-member) in rare inclusions within the pyrope crystals of Dora Maira (Chopin, 1987). Its stability field (Fig. 5) covers a wide range from about 13 to far beyond 50 kbar (Schreyer, 1988).

MgMgAl-pumpellyite, $\text{Mg}_5\text{Al}_5\text{Si}_6\text{O}_{21}(\text{OH})_7$, is the newest synthetic high-pressure phase in the MASH system (Schreyer et al., 1986). It has, thus far, no natural counterpart, and it is indeed somewhat doubtful as to whether or not it can occur within the Earth, because there the possible PT-conditions are generally considered to lie above the $5^\circ\text{C}/\text{km}$ geotherm. Figure 5 shows that the stability field of MgMgAl-pumpellyite (Schreyer, 1988) begins at about 37 kbar, 500°C , and lies thus outside this range. Nevertheless, this synthetic phase is important for subduction zone metamorphism, for it may provide an upper pressure or lower temperature limit to this type of metamorphism by virtue of the presence of alternative mineral assemblages such as pyrope + kyanite + coesite or pyrope + kyanite + talc, both of which are actually encountered in the Dora Maira rocks (Chopin, 1984). It is also of interest to note from Figure 5 that the high-pressure, low-temperature phase MgMgAl-pumpellyite lies entirely within the stability field of diamond.

Figure 5: PT stability fields (heavy lines) of selected high-pressure phases in the MASH system, as projected in the upper left portion. Pressure/depth relation and linear geotherms as in Figure 2. After Schreyer (1988).

Important additional high-pressure phases and assemblages are obtained simply by adding the component K_2O to the synthetic MASH system. This allows the formation of the white mica phengite, a derivative of muscovite with the formula $K(Al_{2-x}Mg_x)^{[6]}[Al^{[4]}_xSi_{3-x}O_{10}](OH)_2$, which was recognized as early as 1963 by Ernst to be characteristic of high-pressure, low-temperature (blueschist) metamorphism. As shown in Figure 6 in a modified AKF plot, the phengite solid solution series extends from pure muscovite, $KAl_2[AlSi_3O_{10}](OH)_2$, by virtue of the inverse Tschermak substitution of $Mg^{[6]}+Si^{[4]}$ for $Al^{[6]}+Al^{[4]}$, towards a theoretical end-member $KMgAl[Si_4O_{10}](OH)_2$ given the name Al-celadonite.

Recent experimental work has shown that Al-celadonite cannot be synthesized under any conditions of water pressure and temperature. The most Si-rich phengites described have about 80 mole % of the theoretical end-member (Massonne and Schreyer, 1986), and white micas with such compositions have been found as mineral constituents of xenoliths ejected by kimberlites.

Figure 6: Triangular projection of portions of the model system K_2O - MgO - Al_2O_3 - SiO_2 - H_2O showing the compositional range of phengites and their possible coexistence at high pressures with other minerals in the system. Note particularly the assemblage phengite-talc-pyrope-kyanite (+ an SiO_2 -phase), which was found at Parigi, Italy (Chopin, 1984). The numbers refer to the Si content of phengites per formula unit, which is 3.0 for pure muscovite, $KAl_2AlSi_3O_{10}(OH)_2$.

Still more importantly, the degree of maximum $MgSi$ -substitution in the phengites (Fig. 6) can be shown experimentally to increase linearly with water pressure and decrease with temperature (Massonne and Schreyer, 1987). Figure 7 depicts the Si-isopleths of phengites as a function of pressure and temperature. Note that the Si content increases with pressure and decreases with temperature. The isopleths apply only to those phengites that coexist, at the requisite end of the substitution line of Figure 6, with their limiting assemblage K-feldspar + phlogopite + SiO_2 + H_2O -fluid.

Reversing the argument, phengites occurring together with this limiting assemblage can be used as rather sensitive geobarometers. Since the dP/dT slopes of the isopleths in Figure 7 are fairly low, the temperatures of metamorphism do not have to be known with a very high degree of accuracy for successful phengite geobarometry. Massonne and Schreyer (1987) have discussed the influences of additional components like Na, Fe, and F as well as of reduced H_2O activity on the phengite barometer. Since biotite + phengite (or muscovite) + K-feldspar + quartz assemblages are com-

mon in granites, metagranites, and meta-arkoses, this method of geobarometry may find a wide range of applications.

The AKF-plot of Figure 6 also shows that phengite-bearing assemblages other than the limiting one just discussed (with phlogopite + K-feldspar + SiO_2) must necessarily contain phengites with lower Si-contents. The Dora Maira rocks with their assemblage pyrope + coesite + kyanite + phengite serve as a good example. This phengite contains only 3.55 Si per formula unit (Chopin, 1984), whereas the maximum phengite for the metamorphic condition of Dora Maira (30 kbar) should have 3.8 Si (see Fig. 7). Such highly siliceous phengites were not reported thus far from any of the Dora Maira rocks. However, at this locality there is no K-feldspar + phlogopite + SiO_2 -assemblage that could be attributed to the highest grade of metamorphism attained.

Among its constituent assemblages the Dora Maira rocks also contain the rather unusual mineral pair phengite + talc (Chopin, 1984). The coexistence of these two minerals in the presence of excess SiO_2 (see Fig. 6), which prohibits other well known pairs like phlogopite + kyanite (+ SiO_2), or phlogopite + chlorite (+ SiO_2), was first discovered in a high-pressure rock from Yugoslavia (Abraham and Schreyer, 1976), but has since been found in many other rocks from the Western Alps (Chopin, 1981). Experimentally, this assemblage was first synthesized at high water pressures by Schreyer and Baller (1977), and Figure 8 shows its stability field as determined in largely unpublished work by Massonne and Schreyer. The overlap of this PT-field with the geotherms between $5^\circ C/km$ and $10^\circ C/km$ associated with subduction-related metamorphism explains why this assemblage, which is not easy to recognize by standard petrographic techniques, is common and actually diagnostic for this kind of metamorphism.

Less Informative and Non-Diagnostic Minerals

The observant reader will have recognized by now that the minerals and assemblages dealt with thus far are exclusively those which - from a bulk chemical point of view - may occur in highly magnesian rocks such as the rare evaporitic

Figure 7: Water PT plot showing the upper thermal stability of muscovite and the limiting stability conditions for phengites of various compositions (numbers indicate Si contents as explained in Figure 6) in assemblage with K-feldspar + phlogopite + quartz + H_2O (compare Fig. 6, lower portion). After Massonne and Schreyer (1987).

metapelites mentioned. What about other, more common protoliths like normal pelites carrying Na, Ca, and Fe as well, sedimentary ironstones, or basalts and other igneous rocks that get involved in subduction zone metamorphism? Undoubtedly, such rocks will develop high pressure mineral assemblages as well, but their breadth of variation is not comparable to that of the magnesian systems and, thus, they are accordingly less useful, even non-diagnostic, for subduction zone metamorphism.

Figure 9 presents a selection of stability fields and reaction curves from Ca, Na, Fe and Mn-bearing systems significant for subduction zone metamorphism. Salient features are, of course, the breakdown reactions of the plagioclase end-members albite and anorthite that form jadeite ($\text{NaAlSi}_2\text{O}_6$)- and grossular ($\text{Ca}_3\text{Al}_2\text{Si}_3\text{P}_{12}$)-bearing assemblages, respectively. Plagioclase breakdown under high pressures is also the fundamental mechanism for the basalt to eclogite transition, which leads to the high-pressure pyroxene omphacite - (Na, Ca) $(\text{Al}, \text{Mg})\text{Si}_2\text{O}_6$ plus a complex garnet solid solution with considerable amounts of grossular component.

Eclogites, the "select or choice" type of rock, are clearly diagnostic of high pressures. Although they cover a wide PT-range from the low-temperature environments of subduction zones up to the much higher temperatures of a normal thermally equilibrated mantle, geothermometry based on element distribution between clinopyroxene and garnet (Ellis and Green, 1979) allows identification of those eclogites formed within subduction zones. Unfortunately, however, the simple two-phase eclogite mineralogy does not lend itself for geobarometry. The same handicap applies to other high-pressure minerals illustrated in Figure 9 such as lawsonite $\text{CaAl}_2\text{Si}_2\text{O}_7(\text{OH})_2\text{H}_2\text{O}$, the hydrous $\text{Fe}^{2+}\text{Fe}^{3+}$ -silicate deerite, or the $\text{Mn}^{2+}\text{Fe}^{3+}$ -garnet calderite, and even to the characteristic blueschist facies mineral glaucophane, the lower pressure limit of which as a function of composition (Fe^{2+}/Mg ; Fe^{3+}/Al) is practically unknown.

Thus, although blueschists and eclogites are certainly good for field evidence of high-pressure metamorphism, their usefulness in geobarometry and, thus, for reconstruction of

subduction depths is rather limited. Furthermore, eclogites as the stable subsolidus mantle facies of the basic melts produced within the mantle may actually never have left their place of origin. These rocks are, therefore, insignificant, and even deceptive rocks for the problem of crustal subduction discussed here. A crustal history of eclogites - either in the continental or the oceanic crust - can only be proven by additional chemical characteristics, as well as by textural features such as pillow structures, as preserved in the Swiss Alps (Fig. 10).

Figure 9: Selected stability curves of other minerals relevant for high-pressure metamorphism. For references see Schreyer (1988, Fig. 2). Thin solid lines are geotherms as in Figure 2.

Figure 8: Stability field of the high-P/low T assemblage phengite + talc (cf Fig. 6). Glaucophane curve as in Fig. 9, linear geotherms as in Fig. 2.

Finally, it must be pointed out that there are minerals and mineral assemblages that have such large stability ranges in terms of water pressure that they may appear in normal, low-pressure crustal metamorphism as well as in the "extreme" metamorphism of subduction zones. Examples are the common garnet end-members grossular, almandine ($\text{Fe}^{3+}\text{Al}_2\text{Si}_3\text{O}_{12}$), spessartine ($\text{Mn}^{3+}\text{Al}_2\text{Si}_3\text{O}_{12}$), and andradite ($\text{Ca}_3\text{Fe}^{3+}\text{Si}_3\text{O}_{12}$) that are stable at 1 atmosphere. However, it is of great interest to note that CaFe-bearing garnets, recently found as inclusions in natural diamonds, contain excess silicon, which can only be accommodated in the sixfold coordinated Al-site of the garnet structure (Moore and Gurney, 1985). Garnets of such chemistry are indeed indicative of high pressures.

Among the non-diagnostic minerals is, strangely enough, the well-known high-pressure Al-silicate polymorph kyanite, which may form as low as about 2 kbar, but persists up to at least 50 kbar. This is equally true for some of its assemblages, even in the diagnostic magnesian systems. One example is the assemblage kyanite + pure Mg-chlorite, which ranges from near 2 kbar to 30 kbar (Schreyer, 1988), or kyanite + phlogopite + SiO_2 covering pressures from near the Al_2SiO_5 triple point at about 3-4 kbar to near 30 kbar (see Fig. 8). For muscovite and phlogopite, both stable at 1 atm, no upper pressure limits have been found as yet, phlogopite persisting to at least 100 kbar. Other selected rock-forming minerals, non-diagnostic for pressure, are Fe-staurolite, chlorite, talc, zoisite and epidote, normal

Ca-bearing pumpellyite, K-feldspar (for H₂O-free conditions), and practically all the pyroxenes, whose transitions to garnet and perovskite structures lie at high mantle pressures greater than 100 kbar.

This list of non-diagnostic minerals and assemblages indicates that there is a good chance that rocks in deeply subducted units of continental crust may exhibit "absolutely normal" mineral assemblages of regional crustal metamorphism. It requires an experienced metamorphic petrologist to discover the characteristic fingerprints of deep crustal subduction among the multitude of metamorphic rock units of all ages the world over. Again, the pyrope-coesite rock presents a good example of what may yet be hidden in our various metamorphic series. Although generations of earth scientists have studied, mapped and described the Dora Maira Massif, and although some geologists had excellent samples of the spectacular pyrope crystals (Fig. 3) sitting on their desks, the geological and geodynamic significance of these rocks was not recognized until 1984, by Chopin.

Re-equilibration to Normal Crustal Pressures

The recognition by petrological methods of very deeply subducted continental crustal material is severely aggravated by the physical chemical principle of equilibration. Our petrological knowledge of deep subduction rests on the assumption that the rocks metamorphosed have attained chemical equilibrium at, for example, 30 kbar and 750°C, as demonstrated by their mineral assemblages now observable in the surface outcrops. Undoubtedly, however, these rocks also had a long way to return to shallower levels, where they were reincorporated into the "normal" continental crust. The fact that the high-pressure equilibrium mineralogy of such rocks can be preserved during tectonic uplift (obduction) is solely due to slow reaction kinetics that prevented re-equilibration to normal crustal pressures of less than 10 kbar.

The petrological features to be observed in the Dora Maira rocks present numerous examples of re-equilibration reactions, which are commonly summed up as retrograde metamorphism. Thus, the huge pyrope crystals (Fig. 3) are commonly found preserved only as pseudomorphs now consisting of chlorite + quartz + some newly grown kyanite (Schreyer, 1985). Chopin (1984) pointed out that the retrograde path taken by the Dora Maira rocks after the 30 kbar event remained entirely within the stability field of phengite + talc (see Fig. 8), because no formation of the assemblage phlogopite + kyanite + quartz was observed. This means that the Dora Maira rocks cannot have been uplifted isothermally, but were gradually cooled during their

tectonic transport back towards the normal crust. If the mechanism of obduction is indeed that proposed by Goffé and Chopin (1986), the thermal regime in the obduction zone (the "root zone" of classical alpine geology?) must be essentially the same as that of the previously active subduction zone. In this respect, the model of Cloos (1982) is of interest with its conveyor-belt system along more or less constant, linear geotherms with descending cold rocks cooling the rising hotter material and vice versa.

Nevertheless, even if decompression and cooling of deeply subducted crustal materials occur along pathways optimal for preservation, there is hardly any chance for the high-pressure mineral assemblages to survive when there is sufficient water present in the fluids penetrating the rock. This is so, because water has a twofold influence on retrograde metamorphism. First, it acts as a catalyst speeding up reactions and, second, it is a necessary chemical ingredient because most of the retrograde reactions are hydration reactions. In this context, many greenschists carrying, for example, chlorite + muscovite + biotite + quartz + albite + epidote, may at some earlier stage have exhibited high-pressure minerals that were wiped out by kinetically favourable retrograde reactions in the presence of excess H₂O fluid. Indeed, the directly adjoining country rocks of the Dora Maira pyrope locality have this very mineral assemblage, and it is only through careful microprobe studies that high-Si phengites (see Fig. 7) were recognized as relic fingerprints of an earlier high-pressure history of these rocks as well (Chopin, 1987).

Reaction kinetics are undoubtedly also favoured by higher temperatures. Thus, decompression of deeply subducted crustal rocks at still high, or even rising temperatures is most likely to be devastating to their mineralogy. If a pure pyrope crystal existing stably at 1000°C and 20 kbar were brought to a pressure of 9 kbar at the same temperature, its stable breakdown assemblage becomes enstatite + quartz + sapphirine. Based on purely empirical results in the requisite experimental system MgO-Al₂O₃-SiO₂, the rate of reaction at this temperature is considerable, so that the pyrope would have disappeared certainly within weeks or months. This is no time compared to orogenic evolution!

The chosen low-pressure conditions fall into the granulite facies of metamorphism, which is believed to prevail in much of the Earth's lower crust, and indeed the assemblage enstatite + quartz + sapphirine has been encountered in granulite terranes (e.g. Grew, 1980). Could this assemblage represent the breakdown product of a former pyrope, thereby indicating an earlier high-pressure metamorphic event? Undoubtedly it could, but evidence is hard to obtain. There can, in fact, only be petrographic textural evidence, for example through symplectitic intergrowths of enstatite + quartz + sapphirine outlining the shapes of the former garnets. Ideally, the relative proportions of the three minerals should add up to equal pyrope composition, provided the breakdown reaction was isochemical. As granulites are notorious for symplectites and corona textures (e.g. Herd et al., 1987), it is a promising new field in petrology to reconstruct former reaction relationships in these rocks and thus the pressure-temperature-time (PTt) paths of metamorphism.

Conclusions

There has been an increasing, if not overwhelming, body of petrologic evidence in recent years indicating that rocks of the continental crust including sediments may be subducted,

Figure 10: Pillow structure in an eclogite from the Alps. The external portions of each pillow (dark) consist of amphibolite as opposed to the main internal volume of eclogite. Lengths of pillows are about 40 cm. Pfulwe near Zermatt, Switzerland. Photo courtesy of M. Frey, Basel.

in the course of continent/continent collisions to great depths, normally occupied by material of the Earth's upper mantle. The recent discovery of diamonds in garnets of gneisses from a metamorphic complex in Kazakhstan (Sobolev and Shatsky, 1987) just underscores this situation. These new facts must change the classical geochemical view that chemical interaction between continental crust and mantle is a "one-way street" from mantle to crust. Since the rather unlikely geodynamic process, of returning to the surface a sediment formerly subducted to 100 km has become reality for the Dora Maira rocks, chances seem good that crustal materials go still deeper and become incorporated into the mantle and, particularly, into its melts produced there. Allègre and Turcotte (1985) envisage such "geodynamic mixing" processes to occur as deep as the 670 km boundary layer.

Considering the deep subduction of continental crust during mountain building and the making of the continental crust throughout Earth history, four different scenarios seem possible for the subsequent fate of continental crust after its subduction (see Fig. 11). The relative likelihood of these scenarios may also provide the answer as to why such an important terrestrial process as crustal subduction was not recognized earlier in the history of the geosciences.

Scenario I: A deeply subducted crustal unit metamorphosed to high grades (30 kbar, 800°C) along a very low geotherm ($\sim 7^\circ\text{C}/\text{km}$) is tapped at a relatively early stage in the orogenic development, i.e. while the thermal regime is still out of equilibrium because of the subduction process itself. Due to back-thrusting or other tectonic mechanisms in the collision zone, portions of this crustal unit come back up essentially along the same PTt path or, as shown in Figure 11.1., along a very slightly hotter geotherm. Thus, many or all of the prograde metamorphic reactions are now reversed, in the presence of excess water, while continued deformation and shearing allow not only the full access by water but also create new schistosity that may erase earlier fabrics. The latest metamorphic facies is one of low grade (green-schist facies) retaining no obvious mineralogical imprint of the subduction history.

Scenario II: The deeply subducted crustal unit is not tapped to return tectonically back into shallower levels as in I, but remains buried at considerable depth within the "mountain root" or the orogenic belt (Fig. 1). The thermal regime in the thickened crust is established at the end of the collision and subduction event (see Thompson and England, 1984, for more details), and the temperature rises in the mountain belt. Thus the PTt loop of curve II on Figure 11 shows increasing followed by decreasing pressures, which continue throughout the subsequent period of partial melting and crustal thinning due to extension tectonics.

By such processes of removing the "mountain root," the crustal unit would become part of the lower crust in a normal-thickness continental crust and enter granulite facies conditions. As most of the water would have left the rock by that time through dehydration reactions and partial melting, the new granulite-facies mineralogy will be attained by a series of anhydrous decompression reactions, causing volume increase such as in the example of pyrope breakdown cited above. Continued regional uplift of the granulite facies terrain accompanied by erosion would then follow classical concepts of basement geology.

In **Scenario III**, the process of subduction of continental crust to about 100 km (the Dora Maira depth) is not halted at this stage but continues to even greater depths in the upper mantle (Curve III on Fig. 11). Due to the intersection of the low geotherm of subduction and the curves for the beginning of melting in crustal rocks (assuming both hydrous and anhydrous conditions in Fig. 11), partial melting will commence within the subducted crustal units at depths of between 80 and 200 km, depending on water activity. Primary melt compositions would generally not be granitic

but poorer in silica, i.e. more syenitic (Huang and Wyllie, 1975). Such melts would probably react with mantle peridotites of the overlying lithospheric plate and cause hybridization, as already envisaged for the subduction of oceanic plates by Sekine and Wyllie (1985). At any rate, metamorphic subsolidus petrology will end in this scenario, and the challenge would be to igneous petrologists and geochemists to detect fingerprints of subducted continental crust in the magmatism of collision zones.

Scenario IV is largely identical to scenario I but relies on the shortage or absence of water during the retrograde PTt-path of the subducted crustal unit undergoing obduction. In the relatively dry environment of the Dora Maira pyrope coesite rock most pyropes are still unaltered (Fig. 3), the coesite to quartz transition has not gone to completion (Fig. 4), and retrograde reaction of jadeite + quartz to form albite has either left jadeite relics (Schreyer, 1985) or produced characteristic and diagnostic textures of the newly formed albite. It seems that the Dora Maira occurrence offers a particularly favourable case for the preservation of deeply subducted rocks of the continental crust. Undoubtedly, the multitude and large size of the pyrope crystals made this rock particularly resistant to retrograde changes. Garnet-free, phyllosilicate-rich, high-pressure schistose rocks are certainly much more sensitive and may lose their former identity completely, as exemplified by the Dora Maira country rocks.

It seems clear that scenario IV is the least likely of the four. But this is not a reason for neglecting these features and the far-reaching and overall geological implications. On the contrary, it should be a future task for petrologists to search for relic minerals or textures in metamorphic rocks indicative of the previous history of high-pressure crystallization.

The wide distribution of generally small occurrences of eclogites, clearly a very resistant rock type, in basement rocks of many regions and ages may indicate that deep subduction of the crust was actually a common, if as yet unrecognized, process, because the more acidic surrounding country rocks of vast extension bear no signs of high-pressure minerals or assemblages. In the Caledonides of western Norway, it became apparent only a few years ago that relic high-pressure assemblages in the acidic gneisses enclosing eclogites "are not uncommon" (Cuthbert et al.,

Figure 11: PT diagram illustrating four potential PTt paths (I-IV) of crustal rocks during subduction and subsequent return transport to the Earth's surface (I, II, and IV only). See text for explanation. Wet and dry melting curves apply to alkali granites and were extrapolated from Huang and Wyllie (1975). Linear geotherms as in Figure 2.

1983). Even in the complete absence of such relics, the eclogite occurrences are of significance, provided they have an unquestionable previous crustal history like that of Figure 10. The long and controversial discussion on the protoliths of the western Norwegian eclogites - crustal rocks versus mantle fragments (Griffin et al., 1985) - is a prime example.

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References

- Abraham, K. and Schreyer, W., 1976. A talc-phengite assemblage in piemontite schist from Brezovica, Serbia, Yugoslavia. *Journal of Petrology*, v. 17, no. 4, p. 421-439.
- Allègre, C.J. and Turcotte, D.L., 1985. Geodynamic mixing in the mesosphere boundary layer and the origin of oceanic islands. *Geophysical Research Letters*, v. 12, no. 4, p. 207-210.
- Chopin, C., 1981. Talc-phengite: a widespread assemblage in high-grade pelitic blueschists of the Western Alps. *Journal of Petrology*, v. 22, no. 4, p. 628-650.
- Chopin, C., 1984. Coesite and pure pyrope in high-grade blueschists of the Western Alps: a first record and some consequences. *Contributions to Mineralogy and Petrology*, v. 86, no. 2, p. 107-118.
- Chopin, C., 1987. Very-high-pressure metamorphism in the western Alps: implications for subduction of continental crust. *Philosophical Transactions of the Royal Society of London, Series A*, v. 321, no. 1557, p. 183-197.
- Chopin, C. and Goffé, B., 1981. High-pressure synthesis and properties of magnesiochloritoid, $MgAl_2(Si_2O_5)(OH)_4$. *Contributions to Mineralogy and Petrology*, v. 76, no. 3, p. 260-264.
- Chopin, C. and Schreyer, W., 1983. Magnesiochloritoid and magnesiochloritoid: two index minerals of pelitic blueschists and their preliminary phase relations in the model system $MgO-Al_2O_3-SiO_2-H_2O$. *American Journal of Science*, v. 283-A, p. 72-96.
- Cloos, M., 1982. Flow melanges: numerical modeling and geological constraints on their origin in the Franciscan subduction complex, California. *Geological Society of America Bulletin*, v. 93, no. 4, p. 330-345.
- Coes, L., Jr., 1955. High-pressure minerals. *American Ceramic Society Journal*, v. 38, no. 8, p. 298.
- Cuthbert, S.J., Harvey, M.A. and Carswell, D.A., 1983. A tectonic model for the metamorphic evolution of the basal Gneiss Complex, Western South Norway. *Journal of Metamorphic Geology*, v. 1, no. 1, p. 63-90.
- Ellis, D.J. and Green, D.H., 1979. An experimental study of the effect of Ca upon garnet-clinopyroxene Fe-Mg exchange equilibria. *Contributions to Mineralogy and Petrology*, v. 71, no. 1, p. 13-22.
- Ernst, W.G., 1963. Significance of phengitic micas from low-grade schists. *American Mineralogist*, v. 48, p. 1357-1373.
- Ernst, W.G., 1973. Blueschist metamorphism and P-T regimes in active subduction zones. *Tectonophysics*, v. 17, no. 3, p. 255-272.
- Goffé, B. and Chopin, C., 1986. High-pressure metamorphism in the Western Alps: zonation of metapelites, chronology and consequences. *Schweizerische mineralogische und petrographische Mitteilungen*, v. 66, p. 41-52.
- Grew, E.S., 1980. Sapphirine + quartz association from Archean rocks in Enderby Land, Antarctica. *American Mineralogist*, v. 65, no. 9-10, p. 821-836.
- Griffin, W.L., Austrheim, H., Brastad, K., Bryhni, I., Krill, A.G., Krogh, E.J., Mork, M.B.E., Qvale, H. and Torudbakken, B., 1985. High-pressure metamorphism in the Scandinavian Caledonides. In: Gee, D.G. and Sturt, B.A. (eds.) *The Caledonide Orogen, Scandinavia and Related Areas*. Vol. II. John Wiley and Sons, p. 783-801.
- Herd, R.K., Ackermann, D., Thomas, A. and Windley, B.F., 1987. Oxygen fugacity variations and mineral reactions in sapphirine-bearing paragneisses, E. Grenville province, Canada. *Mineralogical Magazine*, v. 51, no. 360, p. 203-206.
- Huang, W.L. and Wyllie, P.J., 1973. Melting relations of muscovite-granite to 35 kbar as a model for fusion of metamorphosed subducted oceanic sediments. *Contributions to Mineralogy and Petrology*, v. 42, no. 1, p. 1-14.
- Jacques, A.L., 1987. Diamond-bearing lamproites from Western Australia: multiple tapping of old enriched lithosphere beneath proterozoic mobile belts. *Terra Cognita*, no. 4, v. 7, p. 616.
- Massonne, H.-J. and Schreyer, W., 1986. High-pressure syntheses and X-ray properties of white micas in the system $K_2O-MgO-Al_2O_3-SiO_2-H_2O$. *Neues Jahrbuch für Mineralogie Abhandlungen*, v. 153, no. 2, p. 177-215.
- Massonne, H.-J. and Schreyer, W., 1987. Phengite geobarometry based on the limiting assemblage with K-feldspar, phlogopite, and quartz. *Contributions to Mineralogy and Petrology*, v. 96, no. 2, p. 212-224.
- Mckie, D., 1959. Yoderite, a new hydrous magnesium iron aluminosilicate from Mautia Hill, Tanganyika. *Mineralogical Magazine*, v. 32, no. 247, p. 282-307.
- Moore, R.O. and Gurney, J.J., 1985. Pyroxene solid solution in garnets included in diamond. *Nature*, v. 318, no. 6046, p. 553-555.
- Sibley, D.F. and Wilband, J.T., 1977. Chemical balance of the Earth's crust. *Geochimica et Cosmochimica Acta*, v. 41, no. 4, p. 545-554.
- Schreyer, W., 1968. A reconnaissance study of the system $MgO-Al_2O_3-SiO_2-H_2O$ at pressures between 10 and 25 kb. *Yearbook of the Carnegie Institution of Washington*, v. 66, p. 380-392.
- Schreyer, W., 1977. Whiteschists: their compositions and pressure-temperature regimes based on experimental, field and petrographic evidence. *Tectonophysics*, v. 43, no. 1-2, p. 127-144.
- Schreyer, W., 1985. Metamorphism of crustal rocks at mantle depths: High-pressure minerals and mineral assemblages in metapelites. *Fortschritte der Mineralogie*, v. 63, no. 2, p. 227-261.
- Schreyer, W., 1988. Experimental studies on metamorphism of crustal rocks under mantle pressures. *Mineralogical Magazine*, v. 52, no. 364, p. 1-26.
- Schreyer, W. and Baller, T., 1977. Talc-muscovite: synthesis of a new high-pressure phyllosilicate assemblage. *Neues Jahrbuch für Mineralogie Monatshefte*, no. 9, p. 421-425.
- Schreyer, W. and Yoder, H.S., Jr., 1968. Yoderite: Synthesis, stability, and interpretation of its natural occurrence. *Yearbook of the Carnegie Institution of Washington*, v. 66, p. 376-380.
- Schreyer, W., Maresch, W.V., Medenbach, O. and Baller, T., 1986. Calcium-free pumpellyite, a new synthetic hydrous Mg-Al-silicate formed at high pressures. *Nature*, v. 321, no. 6069, p. 510-511.
- Sekine, T. and Wyllie, P.J., 1985. Hybridization of magmas above subducted oceanic crust. *Geologicky Sbornik*, v. 36, no. 3, p. 259-268.
- Sobolev, N.V. and Shatskii, V.S., 1987. Inclusions of carbon minerals in garnets of metasomatic rocks. *Geologiya i Geofizika*, v. 28, no. 7, p. 77-80.
- Thompson, A.B. and England, P.C., 1984. Pressure-temperature-time paths of regional metamorphism II. Their inference and interpretation using mineral assemblages in metamorphic rocks. *Journal of Petrology*, v. 25, no. 4, p. 929-955.

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