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MATHEMATICAL, PHYSICAL & ENGINEERING SCIENCES

Very-High-Pressure Metamorphism in the Western Alps: Implications for Subduction of Continental Crust [and Discussion]

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Very-high-pressure metamorphism in the western Alps: implications for subduction of continental crust

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The widespread occurrence of coesite and pure pyrope, the presence of several new mineral assemblages and new rock-forming minerals in common lithologies, but also crystal–chemical features such as the existence of silicate–phosphate solid solutions or the repeated occurrence among the new minerals of a new dense structure with face-sharing octahedra, are evidence that unusual pressures have been attained in part of the Dora Maira massif, western Alps, during Alpine regional metamorphism. Mineral assemblages suggest minimum pressures well in excess of 25 kbar (1 bar = 10^5 Pa) which, according to preliminary experimental data, may have reached 35 kbar in this at least 5×10 km^2 continental terrain clearly of supra-crustal origin. Obviously even continental material may be buried to depths of the order 100 km and then uplifted to the surface. Whereas the burial in a low temperature régime is readily explained by subduction of the continental lithosphere (of a peninsula?) 110 Ma ago or earlier, the uplift is much more problematic because it is petrologically constrained to proceed without significant temperature increase. The progressive migration 'in-plate' of intra-continental thrusts once subduction was blocked, accounts for the stepwise decrease in age and grade of high-pressure metamorphism toward the external part of the chain; by the repeated underthrusting of cold material it might also have prevented a temperature increase in the most internal, early subducted zones.

Introduction

The recent discovery of the high-pressure silica polymorph coesite as a product of regional metamorphism in a continental unit of the internal Western Alps (Chopin 1984) raises a number of questions. They bear on the significance of coesite (i.e. on its stable or metastable growth), on the geologic setting of the occurrence (i.e. *in situ* against foreign character of the coesite rock in the surrounding series) and, once these questions have been answered, on the behaviour of continental crust in collision zones. Concerning the first point, a mechanical study by Gillet *et al.* (1984) showed that the coesite relics preserved in pyrope garnets now in a quartz matrix were trapped as coesite and not as quartz during the garnet growth, implying that the whole quartz matrix had formerly been coesite. Coesite relics were selectively preserved within garnet because of the low compressibility and high mechanical strength of the latter mineral. Furthermore Chopin (1984) favoured the idea that coesite had grown stably, both on textural grounds and because the very large deviatoric stresses necessary to shift noticeably the quartz–coesite transition, from its position determined under hydrostatic conditions, are very unlikely to be met in a sample naturally deformed at temperatures well in excess of 600 °C (cf. Green 1974). Accordingly, the coesite rock would have crystallized at minimum pressures of 28 kbar,† thus at depths near 100 km.

† 1 bar = 10^5 Pa.

The purpose of the present paper is to document the regional setting of the coesite rock in the light of recent field and petrographic studies in the Dora Maira massif. It will be shown that coesite and other, even less debatable evidences for very high pressures are, in fact, regionally widespread. Tectonic consequences are then explored at various scales; consequences bearing on the generation of acidic to intermediate magmas at depth in the crust and on mantle geochemistry are out of the scope of this paper.

Geologic setting

The Dora Maira massif is a narrow belt of crystalline Palaeozoic terrains extending over about 70 km from Val Susa in the north to Val Maira in the south, and limited to the east by the Po basin. Together with Monte Rosa and Gran Paradiso they constitute the internal crystalline massifs of the Penninic zone of the western Alps and are thought to represent the eastern edge of the former European plate. They share a common structural position since they appear through windows cut in eclogite-bearing ophiolitic units of the Schistes lustrés nappe (figure 1, inset) such as the Zermatt zone or the Rocciavre and Monviso massifs. However, whereas Monte Rosa and Gran Paradiso show striking lithological similarities, Dora Maira

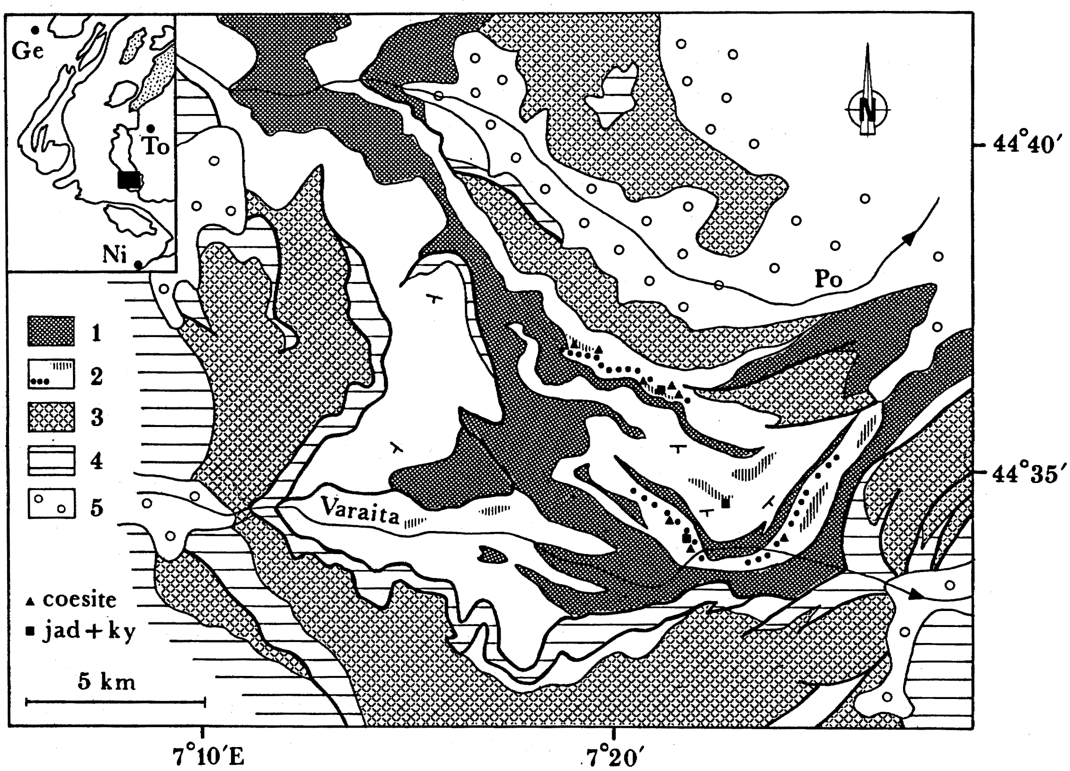

Figure 1. Geological map of the southern Dora Maira massif (mainly after Vialon 1966). (1) Palaeozoic augen gneisses and metagranites; (2) Palaeozoic fine-grained gneisses containing intercalations of metapelite, eclogite and marble (vertical ruling) and of pyrope–quartzite (dots). (1) and (2) constitute the 'polymetamorphic unit'. (3) Overlying and underlying Late Palaeozoic (?) schists and metarhyolites; (4) Mesozoic undifferentiated: Triassic cover (?) series and Schistes lustrés; (5) alluvium. The main occurrences of coesite relics and of the jadeite+kyanite assemblage are shown. The inset locates the area in the western Alps, showing the external crystalline massifs, the Penninic front, the internal crystalline massifs, the Austro–Alpine Sesia zone with its Dent Blanche outlier (both dotted). Symbols: Ge, Geneva; Ni, Nice; To, Torino.

differs by a greater lithological variety and by the absence of more internal units to the east. In Dora Maira, an allegedly Carboniferous sequence is exposed in the eastern part of the massif; it is the structurally lowest unit since the general vergence is toward the east. It is overlain by a polymetamorphic gneiss unit containing abundant metagranites, which is in turn overlain by an allegedly Permocarboniferous sequence characterized by abundant acidic igneous activity (Vialon 1966; Michard 1967) and separated from the overlying ophiolites by Mesozoic strata of Triassic and younger age. The relations between the different Palaeozoic units have been a matter of debate for almost a century, the controversy being between cover to basement (Franchi 1906; 1929; Vialon 1966) and tectonic relations (Argand 1911; Michard 1965).

The first coesite rocks described by the author (Chopin 1984) were found within the polymetamorphic unit in an outcrop of the pyrope–phengite–quartzite layer mentioned by Vialon (1966). The layer is a few metres thick and, as mapped by Vialon, can indeed be followed over kilometres in southern part of the massif (figure 1). The surrounding series consists of monotonous, often layered gneisses, which are rather fine-grained compared with the augen gneisses mentioned below. The gneisses directly enclose the pyrope–quartzite but consistently contain in its immediate vicinity layers of marble and metapelitic kyanite–garnet–schists together with eclogite boudins. Abundant augen gneisses grading locally into metagranites with beautifully preserved textures may represent Palaeozoic granites intrusive into the former gneiss series (Vialon 1966). The interlayering observed in several places of the pyrope–quartzite with the same jadeite–kyanite–quartzite, the lithological suite pyrope–quartzite + kyanite–garnet–metapelites + marble + eclogite and its regional extension constitute a first group of purely geological arguments for considering the coesite rock not as a tectonic block in the surrounding series nor as an igneous product as suggested by Vialon, but rather as part of a metasedimentary series.

Towards a coherent coesite-bearing terrane

Since the first finding, many other occurrences of coesite relics and of pseudomorphs evidently after coesite have been found in the southern part of the Dora Maira massif, both in Val Po and in Val Varaita and consistently within the polymetamorphic unit. Many of them were found throughout the pyrope–quartzite layer but, most significantly, several others were in the country-rocks. A typical pseudomorph after coesite (Smyth 1977; Chopin 1984; Smith 1984) has been observed in omphacite within eclogite in Val Po and, in the same area, relics of coesite are commonly preserved within garnet and kyanite porphyroblasts in metapelites (figure 1). Thus, among the lithologies encountered the only 'barren' ones, with respect to coesite, are so far marbles and the gneisses that make up most of the unit. Nevertheless the presence of coesite in the country-rock and its widespread occurrence corroborate the geological evidence and leave little doubt that physical conditions sufficient to enable coesite growth in various materials have been regionally attained in the polymetamorphic unit, at least in its southern extension.

A fascinating point to speculate upon is the full extent of this coesite zone, because the polymetamorphic unit extends from north to south over the whole massif. The clearest marker, the pyrope–quartzite layer, has not been reported from other parts of this unit. Nevertheless the associated lithologies marble–metapelite–eclogite can be traced over the whole unit, westward of the coesite zone in Val Varaita and northward to the Val Germanasca. There, metapelites are locally very coarse-grained and reminiscent of the metapelites in the coesite zone, with centimetre-large garnets and micaceous prismatic pseudomorphs suggesting former kyanite crystals. Although tentatively thought to be after amphibole (Vialon 1966), the pseudomorphs might have been kyanite as well, the more so because a fine-grained chloritoid of probably late formation (after garnet+kyanite?) is abundant in the matrix. However, definite evidence for coesite is lacking, possibly owing to the severe late overprint under greenschist facies conditions which obliterated most of the former assemblages in the area, making it difficult even to find well-preserved eclogite. In the present state of knowledge the coesite zone extends over an area of about $5 \times 10\ km^2$, but the possibility remains that it might have had a considerably greater extension over the massif. There is, so far, no evidence that coesite was ever present in the neighbouring late Palaeozoic units.

Further evidence for high pressure

Even if it points to unusual conditions, the presence of coesite at a regional scale is not of straightforward interpretation because of the effect of deviatoric stress on phase transitions, as mentioned in the introduction. The purpose of this section is therefore to gather independent mineralogical and petrological evidence that could characterize the conditions experienced by the coesite-bearing unit.

(a) *New rock-forming minerals*

Additional evidence that unusual metamorphic conditions have been met in the coesite-bearing terrane is given by the presence of several new minerals in rocks of common chemical composition.

Ellenbergerite $(Mg, Ti, Zr, \square)_2Mg_6Al_6Si_8O_{28}(OH)_{10}$ occurs as inclusions within pyrope megacrysts in the pyrope–coesite–quartzite layer (Chopin *et al.* 1986). An analysis of its phase relations shows that this highly hydrated mineral is a high-pressure and relatively low-temperature phase with an upper temperature stability limit below 800 °C and a minimum formation pressure of at least 20 kbar (Chopin 1986), in accordance with the unusual, dense structure characterised by face-sharing octahedra (figure 2). Preliminary experimental work led to the synthesis of ellenbergerite and to the puzzling result that titanian, Zr-free ellenbergerite cannot be stable below 35 kbar water pressure, unless it contains considerable amounts of phosphorus substituting for silicon (Schreyer 1985). Most natural crystals are indeed zoned with P-rich cores but, commonly, P-free rims, which (if grown stably) would thus imply much higher pressures than previously thought.

Interestingly, the replacement of silicon by phosphorus (mainly through the $SiAl \rightleftharpoons PMg$ substitution) may be complete in ellenbergerite, which leads to a new Mg-phosphate, with ellenbergerite structure. This is a quite remarkable feature since, under normal crustal conditions, no such bridge exists between the silicate and phosphate realms, even when they share common structures.

In addition, two new polymorphs of already known compounds occur in the pyrope megacrysts: a monoclinic polymorph (or polytype?) of wagnerite, $Mg_2PO_4(F, OH)$, and a hexagonal polymorph of a Mg-bearing dumortierite, ideally $(Mg, Ti, \square)_2Al_6Al_6B_2Si_6O_{34}(OH)_2$. The latter is particularly interesting because its structure is exactly that of ellenbergerite, in which two BO_3 triangles replace two $SiO_3(OH)$ tetrahedra on the threefold axes (Moore & Araki 1978; Chopin *et al.* 1986). It is the third example of an

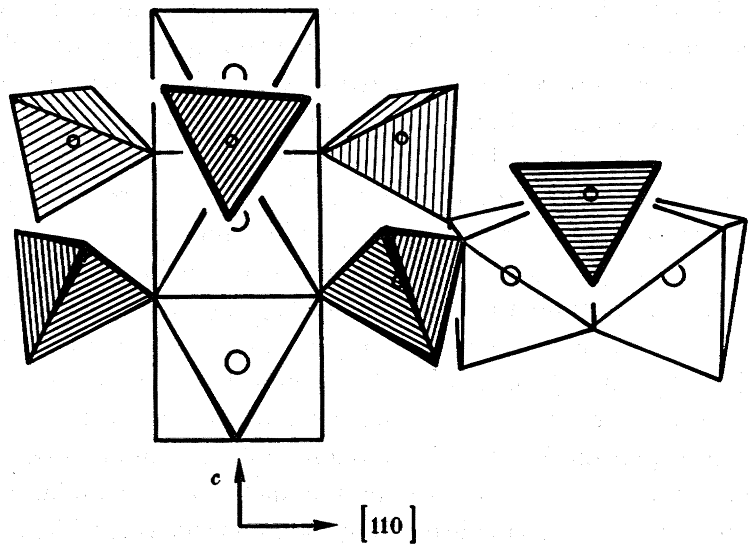

FIGURE 2. Detail of the ellenbergerite structure showing the presence of face-sharing octahedra, both in infinite chains along with six-fold axes (left of the picture) and as octahedral pairs (right) that share edges to form infinite double chains (not represented). In the dumortierite polymorph, the SiO_3 (OH) tetrahedron represented on the right is replaced by a planar BO_3 group.

ellenbergerite structure occurring there; this one is particularly dense, with a packing efficiency near 15.6 Å^3† per oxygen, against 16.0 in pyrope and 16.8 in ellenbergerite.

Magnesiostaurolite is another new silicate found as inclusions in the pyrope megacrysts. This staurolite contains 80–95 mol % of the Mg-end-member, i.e. nearly twice as much as the most magnesian staurolites so far reported (Ward 1984). The end-member is in fact long known from experiment as a high-pressure phase encountered above 15 kbar water pressure in the $MgO–Al_2O_3–SiO_2–H_2O$ system (Schreyer 1968; Schreyer & Seifert 1969).

One might think that this impressive wealth of new minerals in the pyrope quartzite results from the unusual bulk-rock composition, but this is not so, for two reasons. The pyrope–phengite–quartzite is nothing other than a magnesian metapelite, like those known to occur in shelf series sediments with evaporitic tendencies (Moine *et al.* 1981), or like those known in the nearby Gran Paradiso (Chopin 1981) and Monte Rosa (Chopin & Monié 1984) massifs, where these new minerals do not occur in spite of metamorphic pressures having approached 15 kbar. Furthermore, new minerals in the Dora Maira massif do not only occur in the pyrope–quartzite but also in the enclosing series. Indeed a new calcium phosphate, ideally $Ca_2Al(PO_4)_2(OH)$, is found as an accessory mineral both in the pyrope–quartzite and in metapelites with a more normal Fe:Mg ratio. There apatite is not as common as in normal pelitic metamorphic rocks, this new mineral occurring instead but breaking down during the late metamorphic evolution to form apatite, among other products.

(*b*) *High-pressure assemblages*

Regardless of the presence of coesite or new minerals, a few uncommon assemblages of more common minerals point to very high pressures in this terrain.

Kyanite–eclogite and kyanite–jadeite–quartzite occur throughout the area. The typical

† 1 Å = 10^{-10} m = 10^{-1} nm.

assemblage is omphacite or jadeite–garnet–kyanite–quartz–phengite–rutile, which buffers the paragonite content of white mica to a maximum value and thus provides a divariant relation between pressure, temperature and water activity (cf. Holland 1979*a*). Molar paragonite contents in phengite are below 4%, jadeite contents in pyroxene are very close to 50% in eclogite and range between 80 and 92% in the jadeite–kyanite–quartzite. Although the derivation therefrom of absolute values of the intensive parameters is heavily dependent on activity models for the pyroxene and white mica solid solutions, such mineral compositions imply pressure values clearly in excess of any estimate made so far in the metamorphic realm. On the basis of the experimental work of Holland (1979*b*), the order of magnitude of the minimum pressure estimate is 25 kbar (cf. Chopin 1984). In this respect the most critical assemblage is jadeite–kyanite which, to the author's knowledge, has not been reported so far. This assemblage or relics of it occur in quartzite, not only closely associated with the pyrope–quartzite in Val Po and Val Varaita, but also in the surrounding series (B. Lombardo, personal communication 1985). Furthermore the metapelites contain, beside phengite–almandine–kyanite, pseudomorphs made of albite and minor potassic white mica that are most likely after rather pure jadeite. Obviously, the critical pair jadeite–kyanite must have been regionally abundant.

The most definite indication is given by the presence, in the matrix of the pyrope–phengite–quartzite, of the assemblage pyrope–talc–SiO_2. The coexistence within the talc stability field of pure pyrope with any of the polymorphs of silica is restricted to a very narrow wedge of the P_{H_2O}–T field, for which temperatures near 800 °C and minimum pressures of about 28 kbar were proposed in Chopin (1984) on the basis of available experimental data. This value was confirmed by further experimental work (Chopin 1985). Most importantly, reduced water activity shifts the relevant field towards lower temperatures but not towards lower pressures. Thus pyrope may stably coexist with talc and SiO_2 exclusively within the coesite field as determined under hydrostatic conditions. This is the most convincing evidence that the conditions of coesite stability have effectively been reached by this unit.

(*c*) *Summary*

The widespread occurrence of coesite and pure pyrope, the presence of several new mineral assemblages and rock-forming minerals in common lithologies, but also more subtle crystal–chemical details such as the existence of silicate–phosphate solid solutions or the repeated occurrence among the new minerals of a dense, otherwise unknown structure, constitute compelling evidence that the metamorphic terrain considered must have endured exceptionally high pressures for continental crust. The mineral assemblages encountered suggest minimum pressures that are in excess of 25 kbar and may reach 35 kbar, according to preliminary experimental work. We therefore cannot escape the issue that a continental terrain may be buried to depths of the order of 100 km and then uplifted to reach the surface. Since this is a new result and often considered to be at variance with concepts prevailing in geodynamics, an attempt must be made at reconstructing the successive stages of this evolution. In the following, before envisaging consequences, we will shortly consider the few constraints imposed by the metamorphic records and geochronological data on the conditions and timing of burial and uplift.

Constraints on the metamorphic evolution

(*a*) *Burial and prograde path*

It had been hoped that the huge pyrope crystals crowded with inclusions would yield some information about the prograde history of the terrain but they did not, because pyrope growth only began once quite high pressure had already been reached (Chopin 1986). In fact, more information is available from the metapelites of the enclosing series in which normal Fe:Mg ratios caused a more continuous mineral evolution covering most of the prograde history. Indeed the core of centimetre-large almandine garnet porphyroblasts contains tiny inclusions of staurolite, magnesian chloritoid, sometimes paragonite or quartz, rarely kyanite, whereas the rim is usually free of inclusions with the possible exception of a few large ones showing the typical texture of quartz pseudomorph after coesite, sometimes with coesite relics. Chloritoid, staurolite and paragonite are significantly absent in the matrix in which the main assemblage was almandine–kyanite–jadeite (now pseudomorphed)–phengite–coesite (now as quartz). This coexistence of almandine with kyanite and of jadeite with kyanite implies that the upper temperature stability limits of chloritoid and staurolite in the presence of quartz and the upper pressure stability limit of paragonite, respectively, have been overstepped during the metamorphic evolution which led to coesite formation. This is, of course, as theoretically expected (figure 3) but finding milestones along such an unusual path lends further credence to the estimates made above. Interestingly, the presence of paragonite and of magnesian chloritoid (with up to 50 mol % of the Mg-end-member) among the oldest records we have of the metamorphic evolution is indicative of moderate temperatures (Chatterjee 1972) and high pressures (Chopin 1983; Chopin & Schreyer 1983), respectively; that is, of a rather low thermal gradient during the early metamorphic history. This is also taken as evidence that the observed features are the product of a single, Alpine metamorphic cycle.

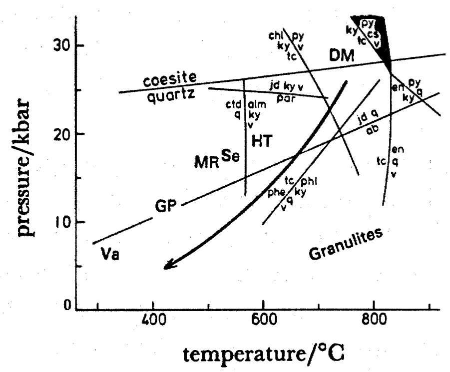

Figure 3. Some petrological constraints on the metamorphic evolution represented in a (P_{H_2O}, T) diagram. The large symbols represent P–T estimates obtained on high-pressure assemblages from several areas of the Alpine chain; Va, Vanoise (Goffé 1982); GP, Gran Paradiso (Chopin 1984); MR, Monte Rosa (Chopin & Monié 1984); Se, Sesia zone (Compagnoni 1977); DM, Dora Maira (this study), all in the western Alps; HT, Hohe Tauern, eastern Alps (Holland 1979*a*). The trace of these estimates forms a 'path' compatible with the loosely constrained prograde path of the coesite-bearing unit, although the latter may lie at lower temperatures. The black solid area represents the conditions for which pyrope may coexist with talc and either quartz or coesite. The figure is reproduced from Chopin (1984), to which the reader is referred for additional data sources. Recent experimental work (Chopin 1985) shows that the pyrope–talc–SiO_2 field must be slightly shifted toward higher pressures, so that it lies entirely within the coesite field. Abbreviations; ab, albite; alm, almandine; chl, chlorite; ctd, chloritoid; en, enstatite; jd, jadeite; ky, kyanite; par, paragonite; phe, phengite; phl, phlogopite; py, pyrope; q, quartz; tc, talc; v, hydrous fluid.

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(*b*) *The uplift path*

Information exists on the 'maximum' metamorphic conditions, that is fragmentary in that maximum pressure and maximum temperature need not coincide. The highest-temperature assemblage found is talc–kyanite–pyrope–SiO_2 which implies temperatures in the range 700–800 °C, depending on water activity, and probably close to 700 °C because of the lack of evidence for partial melting in the country-rock. As already mentioned, this assemblage implies minimum pressures of about 28 kbar but ellenbergerite formation, which occurred at lower temperatures (Chopin 1986), may have occurred at somewhat higher pressure if the first experimental results are to be trusted.

^{40}Ar–^{39}Ar dating on phengite from the former assemblage and from unretrograded kyanite–eclogite yields plateau ages in the range 100–110 Ma (P. Monié, unpublished data). This minimum age for the high-pressure metamorphism is very similar to those obtained from the highest-pressure assemblages in the Monte Rosa massif (Chopin & Monié 1984) and Sesia zone (Oberhänsli *et al.* 1985).

The later evolution under decreasing pressure conditions is rather well constrained by the stability, during most of it, of the talc–phengite pair. As discussed by Chopin (1984), it is very unlikely, for kinetic reasons, that the higher-temperature and lower-pressure stability limit of this assemblage could have been overstepped without producing phlogopite + kyanite + quartz, which is not observed as a paragenesis. According to the experimental work on this reaction by H.-J. Massonne (personal communication 1983, see figure 3), this constrains most of the uplift to have proceeded under relatively high-pressure, low-temperature conditions, and leaves virtually no room for a temperature increase during uplift. A petrographic confirmation of this is the retrograde formation of the high-pressure mineral glaucophane from pyrope in the pyrope–quartzite (Schreyer 1985, fig. 17) and, together with minor paragonite, from garnet + kyanite + jadeite in the jadeite–kyanite–quartzite. Significant in term of the physical processes involved (England, this symposium) is the near absence of temperature increase during uplift implied by these petrological constraints and by other independent factors, such as the preservation of coesite in garnet (Gillet *et al.* 1984) and the likely absence of partial melting in the terrain considered.

Further pressure release still proceeded under decreasing temperature, leading to widespread greenschist facies assemblages. These are particularly well developed in the gneiss series with the assemblage muscovite–biotite–K-feldspar–albite–epidote–Ca-rich garnet–sphene–quartz, in which only high-Si phengite cores in white mica point to former high-pressure conditions. ^{40}Ar–^{39}Ar dating of the late white micas yields consistently 38–40 Ma ages (P. Monié, unpublished data), as everywhere in the western Alps (Bocquet *et al.* 1974; Delaloye & Desmons 1976; Chopin & Maluski 1980; Chopin & Monié 1984) except the internal Sesia zone (Hunziker 1974). Clearly most of the uplift was achieved before the end of the Eocene, i.e. within about 70 Ma.

Consequences

(*a*) *Regionally*

A first consequence of this study is of regional concern, answering a long-standing question. The unique metamorphic grade attained by the coesite-bearing unit rules out any relation other than tectonic between the polymetamorphic unit and the neighbouring ones. The metamorphic gap is considerable, even apparent in the field when comparing the mineralogy developed in the metapelites during the main metamorphic stage. Metapelites typically contain chloritoid (±garnet±glaucophane) in the late Palaeozoic units (Vialon 1966; Michard 1967), whereas they bear the alternative assemblage kyanite+almandine±jadeite in the polymetamorphic unit. Tectonic relations could have been suspected, especially in the southern part of the massif where a narrow band of Mesozoic Schistes lustrés separates the coesite-bearing unit from the overlying one (figure 1). However, given the common interfingering of units on the western side of the massif, it had never been clear whether this allochthony was real or just a late feature, and the massif had always been considered as a whole from a metamorphic point of view. It is clear now that the allochthony must be profound between the polymetamorphic unit and the western, late Palaeozoic unit which, in fact, bears much affinity with the Briançonnais basement (Lefèvre 1982). The same must hold true between the polymetamorphic unit and the underlying Carboniferous where chloritoid is widespread and characteristic in metapelite (Vialon 1966).

(*b*) *Subducting continental crust*

More generally speaking, the main confirmation brought by this study is that fragments of continental crust may be buried to depths approaching or possibly exceeding 100 km, in spite of their low density. Although at variance with common prejudice, this is perhaps not as unexpected at it seems. Underthrusting of continental material over distances of 100–500 km has been postulated in several regions, e.g. below Southern Island, New Zealand (Allis 1981), below Timor (Hamilton 1979) and below Himalaya (Seeber & Armbruster 1984), but in no case does the depth attained exceed 60 km, because of the gentle dip of the subducted plate in the two latter cases. However, Roecker (1982) has presented geophysical evidence suggesting that continental rather than only oceanic lithosphere has been subducted to depths of *at least* 150 km below the Pamir–Hindu Kush region.

The problem of subduction of continental crust has been specifically addressed from a theoretical point of view by Molnar & Gray (1979), who tried to evaluate the two gravitational effects that contribute to subduction of continental lithosphere, the negative buoyancy of the relatively cold mantle part of the continental lithosphere and the pull of a downgoing slab of oceanic lithosphere on continental lithosphere trailing behind it. They concluded that, owing to our uncertainty in the processes occurring and depending on the parameter values, estimates between 10 and 300 km could be made for the length of continent subducted. Because seismic evidence suggests that the lower part of the sinking slab does not exert a pull on upper parts, they favoured the lower values of this range but stated that, in the case of peninsulas or microcontinents, subduction of intact continental lithosphere could be complete. A greater length of subducted crust is expected if the crust is thinner than average. From a physical point of view it thus seems possible to bring continental material to great depth. The geological evidence suggests that this process may have been favoured in the Alpine realm by (i) a complex pattern of oceanic basins, such as Piemont and Valais troughs, and continental fragments (Lemoine 1985), and (ii) the presence of thinned continental crust that probably formed during the extensional régime that prevailed from Trias to, approximately, Lower Cretaceous (see Lemoine 1985, for the western Alps, and Le Pichon & Sibuet 1981, for a general mechanism).

(c) *A problematic uplift*

Therefore, the most formidable problem is not burying continental crust to great depth but finding a mechanism able to bring it back to the surface *without major temperature increase*. Considering very crudely the time constraint of 70 Ma to release 20 kbar would lead to a moderate average uplift rate close to 1 mm a^{-1}. This might, in turn, suggest that erosion alone could have been sufficient to counteract the rise of the isotherms once the subduction was blocked. On the other hand, the negligible temperature increase during uplift is a very characteristic feature, which precludes erosion as a sufficient factor and implies tectonic uplift (see England, this symposuim; England and Thompson 1984). Various tectonic mechanisms are presently explored which may be able to prevent the rise of the isotherms, either by extension of a thickened crust (England, this volume; Platt, this volume) or by upward thrusting of deep units onto cooler ones, as exemplified by Goffé and Velde (1984) in a more external part of the chain.

The latter mechanism may have repeatedly acted during the Alpine evolution, as indicated by the presently exposed metamorphic pattern. Indeed the characteristic feature in the western Alps is that major tectonic contacts bring to exposure metamorphic units (involving continental blocks) the age and grade of which increase stepwise toward the internal part of the chain (figure 4). This pattern reflects the migration both in time and space of intra-continental thrusts within the plate, once crustal shortening could not be achieved anymore by the subduction of complete continental lithosphere (figure 4). The underthrusting of relatively cold panels of continental crust below the overthrusted, higher-grade metamorphic units may be the major factor preventing the rise of the isotherms in the higher-grade units and generating high-pressure, low-temperature metamorphism in the underthrusted, more external unit. This works well in intermediate parts of the chain (Goffé & Velde 1984), where the pressures achieved were less than 10 kbar. Concerning the most internal, deeply subducted continental fragments, such a mechanism works as well from a thermal point of view according to Gillet *et al.* (1985); it seems to be more problematic from a gravitational point of view because it would imply the burial of other, more external units to nearly as great depths. Although this is not completely unlikely, because relevant rocks might not be exposed as yet, alternative possibilities exist, such as the breaking of the subducted plate or the detachment of the (upper?) crust from the rest of the subducted slab, that would restore the buoyancy of the crustal fragment and allow its more rapid uplift. In any event, if it is possible to emplace garnet–peridotite bodies into continental crust, as known in the Central Alps (Evans & Trommsdorf 1978), it must also be possible to uplift more buoyant deeply buried continental crust. In this respect it must be kept in mind that the Ivrea gravimetric anomaly extends southwards through the Sesia zone and the Lanzo ultramafic body to the southern part of the Dora Maira massif. This implies an uplift of mantle material below the most internal and highest-pressure metamorphic units (figure 4).

Clearly, however, there is no unique solution to our problem. As a challenging illustration is the recent fossil finding by Dumont *et al.* (1984) that would imply that sedimentation persisted on some parts of oceanic crust until Upper Cretaceous, while continental units were already deeply subducted. This strengthens the idea that these continental blocks were like peninsulas, but leaves us little hope that we can approach the actual processes with our two-dimensional representations. The combination of detailed geophysical modelling with structural, geochro-

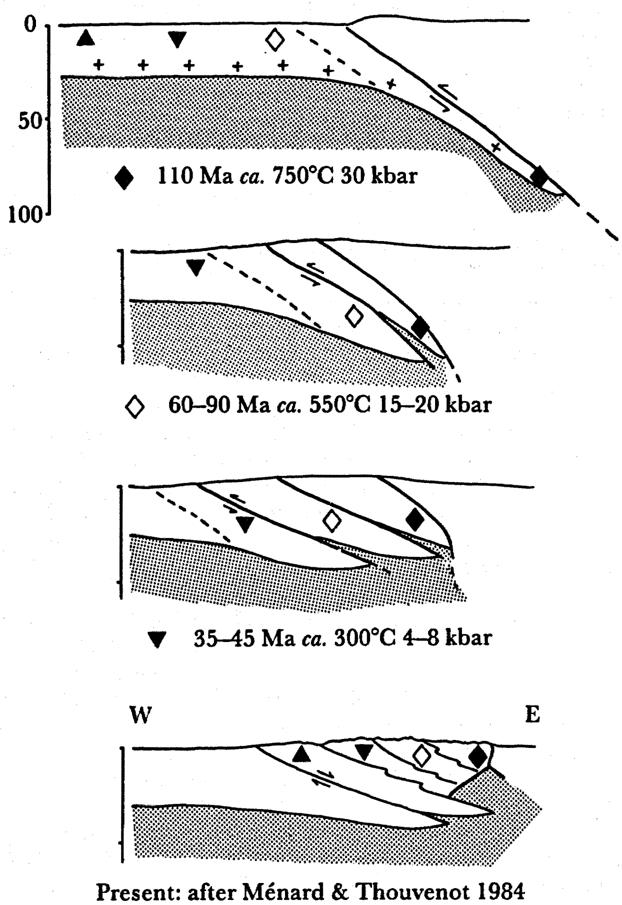

FIGURE 4. Tectonometamorphic evolution of the western Alps. Oversimplified sketch drawing emphasizing the westward migration, once the subduction blocked, of major thrusts within the continental crust, and their possible role in preventing a temperature increase in the early-subducted units, and in creating major metamorphic gaps. Only the European plate is considered here; the eastern, not detailed, part of the sections would show obducted oceanic crust, scraped and subducted continental fragments of the Apulian plate (like the Sesia zone?), and the continental lithosphere of that plate. The four symbols could represent from east to west: Dora Maira, something like Gran Paradiso, internal Briançon zone, and Pelvoux massif, respectively. Although emphasis is put on underthrusting, note that the figure implies important erosion as well.

nologic and petrologic work may help sorting out a most plausible scenario among the many possibilities for this very underdetermined problem. The main contribution of this essentially petrological study is that there must exist a solution.

I wish to express my gratitude to the organizers of the Discussion Meeting for the invitation, to P. C. England and W. S. Fyfe who have drawn my attention to important references, and to T. J. B. Holland and M. J. O'Hara for reviews. Endless discussions with B. Goffé have also materially contributed to this paper.

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Discussion

W. GIBBONS (*Geology Department, University College, Cardiff*). The uplift, preservation and ultimate exposure of high-*P*–low-*T* metamorphic rocks is less likely to result from the growth and development of an accretionary wedge during oceanic plate consumption than from the termination of subduction by some form of plate collision. In the Alps, for example, the main phase of high-*P* metamorphism (as recorded by the oldest radiometric dates) took place around the beginning of the late Cretaceous during the burial and high-*P* metamorphism of 'European' continental crust beneath the whole accretionary system. A modern analogy is perhaps provided by the present subduction of N. Australian continental crust beneath the Indonesian Banda Arc; presumably the leading edge of the Australian plate is now undergoing high-*P*–low-*T* metamorphism. Most blueschists around the world are similarly associated with collisional orogenic belts; even the high-*P*–low-*T* Franciscan Complex of western U.S.A., the uplift of which is not obviously related to plate collision, has been shown by recent geophysical evidence to be underlain by continental crust. Therefore, any model that examines the tectonic processes responsible for the uplift of high-*P*–low-*T* metamorphic rocks should consider the effect of the arrival and subduction of continental crust beneath the accretionary system. In the geological record of formerly destructive plate margins, it is more often the termination of oceanic plate consumption that is recorded, rather than the preceding history of 'steady-state' subduction.

C. Chopin. Obviously. Besides, the concept of an accretionary wedge is probably superfluous in the Alps.

J. P. Platt. It is difficult to reconcile the idea of a mid-Cretaceous continental collision in the Alps with the evidence from plate-motion analysis (Dewey *et al.* 1973) and palinspastic reconstruction (Trümpy 1980) for 400–700 km of shortening in late Cretaceous and Tertiary time. The problem arises because some bodies of continental basement material (such as the Sesia Zone and the Monte Rose massif) were affected by high-*P*-low-*T* metamorphism at a very early stage in Alpine orogenic history. In fact, these bodies of basement were probably crustal fragments that either lay within the Neotethys, surrounded by deep basins underlain by oceanic or greatly attenuated continental crust, or (in the case of the Sesia Zone) were attached to its southeastern (Adriatic) margin. They were themselves probably fairly thin (10–15 km), having been produced during the mid-Jurassic rifting event that produced the Neotethys. Their progressive underthrusting and accretion undoubtedly contributed to uplift (that is the whole thesis of my paper) but these events cannot properly be treated as continental collisions. I think we should reserve the term collision for the entry into the convergent margin of a substantial body of normal-thickness continental crust. This began in Oligocene time, as shown by the start of deformation in the Helvetic shelf, but was probably not complete until late Miocene to Pliocene time (Milnes 1978; Hsü 1979). Most of the preserved high-*P*–low-*T* rocks in the Alps had already reached comparatively shallow depths by the onset of the meso-Alpine metamorphic event, at the beginning of the Oligocene; and their uplift history is therefore largely pre-collisional.

It is certainly not true that most blueschists are associated with collisional orogenic belts in the sense used here. Nearly all the extensive tracts of high-*P*–low-*T* rocks occur in non-collisional active margin settings, mainly around the Pacific: New Caledonia, Papua-New Guinea, Celebes, Sanbagawa belt of Japan, Kamchatka, the Klamath Mountains, the Franciscan Complex, and the Hellenic arc come immediately to mind. Note that recent geophysical evidence for the extension of Sierran basement a few tens of kilometres beneath the eastern margin of the Franciscan (Wentworth *et al.* 1984; Zoback & Wentworth 1986) in no way justifies a collisional interpretation for this belt, which lay at the active leading edge of North America from mid-Jurassic to Miocene time. Blueschist terrains in collisional orogens are, in fact, rare; where preserved, they were clearly both formed and elevated to a fairly high crustal level before the collisional event. More usually, they have been destroyed by the thermal effects associated with collision.

It is undoubtedly true that the elevation of high-*P* terrains depends on the underplating of large volumes of low-density rock. In the Alps this was the Penninic basement massifs; in the Franciscan it was probably immense thickness of Late Cretaceous greywacke. It is also true that underplating alone, whether by subduction or collision, is insufficient as an explanation; we need a mechanism to remove material from above the high-*P* rocks.

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E. V. Artyushkov (*Institute of Physics of the Earth, Moscow, U.S.S.R.*). The existence of coesite in the Dora Maira massif requires not only a burial of continental crust to depths of *ca.* 100 km under a temperature *T ca.* 800 °C. To preserve coesite under a decreasing pressure, this block of continental crust should probably have ascended to the surface quite rapidly. It seems that subduction of continental crust is unable to ensure both a rapid subsidence and rapid return of continental crust to the surface. Another possible mechanism can be suggested.

The present folded structure of the Alps was mostly produced by an intense shortening of deep basins on the attenuated continental crust. These basins formed by rapid subsidence of continental crust without significant stretching (Artyushkov & Baer 1984). The subsidence was caused by the destruction of the lower continental crust under upwelling of the asthenospheric anomalous mantle. One of the possibilities that can be discussed is a transformation of gabbro into eclogite in the lower crust. The latter rock has a high density $\rho \approx 3.55$ g cm^{-3} and should sink into the underlying anomalous mantle ($\rho \approx 3.25$–3.30 g cm^{-3}) of low viscosity.

Some large blocks of dense eclogite sinking into the mantle can be coupled with considerably smaller blocks of sialic rocks. Sinking of the crustal rocks to a large depth will produce a metamorphism typical of very high pressure that can never be attained in continental crust. If detachment of a sialic block from eclogite happens it will ascend to the base of the attenuated crust. In the anomalous mantle, this ascending movement can be quite rapid. For instance, under $\eta \approx 10^{16}$ Pa s a block, several kilometres in size, will ascend to the base of the crust from the depth of *ca.* 100 km in *ca.* 10^4 years. This does not exceed a thermal relaxation time for such a block.

The following crustal shortening results in piling of thin plates of the attenuated crust and an intense erosion. As a result, sialic rocks that ascended to the crust from a large depth can be exposed to the surface.

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Artyushkov, E. V. & Baer, M. A. 1986 Mechanism of formation of hydrocarbon basins: the west Siberia, Volga–Urals, Timan–Pechora basins and the Permian basin of Texas. *Tectonophysics* **122**, 247–282.

C. Chopin. The proposed mechanism may account for rapid uplift of deeply buried upper crustal blocks. However, it is unclear to me how the upwelling of 'asthenospheric anomalous mantle' may lead to eclogite formation in the lower crust, and then to subsidence without stretching. This upwelling is likely to result in a temperature increase, not in the pressure increase required to form eclogite.