More and more evidence is being discovered in Phanerozoic collision belts of the burial of crustal rocks to previously unsuspected (and ever increasing) depths, presently on the order of 150–200 km, and of exhumation from such depths. This extends by almost one order of magnitude the depth classically ascribed to the metamorphic cycling of continental crust, and demonstrates its possible subduction. The pieces of evidence for this new, ultrahigh-pressure (UHP) metamorphism exclusively occur in the form of relics of high-pressure minerals that escaped back-transformation during decompression. The main UHP mineral indicators are the high-pressure polymorphs of silica and carbon, coesite and microdiamond, respectively; the latter often demonstrably precipitated from a metamorphic fluid and is completely unrelated to kimberlitic diamond or any shock event. Recent discoveries of pyroxene exsolutions in garnet and of coesite exsolutions in titanite suggest a precursor garnet or titanite containing six-fold coordinated silicon, therefore still higher pressures than implied by diamond stability, on the order of 6 GPa. The UHP rocks raise a formidable geological problem: that of the mechanisms responsible for their burial and, more pressingly, for their exhumation from the relevant depths. The petrological record indicates that large tracts of UHP rocks were buried to conditions of low $T/P$ ratio, consistent with a subduction-zone context. Decompression occurred in most instances under continuous cooling, implying continuous heat loss to the footwall and hangingwall of the rising body. This rise along the subduction channel – an obvious mechanical discontinuity and weak zone – may be driven by buoyancy up to mid-crustal levels as a result of the lesser density of the acidic crustal rocks (even if completely re-equilibrated at depth) after delamination from the lower crust, in a convergent setting. Chronological studies suggest that the rates involved are typical plate velocities (1–2 cm/yr), especially during early stages of exhumation, and bear no relation to normal erosion rates. Important observations are that: (i) as a result of strain partitioning and fluid channelling, significant volumes of subducted crust may remain unreacted (i.e. metastable) even at conditions as high as 700°C and 3 GPa – with implications as to geophysical modeling; (ii) subducted continental crust shows no isotopic or geochemical evidence of interaction with mantle material. An unknown proportion of subducted continental crust must have escaped exhumation and effectively recycled into the mantle, with geochemical implications still to be explored, bearing in mind the above inefficiency of mixing. The repeated occurrence of UHP metamorphism, hence of continental subduction, through time and space since at least the late Proterozoic shows that it must be considered a common process, inherent to continental collision. Evidence of older, Precambrian UHP metamorphism is to be sought in high-pressure granulite-facies terranes.

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1. Introduction

One of the key features in the theory of plate tectonics is the opposition of a young, continuously renewed and subducted oceanic lithosphere, to an old and everlasting continental lithosphere. In this paradigm, continental crust was assigned, because of its low density, the role of a floating object which could be accreted, deformed, thickened and eroded in collision zones, but the fate of which was restricted to an erosion–sedimentation–metamorphism/melting cycle encompassing at most the 30 or 40 outer kilometers of the solid Earth. The existence of crustal mountain roots as deep as 70 km was evident from gravity and seismic data, but it was unclear whether samples could ever be brought from such depth to the surface by tectonic processes.

A first hint was given by the recognition that rocks of oceanic origin in the central Alps, after they were hydrothermally transformed under near-surface conditions, were metamorphosed to conditions of ca. 3 GPa (i.e. near 100 km depth) [1] and that the enclosing continental basement may have shared the same burial [2]. In the early 80s the successive discoveries of coesite, a high-pressure polymorph of quartz, in the western Alps [3] and the Caledonian chain of Norway [4] led to the inescapable conclusion that continental crust can be buried to 100 km depths and be brought back to outcrop during the course of orogenic processes. The geological problems at stake became even more formidable after the discovery of demonstrably metamorphic diamond in the Cambrian orogen of the Kokchetav massif, Kazakhstan [5], implying minimum formation pressures of about 4 GPa, i.e. burial to depths exceeding 120 km.

These findings, which rely entirely on careful microscopic observation, have been the incentive for a blooming field-oriented high-pressure research activity. A wealth of new findings over a decade confirms worldwide the generality of the early discoveries, brings answers to the more pressing questions but also pushes ever deeper the limits of continental metamorphism, through the recognition of ever more tenuous indicators of ever higher pressures – and so raises new questions. We examine here these recent developments – made in parallel with those of mantle mineralogy [6] as revealed by high-pressure experimental work and by mineral inclusions in ‘super-deep’ kimberlitic diamonds – as well as their limitations and implications.

2. The force of mineralogical evidence

2.1. The classics: simple polymorphic indicators

The recognition of ultrahigh-pressure metamorphic rocks has essentially relied on the identification of the high-pressure polymorphs of silica (coesite) or of carbon (diamond). The simplicity of the first-order phase transitions involved indeed makes the very presence of one such phase, whatever its grain size or abundance, an immediate indicator of minimum pressures attained, on the order of 3 and 4 GPa, respectively (Fig. 1). The underlying assumption that these phases formed in their stability field is borne out by independent lines of evidence, as shown below.

The rock mineralogy tending to adapt to changing pressure (P) and temperature (T) conditions, such UHP phases normally disappear during decompression, i.e. during the movement of the rock toward the surface (= exhumation), this the more so if the rock enters the stability field of the respective low-pressure phase at high temperature. As a matter of fact, the metastable persistence of these indicators up to surface conditions is commonly the result of their inclusion in mechanically strong host minerals like garnet or zircon (Fig. 2A,B,D). These act as pressure vessels around each inclusion, insulating it from metamorphic fluid and preventing the volume increase of the back-transformation, i.e. maintaining the inclusion pressure on the relevant transition curve.
as long as the tensile strength of the host mineral can sustain the difference between the internal (including) and external pressure. Although early anticipated [7], the effectiveness of this mechanism was only recently demonstrated for small (10–20 μm) coesite inclusions in garnet and zircon, in which Raman microspectroscopy revealed residual internal pressures that could exceed 2 GPa [8,9]! Such residual pressures are also evidence against metastable coesite formation. Note that, once the host mineral is fractured (Fig. 2A) – or in the absence of a ‘container’ – there is no other obstacle than kinetics to the complete transformation of the high-PR phase into the lower-P polymorph, i.e., for coesite/quartz, temperatures lower than about 375–400°C [10] or a completely dry system [11].

2.2. Relics and pseudomorphs: the limits of evidence

The tiniest relic of a UHP indicator in a metamorphic rock is in itself compelling evidence that the rock passed at some stage through UHP conditions, and this may imply entire reconsideration of the geological or structural record for the area. Given these far-reaching implications, a prerequisite is that the mineralogical evidence is unambiguously characterized – and demonstrably in situ. Techniques with high spatial resolution, such as Raman microspectroscopy, are invaluable complements of optical observation in this respect. Yet, the limit of what can be done convincingly has probably been reached with a weak coesite Raman signal in what is optically a polycrystalline quartz inclusion in Antarctic eclogite [12], or with the report of diamond from the Rhodope, Greece [13], in which the Raman spectrum of ‘diamond’, usually an intense sharp band, can be interpreted as well as that of a defective graphite [14]. Besides, diamond is so prone to artifact, to be plucked off or introduced into a sample during processing (sawing, grinding) that utmost care and special techniques are required to ensure its identification in situ (e.g. [15]), the safest way being to work exclusively with crystals that are entirely included in another mineral, small size and confocal techniques permitting [5,16]. This may be the reason why the original reports of large UHP diamond from Dabie Shan, eastern China [17], have not yet been independently confirmed by micro-findings.

However, absence of evidence is not evidence of absence, and this may be a golden rule for UHP studies. The original coesite report from the Caledonides [4] could not be duplicated for a decade in spite of the efforts of several expert groups – until a Ph.D. student was able to demonstrate the regional occurrence of coesite there [18,19]. New eyes are sometimes required to open frontiers...
Fig. 2. Photomicrographs of characteristic or potential mineral indicators of UHP conditions (plane-polarized light unless otherwise specified). (A) Coesite (coes) inclusion in pyrope garnet (gt); note the incipient transformation into quartz (qz) and, as a result of the related volume increase, the radial cracks in the host garnet [3]; Parigi, Dora-Maira massif. (B) Ellenbergerite (Elb) included in pyrope garnet along with rutile (ru); Parigi, Dora-Maira massif. This Mg–Al–(Ti,Zr)-silicate contains 8 wt% H$_2$O and is stable at pressures exceeding 2.7 GPa (and $T < 725^\circ$C). (C) Oriented quartz needles precipitated within omphacitic clinopyroxene (cpx); Blumenaue eclogite, Erzgebirge [86]. This feature is often encountered in UHP eclogite. (D) Microdiamond (diam) inclusions in zircon (zr); garnet gneiss, Seidenbach, Erzgebirge [16]. (E) Oriented orthopyroxene (opx) precipitates in garnet, evidence of a former super-silicic, majoritic garnet; Ugelvik peridotite, Otterøy island, western Norway [28]. (F) Oriented K-feldspar (Kfs) precipitates in clinopyroxene; crossed nicols; pyroxenite in marble, Kumdi Kol, Kokchetav massif.
The evidence may also become more elusive, when the indisputable relics of UHP minerals have disappeared and only pseudomorphs thereof, i.e. breakdown products retaining the original shape of the parent mineral, are present. Graphite octahedra [20] are a straightforward example of pseudomorphs after diamond, but what is the diagnostic value of graphite cuboids in schist [21]? Likewise, the texture of quartz pseudomorphing coesite (Fig. 2A) is characteristic [7] but tends to disappear through recrystallization upon prolonged thermal annealing. What then is the diagnostic value of a well-annealed polycrystalline quartz inclusion in fractured garnet? Clearly at some stage the force of the evidence declines, and only an array of converging hints may then convince the sceptical petrologist.

2.3. Beyond the polymorphic record: deeper and deeper with other UHP indicators

The intensive study of ascertained or potential UHP rocks has revealed a number of uncommon mineralogical features and assemblages, of more or less definite UHP value. Some of these mineralogical hints for UHP are illustrated in Fig. 2; many have recently been reviewed [22,23], so only the most salient or new features are addressed here. Importantly, none of them involves a polymorphic transition, thus avoiding the suspicion bearing on such phase transitions, namely that they may be affected by high deviatoric stresses (e.g. [3]) – which is definitely not the case for microdiamond in the Erzgebirge UHP gneisses, which precipitated as a daughter mineral in supercritical COH fluid/melt inclusions in garnet [24, 25].

2.3.1. K-bearing clinopyroxene

The occurrence, in calc-silicate rocks of the diamond-bearing Kokchetav massif, of clinopyroxene with extremely high potassium contents (up to 1.5 wt% K₂O) [5] was a confirmation of stable diamond occurrence. So far solely known from a few eclogite xenoliths in kimberlite, this feature was experimentally shown to be stable at pressures of 4–10 GPa. Upon decompression this potassic pyroxene shows characteristic textures, with oriented precipitates of K-feldspar (Fig. 2F) [5, 26], sometimes of a later K-bearing mica [27].

2.3.2. Majoritic garnet

Another most significant discovery is, in the peridotite body of Otøy Island, western Norway, that of microtextural evidence for exsolution of orthopyroxene in coarse garnet, implying a ‘super-silicic’ precursor garnet, i.e. the presence of several mol% majorite component [28,29]. It was known from experiment [30] that with increasing pressure pyropic garnet can incorporate more and more MgSiO₃ as majorite component (to fit the garnet formula, better written Mg₃Mg⁶Si⁴Si₃O₁₂, in which one-fourth of the silicon atoms is six-fold coordinated). Modal estimates of the majorite component originally present in the Norwegian garnet, i.e. before exsolution, lead to formation pressures exceeding 6 GPa ( > 200 km). Such sensational exsolution had been discovered in garnet from kimberlite xenoliths a few years earlier [31] but it was completely unexpected that the delicate textures (Fig. 2E) could be preserved during the long ascent and slow cooling of an orogenic peridotite body, as opposed to quenching by explosive kimberlite sampling. Admittedly, mantle rather than crustal material was involved in either case, but the demonstration was made that, even with the rates attending metamorphic processes, such evidence could be preserved – opening the prospect that it may be found in UHP crustal rocks too.

Since then, indeed, three reports of related textures were made in UHP terranes. In garnet–peridotite of the Su-Lu terrane, eastern China [32], the textural evidence of pyroxene exsolution from garnet is convincing but also involves oriented micrometer-size rods of rutile and apatite; however, it is unclear whether the extreme conditions recorded by the ultramafic body were shared by the associated UHP (coesite-grade) crustal rocks. As in Norway [28], cf. [33,34]), the extreme conditions may record an earlier, deeper mantle stage before the body was amalgamated with crustal material during ‘normal’ UHP metamorphism. In two other reports from the Greek Rhodes [13,35], the crustal nature of the rocks is clear but the evidence of exsolution from a former
majoritic garnet is less compelling [14]; in garnet from kyanite-biotite gneiss, no pyroxene occurs but oriented rods of quartz and rutile [13]; and in garnet from metabasite, with oriented rutile rods, clinopyroxene occurs as blebs [35], which may or may not represent recrystallized exsolution products (alternatively they could be inclusions trapped during garnet growth).

2.3.3. ‘Super-silicic’ titanite

Other spectacular evidence for the metamorphism of crustal material at unsuspected depths is the discovery of coesite precipitates in titanite, CaTiSiO$_5$, in marble from the Kokchetav massif [27]. This suggests the existence of a super-silicic precursor titanite, as experimentally anticipated in a high-pressure study of the series CaTiSiO$_5$–CaSi$_2$O$_5$, in which the Ca-silicate end-member is stable above 8 GPa and has the titanite structure with 50% six-fold coordinated silicon [36]. The precursor compositions reconstructed by integration of the exsolved phases (also minor calcite and apatite) point again to pressures that may have exceeded 6 GPa [27], i.e. near 200 km burial. Paradoxically, coesite in such a case plays the role of a decompressional, lower-pressure phase!

A general feature of these UHP conditions, whether they approach the 3 GPa/650–800°C range in ‘standard’ coesite-bearing terranes or the 4–6 GPa/900–1000°C range in the diamond-bearing terranes, is that they represent a low $T/P$ ratio at mantle depths. The most likely if not the sole geodynamic context in which such a low ratio is realized is subduction zones.

3. UHP through space and time

After an incubation period of 5 years after the first coesite reports, more and more evidence for UHP metamorphism is being discovered worldwide (e.g. [23]), even in orogens where high-pressure metamorphic rocks were so far virtually absent or terribly underestimated, like the Himalaya.
In short, definite evidence of UHP metamorphism is now missing only from Australia and North America (Fig. 3).

3.1. Coherent UHP terranes

A major result concerns the actual extent of UHP rocks within a given metamorphic terrane. Pieces of definite evidence for UHP remain rare and occur in rock types (eclogitic metabasite, magnesian schist, marble or pelitic metasediment) that are subordinate with respect to the overwhelming country rock of felsic gneiss (metagranite), in which evidence for UHP has long remained elusive. Yet the areal extent of UHP occurrences is over kilometers in the Alps [40], tens of kilometers in Norway [19] and the Kokchetav massif [9], and hundreds of kilometers along the Hong’an-Dabie-Su-Lu Triassic belt of east central China (e.g. [22]).

The key point is whether the UHP rocks and their country-rock gneiss, which bear only low-grade assemblages, share the same UHP history or were tectonically amalgamated during a later stage, i.e., whether the metamorphic terrane as a whole forms a coherent UHP terrane or a mixture of UHP and low-pressure rocks. New problem, new methods: beside extensive zircon U-Pb point-dating [41] and stable-isotope data [42], the systematic study by Raman spectroscopy of the minute mineral inclusions in zircon concentrates extracted from the country-rock gneisses has proved to be most effective in unraveling the former extent of UHP conditions in rocks that bear now only low-grade mineral assemblages, especially in the huge Dabie-Su-Lu belt [43–45].

There the resulting evidence is that the gneisses, whether of sedimentary derivation (paragneiss associated with eclogite and marble), or of magmatic derivation (orthogneiss, former granite forming large tracts of country rock), both commonly bear coesite relics in zircon, thereby demonstrating the coherency of the UHP terrane. A similar conclusion was reached in the small UHP terrane (15×5×1 km) of the Dora-Maira massif, western Alps [40] and might hold in Norway [18,19,46], as well as in the Erzgebirge [47] and the Kokchetav massif [9], in both of which micro-diamonds are also gneiss-hosted. However, the implications are particularly formidable in China, given the scale of the UHP terrane: several hundred kilometers in length, a few tens of kilometers lateral exposure, and some kilometers thickness! Obviously large tracts of continental crust were involved in UHP metamorphism.

3.2. Shaping the continents

One of the most fascinating aspects of recent findings is the repeated occurrence through time of UHP metamorphism of continental crust. The composite Eurasian continent bears the scars of the successive orogenies that make it a collage of microcontinental blocks. In each of these scars UHP rocks do occur, marking the steps of continental accretion (Fig. 3): from the Cambrian Kokchetav massif in Kazakhstan (530 Ma [48,49]), to the Early Paleozoic central China belt (Altun-Qaidam and Qinling; ca. 500 Ma [50], Jingsui Yang et al., article in revision), to the Triassic Hong’an-Dabie-Su-Lu belt (220–240 Ma, e.g. [41]), to the Cretaceous merging of the Lhasa block prolongation in Indonesia (110–120 Ma [51]), to the Tertiary Himalayan belt (45 Ma?). On the western side of the Eurasian block, UHP crustal metamorphism is likewise well represented in the Caledonian chain (Laurentia–Baltica collision) from eastern Greenland [52] to Norway, in the Variscan belt (Laurentia–Gondwana collision) from Central to Western Europe (Bohemian massif [16] and Monts du Lyonnais, French Massif Central [53], respectively), but also in the Meso-Cenozoic Alpine chain (Europe–Apulia collision), from the Rhodope massif (70–75 Ma [35]) to the western Alps (38–35 Ma [54–56]). By contrast, the processes active along the circum-Pacific margins seem to be unsuitable to produce or, more likely, to exhume UHP rocks. In the case of Eurasia, the UHP material involved may be of oceanic origin in a few instances (Sulawesi, Monts du Lyonnais, Cignana) but continental in most cases, attesting to the general rule that continental subduction has operated at least since the late Proterozoic.

This apparent age limit as well as the paucity...
of Precambrian high-pressure, low-temperature rocks (blueschists and eclogites) have led to the suggestion of a rapid decrease of the subduction-zone thermal gradient during the late Proterozoic, after a massive heat release made possible by the dislocation of the Rodinia supercontinent [57]. However, the existence of Proterozoic UHP metamorphism in the Pan-African belt of Mali (ca. 600 Ma [58]) and Brazil [39], of 2 Ga old eclogites in the Usagaran belt of Tanzania [59] and of 2.6 Ga high-pressure granulites in the western Canadian shield [60] indicates that zones of low \( T/P \) ratios did exist early in the Earth’s history. The preservation of the HP/UHP rocks formed was admittedly more difficult because of the Earth’s higher heat production, and the likely higher average geothermal gradient and lesser plate thickness. This may be sufficient to account for the paucity of preserved Precambrian blueschists and eclogites.

The important point is that in collision belts – as opposed to Pacific-type margins – subduction of continental crust to depths reaching 100–200 km appears to be the rule since at least the Proterozoic. The implications in terms of tectonics and geodynamics are far-reaching. Were it simply for the Himalaya, most of what has previously been considered relevant for the archetype of the collision belts – in terms of thermal regime, metamorphism, partial melting, uplift and cooling – actually pertains to a late stage of the chain evolution: metamorphism of continental crust started about 20 Ma earlier with the subduction of the leading edge of the Indian plate to coesite-forming depths (90–100 km); the present-day shallow dip of India below southern Tibet bears no relation to an initially much steeper dip, and the average exhumation rates are twice as high as previously thought (ca. 10 mm/yr) [37].

4. A window into subduction-zone processes

Another important aspect of these findings is that the exposure of large tracts of HP/UHP rocks represents windows opening into subduction-zone processes. Aspects thus accessible to observation and measurement include the extent of mineral transformation and the role of fluids and deformation under such conditions, the degree of interaction with mantle material – in the footwall of the subduction zone or in the overlying mantle wedge – and its possible geochemical signature. Admittedly, the message is in essence a palimpsest that has to be deciphered, since the features acquired under UHP conditions may have been altered, overprinted or erased during the later decompression history. Yet it is a unique record.

4.1. Extent of rock transformation at UHP

A repeated observation in HP/UHP terranes is that significant rock volumes may escape deformation and mineral equilibration at the prevailing \( P-T \) conditions, despite temperatures as high as 650–750°C for UHP rocks. Famous examples are the metagranite body of Flatraket in the Western Gneiss Region, Norway [61], or the Brossasco metagranite in the Dora-Maira massif [62] (Fig. 4). Pure volume diffusion is clearly an inefficient process: the partitioning of deformation and the related fluid access, both triggering nucleation and mass transfer [63,64], are actually much more efficient than temperature.

This has led thermomechanical modelers to consider not only the changing mineralogy according to \( P-T \) conditions, in order to account for changing physical properties like density or...
rheology, but also the effectiveness of this transformation (e.g. [65–67]), i.e. metastable persistence or absence thereof. The consequences in terms of density changes and buoyancy are significant (see below).

4.2. Isotopic behavior at UHP

This contrast between reactive and metastable systems, regardless of the high temperatures attained, is also entirely reflected in their geochemical and isotopic behavior, thereby questioning the meaning and the applicability of closure temperatures. Monazite U–Pb ages in undeformed facies of the Brossasco metagranite are intermediate between the Hercynian intrusion age and the Alpine age of the UHP event, whereas they are definitely Alpine in nearby UHP rocks like magnesian schists [68] which were completely reworked during Alpine times. The same holds for U–Pb in titanite, for which formation ages allow one to date the successive steps of UHP metamorphism, incipient overprint and low-grade overprint, depending on the rock type addressed [56]. Evidently the rocks showing the largest mineral metastability under UHP conditions show Rb–Sr disequilibrium among their phases [69] and are the worst target for Ar–Ar dating (e.g. [70]).

Independent of the problem of the equilibration scale, there is an inherent problem to K–Ar systemsatics in UHP terranes. Under UHP conditions the partial pressure of argon is expected to be one order of magnitude higher than in low-pressure terranes, leading to significant argon incorporation during the growth of some UHP phases, in particular white mica. The result is commonly that the K–Ar ages of UHP micas (even Ar–Ar plateau ages) are paradoxically higher than those derived from the more retentive U–Pb or Rb–Sr systems, and geologically meaningless [70,71].

4.3. Fluid behavior: open vs. closed system, mantle–crust interaction, rheology

In contrast to reports of low-pressure terranes flushed by ‘oceans’ of hydrothermal fluids, the experience gained in HP and UHP terranes is that of very limited fluid flow and at best centimeter-scale isotopic equilibration, in spite of the uncommon P–T conditions. As summarized by Rumble [42], ‘stable isotope evidence denies the existence of a pervasive fluid free to infiltrate across lithologic contacts during UHP metamorphism. Furthermore, the preservation of high-temperature oxygen isotope fractionations among minerals argues against the presence of free fluid after peak metamorphism, during exhumation and cooling. Residence in the upper mantle had no discernible metasomatic effect on the stable isotope composition of crustal rocks subducted during continental collision’. The best evidence for this is the preservation, throughout the UHP event in Dabie-Su-Lu, of a huge oxygen and hydrogen isotope negative anomaly that dates back to a premetamorphic, Neoproterozoic hydrothermal system involving meteoric water from a cold climate [72]. This does not mean the rocks were dry: most UHP assemblages bear hydrous minerals like micas or epidote that are stable under these conditions, and therefore evolve little fluid.

The presence or absence of a fluid also directly affects the temperature of partial melting (Fig. 1). Therefore, depending on both fluid availability and the P–T path followed, in particular during decompression, melting may be expected to be minimal or quite general. Interestingly, under UHP conditions there may be complete miscibility between silicate melts and hydrous fluids for a range of compositions, with the consequence that the supercritical fluids generated at UHP by dehydration reactions during subduction can dissolve considerable amounts of silicate material with increasing T, and therefore be of restricted mobility (as compared to the more hydrous fluids produced at lower pressure, in the subcritical range) [73,74]. Besides, the presence of such fluids, whenever water is available, has a decisive weakening effect on the rheological strength of rocks during deep subduction, precluding notable shear heating [75].

5. Exhumation: rates and processes

The realization that crustal segments may re-
turn to the surface from depths exceeding 120 km raises a difficult geological problem: that of the mechanisms responsible for their burial and, more pressingly, for their exhumation from such depths. The petrological record indicates that large tracts of crustal rocks were buried to UHP conditions of low $T/P$ ratios, which is consistent with and best accounted for by a subduction-zone context. It has indeed long been recognized by modeling that several hundred kilometers of the leading edge of a continental plate can be entrained into a subduction zone by the sole effect of body forces, all the more so as the continental margin is thinner and the cohesion between continental and sinking oceanic lithosphere is higher. Continental subduction during incipient collision is therefore an easy way to account for the relevant burial.

As to exhumation, the petrological record in most instances constrains decompression to have occurred under continuous cooling [3,9,23,40,64], implying continuous heat loss to the footwall and/or hangingwall of the rising body. In addition, isotopic dating shows that formation ages and cooling ages obtained on UHP rocks span a narrow range of a few million years in most instances [35,41,48,49,54–56,76], implying high cooling rates and, through the petrological constraints, uncommonly high decompression rates, especially during early stages of exhumation. The exhumation rates obtained (up to 1–2 cm/yr) are indeed more akin to plate velocities and bear no relation to normal erosion rates. It was soon recognized that erosion and tectonic processes like extensional faulting or ‘corner flow’ are unable to account for the critical, early stage of extrusion, even if they may become effective during later stages, at higher structural levels. Besides, it is certainly not fortuitous that most UHP terranes presently exposed are essentially made of light, upper-crustal material, thus pointing to the role of buoyancy as a main (?) factor of exhumation (e.g. [77]). The constraint of continuous cooling precludes any mechanism like diapiric rise across the overlying mantle wedge and makes the return back along the subduction channel – a ‘two-way street’ and obvious mechanical discontinuity – the most likely exhumation path. The process is essentially driven by buoyancy up to mid-crustal levels, as a result of the smaller density of the acidic crustal rocks (even if completely re-equilibrated at depth), after delamination from the lower crust. Field data (e.g. [76,79,80]) as well as physical [81], thermal [82] or thermo-mechanical [66,67,87] modeling support the idea that most of the exhumation of UHP rocks may be achieved by such ‘extrusion’ of light crustal material along the subduction channel, in a continuously convergent setting, this without the need to resort to slab break-off as a ‘deus ex machina’.

6. Consequences and prospects

A new picture emerges for the fate of continental margins and microcontinents. Continental subduction must now be considered a normal process, inherent to continental-plates convergence, collision and orogeny – at least since the late Proterozoic. Part of (most of?) the subducted upper crust may be delaminated and exhumed, bringing testimony of subduction-zone processes back to the surface, with a clear geochemical message: the inefficiency of mixing. Such a message could help us shape our view of the deeper mantle in terms of its heterogeneity and inheritance from subducted slabs (e.g. [83]). An open question is still the extent to which continental lithosphere, upper and lower crust possibly included, may actually disappear into subduction zones and contribute to magma generation [78], mantle geochemistry, geophysical signature, etc. Another concern is that our present view of UHP metamorphism might be intrinsically biased by the necessary preservation of a mineralogical record. It may well be that subducted continental fragments were exhumed and decompressed under near-isothermal or even increasing temperature conditions, leading to dehydration reactions and partial melting (Fig. 1). This would likely erase most records of the early UHP stage. High-grade, granulite-facies metamorphic terranes are therefore also open to UHP detective work [84].

Given the exploding number of UHP findings during the past few years, the author is confident that UHP metamorphism will sooner or later be
recognized on all continents and back into the Precambrian as well. An important issue will be increasing chronological and spatial resolution in the dating of the highest-pressure and early-exhumation stages, for instance in the dating of successive growth zones of crystals. Conversely, the ever-increasing spatial resolution of characterization techniques also hints at one of the limits to be encountered. More and more polymorphs having a UHP thermodynamic stability field will be identified at the submicrometer scale, down to ‘crystals’ having only a few unit-cells extension; thereby the limit is reached under which surface energies may become preponderant over enthalpic properties, casting some doubt as to the actual P–T formation conditions. The recent findings of nanometer-size TiO$_2$ with the α-PbO$_2$ structure in Erzgebirge [85] show that this concern is already no longer a purely theoretical one.

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