Mineralogical and experimental evidence for very deep exhumation from subduction zones

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Abstract

In 1996, we proposed that the Alpe Arami lherzolite, Switzerland, contains evidence within it implying origin at a depth of greater than 300 km. Suggestion of such extraordinary depth of exhumation of these rocks has been controversial. Principal amongst the original evidence was a very high concentration of oxide precipitates in olivine. From the abundance, morphology, crystallography, and topotaxy of these oxides, it was argued that the inferred very high solubility of highly-charged cations (principally Ti and Cr) represented a previously unrecognized mantle environment and that pressures in excess of 10 GPa were the most likely explanation of the observations. We have now successfully completed high-pressure experiments to test whether there are any conditions under which olivine can dissolve the high concentrations of TiO2 we originally inferred (>0.6 wt%). We answer the question in the affirmative for P≥10 GPa, consistent with our hypothesis. In addition to these experimental results, we also have discovered in the same rocks exsolution lamellae of clinoenstatite in diopside. These lamellae contain antiphase domains which indicate that the originally-precipitating phase was a C2/c pyroxene; additional geologic and crystallographic observations strongly suggest that the precipitating phase was high-pressure clinoenstatite, thereby providing independent evidence of a minimum depth of origin of this massif of 250 km. To those observations, we add here exsolution of SiO2 from omphacite of Alpe Arami eclogite. Discoveries similarly implying very great depth of exhumation of mantle rocks and/or subduction to and return from such depths are now known from other continental collision terranes. In particular, recent discovery of exsolution of pyroxenes from garnets in the peridotites of the Western Gneiss terrane of Norway provides unambiguous evidence of a very deep origin for these peridotites (>185–200 km), and discovery of microdiamonds in metasediments of the Erzgebirge of Saxony adds to the growing list of continental terranes exhumed from >100 km. Realization that such deep subduction and exhumation have occurred multiple times
spanning the entire Phanerozoic strongly suggest that this phenomenon is a normal process of continental collision rather than a bizarre curiosity. Whether such exhumation is a single- or multiple-step process is an important question for future research. © 2000 Elsevier Science Ltd. All rights reserved.

1. Introduction

It was postulated previously that precipitation of abundant (~1 vol%) FeTiO₃ rods in olivine of the Alpe Arami garnet lherzolite, Ticino, Switzerland, provided evidence of a previously unrecognized geological environment, in which TiO₂ has much higher solubility than under other conditions represented by crustal peridotites and mantle xenoliths (Dobrzhinetskaya et al., 1996). From the observed shape and inferred crystallography of titanate rods, it was postulated that the new environment represents very high pressure conditions. Despite the controversy that has been stimulated by this proposal (Hacker et al., 1997; Green et al., 1997a), we have continued to study these rocks and have performed experiments to test our hypothesis of high solubility of TiO₂ in olivine at high pressure. We have found that: (1) these rods are older than any other microstructural features of the rock (Green et al., 1997b); (2) the solubility of TiO₂ in olivine increases markedly with pressure and is quantitatively in agreement with our original proposal that the rocks may have originated at depths in excess of 300 km (Dobrzhinetskaya et al., 1999); (3) exsolution of probable high-pressure clinoenstatite in diopside in the same Alpe Arami specimens as display the abundant FeTiO₃ rods provides independent evidence supporting depth of origin greater than 250 km (Bozhilov et al., 1999). Here, we review these findings and describe exsolution of quartz from omphacite in eclogites, a phenomenon not previously described from Alpe Arami, but well-known from deep mantle xenoliths, high-pressure experiments, and also has been described from other ultra-high-pressure metamorphic rocks. The new observations of others similarly documenting exhumation of rocks from extreme depths in the mantle are also discussed to explore some of the implications of such very deep rocks in subduction-zone terranes.

2. Alpe Arami olivine

Olivine in the Alpe Arami massif consists of two readily-recognizable generations — an older generation displaying exsolution of up to 1% by volume FeTiO₃ rods oriented parallel to [010] (Fig. 1), and a younger generation devoid of precipitates. Electron-diffraction evidence was originally interpreted to suggest that these rods had several different crystallographic structures, including ilmenite (Dobrzhinetskaya et al., 1996). In our continuing work, we have been unable to verify the postulated metastable phases by completing the characterization of their structures, suggesting that all of the rods may be ilmenite (for discussion, see Hacker et al., 1997; Green et al., 1997a). The ilmenite precipitates display toptaxy of [0001]lim // [100]ol; [1120]lim // [010]ol; [0110]lim // [001]ol, exactly as reported previously for ilmenite exsolution in olivine (Mosely, 1981; Drury and van Roermund, 1988). However, the precipitates in Alpe
Arami olivine are in much greater abundance than previously reported and their rod shape is inconsistent with the disk shape observed for naturally precipitated ilmenite.

The older generation of olivine in the Alpe Arami massif also exhibits a lattice preferred orientation (LPO) in which [100] is normal to the foliation (Möckel, 1969; Buiskool Toxopeus (1976, 1977a, 1977b)). This pattern is unique to these rocks and is incompatible with the
known dislocation glide mechanisms that are responsible for the common LPO of olivine that develops in other peridotites (e.g., Nicolas and Poirier, 1976; Green, 1976) and is responsible for the seismic anisotropy of the shallow oceanic mantle. These glide mechanisms are readily reproducible in deformation experiments in the laboratory (e.g., Raleigh, 1968; Durham et al., 1977; Zeuch and Green, 1984) and produce a set of very well-defined dislocation configurations (e.g., Green and Radcliffe (1972a, 1972b); Durham et al., 1977; Jin et al., 1989). The younger generation of Alpe Arami olivine exhibits the normal LPO and contains no titanate rods; ilmenite is found in the interstices between the other crystals. Examination of the dislocation substructure of olivine by the technique of internal oxidation (Kohlstedt et al., 1976) showed that the dislocation configurations in both generations are the same and are those that are to be expected for the normal LPO of olivine (Green et al., 1997b). Such examination also showed clearly that the ilmenite rods are older than the dislocation substructure and have interfered with propagation of the later dislocations of the \( b = [100] \) type (Fig. 2). The second generation of olivine is associated with hydrous alteration of the garnets and pyroxenes and is clearly Alpine in age. The first generation developed its LPO and exsolved the titanate rods under unknown conditions at an unknown earlier time.

3. Experimental study of \( \mathrm{TiO}_2 \) solubility in olivine

In order to test our hypothesis of very high pressure origin, we conducted a series of
experiments on lherzolite powders with added ilmenite at high pressure and high temperature and measured the solubility of TiO$_2$ in olivine in those experiments. The results for nominally anhydrous compositions are shown in Fig. 3 (Dobrzhinetskaya et al., 1999). Only experiments held at the experimental conditions for $\geq$12 h are shown here; shorter-time experiments showed time-dependence of TiO$_2$ contents of olivine, indicating lack of equilibration. For times longer than 12 h, repeated experiments at the same conditions showed no significant differences. Circles in Fig. 3 indicate single-stage experiments in which the starting material was held at the indicated conditions and then quenched. Two-stage experiments were also conducted, in which the specimen was first held at a chosen $P$ and $T$ and then subsequently subjected to a second anneal at lower $P$ or $T$ under conditions where the solubility of TiO$_2$ would be lower. Ellipses in Fig. 3 represent the conditions of the second stage of these experiments. These reversal experiments not only confirmed the solubilities measured in single-stage experiments, but also displayed exsolved disks of ilmenite showing the same topotaxy as determined previously for ilmenite precipitates in olivine (Mosely, 1981; Drury and van Roermund, 1988).

Under conditions comparable to those indicated for Alpe Arami by thermobarometry ($\sim$5 GPa, 1400 K), solubility of TiO$_2$ was below the sensitivity of our EDS X-ray technique (i.e., $<0.1$ wt%). The solubility of TiO$_2$ in olivine implied by the image-analysis measurements ($\sim$1 vol% FeTiO$_3$) is approximately 0.6 wt%. Hence, the data of Fig. 3 indicate that such high solubility would only be expected at pressures of at least 8–10 GPa, depending on temperature (but see Section 6). Exploratory experiments in this same system but with added antigorite as a source of H$_2$O also were performed (Dobrzhinetskaya et al., 1998). Under all conditions investigated, olivine reacted with the fluid to form high-density titaniferous hydrous phases such as titanoclinohumite and titanochromrodrite (Wirth et al., 1999) and the stable titanium oxide was rutile; none of these phases are present in Alpe Arami lherzolite, except for the latter which occurs as precipitates in garnet. Olivine in these charges showed TiO$_2$ content below the

Fig. 3. TiO$_2$ solubility in olivine as a function of pressure and temperature. Circles represent single-stage experiments; ellipses represent the second stage of two-stage (reversal) experiments. Open symbols signify that the coexisting (Fe,Mg)TiO$_3$ oxide contained Si; solid symbols indicate no Si in the oxide. Lower case ‘r’ signifies coexisting rutile. Asterisk is an experiment from Green et al. (1997b). Figure from Dobrzhinetskaya et al., 1999.
EDS detection limit. Similar results for hydrous systems have been reported by Ulmer and Trommsdorff (1997).

4. Alpe Arami diopside

Yamaguchi et al. (1978), in a short paper about amphibole lamellae in diopside of Alpe Arami lherzolite, mentioned in passing that they had found clinenoestatite lamellae. In 1978, the existence of high-pressure clinenoestatite (HPclen) was not known. At the end of Green et al. (1997b), we pointed out that if those lamellae could be relocated, they could be examined for antiphase domains (APDs) that, if present, could identify the originally exsolving phase as HPclen. We have been successful in relocating clinenoestatite lamellae in Alpe Arami diopside (Bozhilov et al., 1999).

Diopsides from two microenvironments were examined — clumped around and included within garnets, and scattered in the olivine + enstatite matrix. In the latter group, diopside exhibits a nearly-perfect substructure except for widely-spaced lamellae of amphibole. Diopsides of the former group, however, exhibit 1–5 vol% clinenoestatite lamellae (Fig. 4(A)) with space group P2₁/c determined by electron diffraction (Fig. 4(B)). The clinenoestatite lamellae and diopside host share the b axis, with their c axes subparallel at an angle of about 2°. The lamellae lie at an angle of 20–24° from the (100) planes of both phases, approximately parallel to (010) of diopside; bright field images exhibit blotchy contrast within the lamellae on a wave-length of a few nanometers, which is not observed in the diopside host (Fig. 4(A)). Dark field imaging of the lamellae with h + k = even reflections show the same mottling, with alternating darker and lighter contrast regions elongated across the lamellae (Fig. 4(C)). By contrast, imaging with h + k = odd shows generally equidimensional domains that are smaller than the strain domains and exhibit much higher contrast (Fig. 4(D)). These marked differences between dark field images with h + k = odd and h + k = even suggest that the dark areas within the lamellae in Fig. 4(D) are antiphase boundaries and the regions between them are APDs related by displacement vector 1/2(a + b). High resolution TEM imaging down the b axis showed numerous places where the (100) lattice fringes are offset by a/2 (Bozhilov et al., 1999), providing strong support for this interpretation. Careful examination of the interfaces between lamellae and host on all scales showed them to be coherent; they lack misfit dislocations and exhibit crystallographically-controlled steps accompanied by very small adjustments of the lattice fringes as they cross the boundary. EDS analysis in the TEM gave the following compositions: diopside host — (Ca₀.₉₄Na₀.₀₆Mg₀.₈₄Fe₀.₀₈Cr₀.₀₃Al₀.₀₄Ti₀.₀₁)Si₂₂O₆; clinenoestatite lamellae — (Mg₁.₅₇Fe₀.₂₅Na₀.₀₃Cr₀.₀₀₁Al₀.₀₂Ca₀.₁₄)Si₂₂O₆.

Of the five different polymorphs known for low-calcium pyroxenes, three are monoclinic. Two have space group C2/c, High Temperature clinenoestatite (HTclen) and High Pressure clinenoestatite (HPclen); the remaining monoclinic phase, low clinenoestatite (Lclen), has space group P2₁/c and is stable at Earth’s surface. The electron-diffraction information (Fig. 4(B)) clearly demonstrates that these lamellae have space group P2₁/c, but the antiphase domains imply that this phase has formed by inversion of a pre-existing phase with space group C2/c. Neither HTclen nor HPclen can be quenched to room temperature and pressure; both undergo spontaneous and rapid displacive transformations to Lclen (Angel et al., 1992; Angel and
Fig. 4. (A) Transmission electron micrograph of clinoenstatite exsolution lamellae in diopside from Alpe Arami lherzolite imaged approximately down the b axis; interfaces between host and lamellae are coherent. (B) Selected area electron diffraction (SAED) shows [010] zone axis patterns of both the C-centered diopside (strong spots) and the P2₁/c clinoenstatite (weak spots). Note that the spots with \( h+k = \text{even} \) are present for both pyroxenes whereas the spots with \( h+k = \text{odd} \) are absent for the diopside host; the angles between the crystal axes of the two phases are about \( 2^\circ \) for the \( c^* \) axes (labeled) and \( 1^\circ \) for the \( a^* \) axes. (C,D) Pair of dark field images of a single lamella with operating reflections \( h+k = \text{even} \) (C) and \( h+k = \text{odd} \) (D). Mottled contrast within the lamella in (C) is like that seen in bright field illumination (A) and is due to elastic lattice distortions. Note that neither the shape nor size of the bright areas in (D) correlates with the strain-induced contrast features in (C). The diopside host appears bright in (C) because of the overlap of diopside and \( h+k = \text{even} \) clinoenstatite reflections. (C and D from Bozhilov et al., 1999).
Hugh-Jones, 1994; Hugh-Jones et al., 1994, 1996). Thus, either HTclen or HPclen could have been the precursor. Bozhilov et al. (1999) discuss all of the possibilities of origin of these lamellae, including the possibility that orthoenstatite (Oren) could have formed first, followed by inversion or shearing to form Lclen (Coe and Kirby, 1975). The only precursor consistent with all of the crystallographic and geologic evidence is HPclen.

Previous thermobarometric and geochronological investigations of the Alpe Arami massif have documented in detail the $P$–$T$–$t$ path this peridotite has followed to the surface from ~160 km, 1400 K, the maximum conditions recorded by the most sluggish ion-exchange reactions in the system (Fig. 5). According to these analyses, the massif has not experienced temperatures in excess of 1100 K at pressures less than 2 GPa (60 km depth) (Brenker and Brey, 1997) — conditions much too cold to have accessed the HTclem stability field (e.g., Cameron and Papke, 1981; Gasparik, 1990). Indeed, it is clear that if the massif had reached the HTclem stability field, even if anhydrous, it would have been well above its solidus and

![Fig. 5](image_url)

Fig. 5. Constraints on exhumation path of the Alpe Arami peridotite. Stippled and solid circles represent maximum metamorphic conditions determined by Ernst (1978) and Medaris and Carswell (1990), respectively, using thermobarometry. Exhumation curve from 5 GPa to surface (boxes connected by solid arrows) from Brenker and Brey (1997) and references therein. ($Mg_{0.9}Fe_{0.1})SiO_3$ phase diagram from Woodland and Angel (1997); pyroxene phase fields labeled with the space group of the stable phase. Experimental data points from Fig. 3 also shown (squares represent initial conditions of two-stage experiments). Dashed arrow below 5 GPa is a possible exhumation path consistent with data currently available. Olivine phase fields from Akaogi et al. (1989).
would have left a dominant igneous signature in the microstructure. Moreover, such a high temperature path would have led the garnets to break down into pyroxene + spinel symplectites, the cation-exchange reactions upon which thermobarometry is based would not have closed until very shallow depths, and mineralogical and microstructural memory of the Alpine metamorphism would have been seriously modified or erased. In contrast to this scenario, if the massif came from depths greater than the Oren stability field (Woodland and Angel, 1997) along a trajectory consistent with the exhumation curve of Brenker and Brey (1997), exsolution from diopside above the Oren/HPcLEN phase boundary would have produced HPcLEN and would be consistent with all other observations concerning this peridotite.

We attribute metastable preservation of HPcLEN (rather than inversion to Oren) primarily to the low energy of the coherent boundaries between the HPcLEN lamellae and their diopside host, and to the reconstructive nature of the C2/c → ortho transformation. In addition, the low-energy boundary between diopside and Oren is parallel to (100), far from (401). Thus, given the surface-to-volume ratio in these lamellae, conversion to Oren in situ would have provided little or no savings in free energy, and reconstitution of the lamellae parallel to (100) would have required a large activation energy compared to the free energy reduction to be attained. At some undetermined lower T and P, the unquenchable C2/c → P21/c reaction ran within the HPcLEN lamellae, yielding the observed antiphase boundaries. That transformation could occur at any point beyond the metastable extension of the HPcLEN/LcLEN boundary (Fig. 5). The strong strain contrast that exists inside the lamellae is distinctly different from the lack of such effects in volcanic rocks (e.g., Nord et al., 1976; Robinson et al., 1977). We attribute this difference at least in part to the significant volumetric expansion (~3%) between HPcLEN and the LcLEN inversion product, whereas diopside parameters vary smoothly with pressure and temperature. The much larger scale bending in the diopside due to strain concentrated at the lamellae tips also is probably a consequence of the volumetric expansion upon transformation to LcLEN.

It is worthy to note here that thus far clinoenstatite lamellae have been found only in diopside crystals of Alpe Arami lherzolite and only in those that are spatially associated with garnets. Similarly, those diopsides not associated with garnet do not show the large elastic strains present in those arranged around or in garnet. These observations suggest that, despite the observation that all diopsides currently have the same composition, the matrix diopsides grew or recrystallized at lower pressures. A corollary to this observation is that the detailed chemistry of these phases cannot be used to calculate conditions of origin of the lamellae because cation exchange amongst all of the phases must have continued after formation of the lamellae.

We conclude from all the crystallographic and geologic information that the only scenario consistent with the observations is that HPcLEN exsolved from diopside at very high pressure. The position of the Oren/HPcLEN boundary has been determined for both Mg- and Fe-end members (Woodland and Angel, 1997; Woodland, 1998), and it has been shown that Ca has little effect on the position of the boundary (Herzberg and Gasparik, 1991; Woodland, 1998). Therefore, we adopt the phase diagram for (Mg0.9Fe0.1)SiO3 (Woodland and Angel, 1997), which represents a typical mantle composition (Fig. 5). Because it is clear that cations remained mobile during and after the conditions of exsolution of HPcLEN, it is not possible to estimate the conditions of exsolution within the HPcLEN field. However, the composition of
diopside-enstatite solid solutions shown in Fig. 4 of Herzberg and Gasparik (1991), suggests an upwelling path for Alpe Arami with a slope as shown in Fig. 5. Other paths are possible, of course; higher temperature paths would also fit the observations readily, but paths starting at lower temperatures imply an increase in solubility of enstatite components in diopside along parts of the path in which temperature is increasing (Herzberg and Gasparik, 1991).

Also shown in Fig. 5 are the conditions of the experiments of Fig. 3, where we have determined experimentally the solubility of TiO$_2$ in olivine. At 12 GPa, all experiments showed solubility $\geq 0.6$ wt%; at 10 GPa, conditions of solubility $> 0.6$ wt% lie to the right of the data. Thus, exhumation curves simultaneously consistent with exsolution of HPclen from diopside and the thermobarometric data of others also imply high solubility of TiO$_2$ in olivine (see Fig. 3 for $P/T$-dependence of TiO$_2$ in olivine). Potential exhumation paths with slopes significantly steeper than that of the dashed arrow in Fig. 5 would not be consistent with solubility trends of MgSiO$_3$ in diopside (Herzberg and Gasparik, 1991).

5. Omphacite with quartz precipitates in Alpe Arami eclogite

We have recently discovered abundant quartz precipitates in omphacite from eclogite bordering the Alpe Arami massif (Fig. 6; Dobrzhinetikaya et al., 1999). To our knowledge, evidence for such previously Si-rich nonstoichiometric pyroxenes has not been described from

![Fig. 6. Optical micrograph, plane-polarized light, showing a ‘cloud’ of SiO$_2$ precipitates in Alpe Arami omphacite. The precipitates are quartz rods which are preserved only in the central portions of crystals, a common observation for exsolution in materials of all kinds; excess silica from the remainder of the crystal escaped to grain boundaries where it participated in retrograde mineral reactions.](image-url)
Alpe Arami nor from elsewhere in the Alps. However, they are well-known from eclogite xenoliths in kimberlite pipes (e.g., Smyth, 1980), and similar omphacite has been reported from eclogites of the Sulu area in China (Liou et al., 1988), where associated garnet peridotite carries thermobarometric evidence of pressures of 6 GPa (Yang et al., 1993). The significance of such omphacite as an indicator of very high pressures is not yet fully understood, but suggests pressures greater than those yielded by thermobarometry. Pyroxene with composition \( \text{Na(Mg}_{0.5}\text{Si}_{0.5})\text{Si}_{2}\text{O}_{6} \) has been synthesized at 15 GPa, 1900 K (Angel et al., 1988).

6. Discussion

The experimental data presented here indicate that dry lherzolite saturated with TiO\(_2\) would exhibit high solubility of that component in olivine. However, a critical and still unresolved conflict between two methods of measurement remains concerning the ilmenite content of the older generation of Alpe Arami olivine. Image analysis of 1 vol\% FeTiO\(_3\) crystals implies an original TiO\(_2\) content \( \geq 0.6 \) wt\% (Dobrzhinetskaya et al., 1996; Green et al., 1997a). Such measurement of relative abundances of intergrown phases is the usual method for estimating original bulk compositions and has been applied in other cases of implied extreme depth of origin of mantle rocks (Haggerty and Sautter, 1990; Sautter et al., 1991; van Roermund and Drury, 1999; van Roermund et al., 1998, 1999). However, Hacker et al. (1997), despite confirming our measurement of 1 vol\% ilmenite precipitates in our Alpe Arami olivine crystals, argued that broad-beam microprobe analysis of one of our crystals indicated a much lower TiO\(_2\) content. They also measured several other crystals from their own collections of titanate-rod-bearing olivine by this technique; although the crystals analyzed contained abundances and sizes of ilmenite rods that varied widely, their microprobe technique yielded the same low TiO\(_2\) content in all crystals. Hacker et al. (1997) provided no explanation for why they found image analysis and broad-beam probe analysis to yield uncorrelated results; these differences remain unresolved (Green et al., 1997a). At this point, therefore, the experimental measurements demonstrate that in a TiO\(_2\)-saturated five-phase lherzolite (olivine, garnet, diopside, high-pressure clinoenstatite, ilmenite) at \( P > 10 \) GPa, olivine will dissolve in excess of 0.6 wt\% TiO\(_2\) and that it will not dissolve such high amounts at lower pressures. Whether or not Alpe Arami olivine contains that much TiO\(_2\) remains controversial. If the image analysis is correct, then the sum of all data currently known suggests origin of the Alpe Arami peridotite at \( P \geq 10 \) GPa, \( T \geq 1600 \) K, depth \( \geq 300 \) km. If the broad-beam microprobe data are correct, then the Alpe Arami peridotite had to have been greatly undersaturated in FeTiO\(_3\) under conditions of exsolution of high-pressure clinoenstatite and no minimum depth can be estimated.

Relevant to this controversy is that in experiments in which we varied conditions to first produce a high concentration of dissolved TiO\(_2\) in olivine and then changed the conditions to approach the equilibrium solubility from higher concentrations, disks of ilmenite were produced lying in (100) of olivine, as was found earlier for natural precipitation of ilmenite (Mosely, 1981; Drury and van Roermund, 1988). These results support our previous speculation that the rod shape of FeTiO\(_3\) precipitates in Alpe Arami olivine is unlikely to be due to direct precipitation of ilmenite. Thus, origin of the rod morphology of Alpe Arami titanate precipitates in olivine remains unexplained.
7. Very deep origin of subduction-zone peridotites

Subduction-zone peridotites are in general garnet peridotites. That in itself requires a depth of origin in excess of 60–80 km. It follows, therefore, that at minimum all of these rocks were exhumed from such depths, that the processes that operated to bring them to the surface extend to at least that depth, and that such processes have operated in many subduction zones of many time periods. Moreover, the exhumation processes must have occurred at sufficiently low temperature or sufficiently rapidly that the garnets only partially reacted to the spinel facies. Beyond these arguments based purely on the presence of garnet, thermobarometric calculations based on various mineral combinations in individual garnet peridotites record pressures of 3–3.5 GPa in the Western Gneiss Region of Norway, 4–5 GPa for the Alpe Arami peridotite in Switzerland (Ernst, 1978, Ernst, 1981; Medaris and Carswell, 1990; Becker, 1993; Brenker and Brey, 1997), and in excess of 6 GPa for the Sulu terrane of Eastern China (Yang et al., 1993). Specific reservations can be raised concerning individual measurements, but the general agreement of these measurements and their basis in many and diverse experimental studies makes it highly unlikely that such measurements are seriously in error. It also must be accepted that the thermobarometric measurements reflect minimum depths of origin because they represent closure of cation-exchange reactions as temperatures fall (for further discussion, see Brenker and Brey, 1997). Thus, there is every reason to expect that these rocks have had a history involving greater depths. Indeed, plate tectonics tells us that the mantle convects in the solid state, hence it is to be expected that most, if not all, peridotites have experienced long and complicated histories involving circulation to great depths, with some likely to have reached the core-mantle boundary (2900 km depth). Some key questions are, therefore: (1) What evidence might there be of such history? (2) What conditions would be required to retain such evidence? (3) When was such evidence imprinted in the rocks — just before exhumation? — long before exhumation? (4) If long before exhumation, how and where was the evidence preserved until exhumation?

Because cation-exchange reactions can only freeze in conditions below some specific (pressure-dependent) temperature, thermobarometry is limited to closure of the most refractory reaction in a given rock type. Peridotites are the most refractory rocks, hence they present the greatest opportunity to preserve evidence of great depth. It is not surprising, therefore, that it is this rock type that records the greatest depths in any given terrane. These systematics should also cause us to question whether the lower pressures recorded in basic and acidic rocks of the same or related terranes is because they are unrelated to the peridotites (and have become associated with them during exhumation), or whether these rocks show lower pressures simply because thermobarometry is unable to record higher temperatures and, therefore, higher pressures in these rock types. Quartz rods in omphacite are consistent with the latter interpretation. If the chemical composition of currently existing phases cannot record depths beyond those at which the closure temperatures of their cation-exchange reactions is reached, then only metastable preservation of high-pressure phases or indirect methods of recovering earlier compositions are available. An excellent example of the former is the discovery of diamond and coesite in ultra-high-pressure metamorphic terranes. Such discoveries were questioned seriously early-on in the evolution of this problem, simply because the community was unprepared to accept that such high-pressure rocks could reach the surface by tectonic
means. A good example is the evolution of acceptance of the depth of origin of the rocks in the Western Gneiss Region of Norway. The early discovery of coesite by Smith (1984) was not seriously entertained for many years. Discovery of microdiamonds in this same terrane (Dobrzhinetskaya et al., 1993, 1995) was greeted by some with suspicion of contamination rather than confirmation of Smith’s observations. Only discovery of a large volume of coesite-bearing eclogite (Wain, 1997) has finally put this question to rest. History repeats itself. Another terrane of microdiamonds in quartzofeldspathic gneisses has recently been discovered in the Variscan Erzgebirge of Saxony, Germany (Massonne, 1998). The general reaction of the community is again disbelief. However, Professor Massonne has provided us materials from the diamondiferous locality and we have confirmed microdiamonds in situ in samples for which no diamonds have been used in the preparation process (Dobrzhinetskaya, Green, and Bozhilov, unpublished results).

Another method by which rocks can have memory of earlier higher T or P conditions is by microstructural preservation of phase geometries that point either to the presence of previous compositions of phases that adjusted to changing conditions by precipitation reactions or to previous phases that became unstable and broke down in recognizable ways. The example of preservation of precipitates of one phase in another has been used in this way for many years. For example, recapture of the conditions of higher temperature and/or pressure conditions by recalculating compositions from exsolution lamellae of one pyroxene in another or one feldspar in another is a common petrological tool. Readily recognizable examples of break-down reactions are symplectites, such as those of spinel + pyroxene after garnet in peridotites, plagioclase + quartz after omphacite in eclogites, olivine + ilmenite after titanoclinohumite, and many others. Thus, these microstructural tools are not new, nor is acceptance of their validity. Their application to infer very high pressures is controversial, therefore, either because the experimental basis upon which to base such inferences is absent or because of reluctance to believe that exhumation from hundreds of kilometers is possible. Continuing the example of the Norwegian peridotites and eclogites, depths of 100 km are now considered acceptable for this terrane because of discovery of abundant coesite. However, van Roermund and Drury (1998), van Roermund et al. (1998, 1999) have discovered exsolution of pyroxenes from garnet in peridotite, implying minimum depths of order 200 km (Haggerty and Sautter, 1990; Sautter et al., 1991). Thus, the Alpe Arami peridotite is no longer the only subduction-zone peridotite with microstructural evidence suggesting depths significantly greater than given by thermobarometry.

These results and the validation of thermobarometry by continuing experimental work in progressively more complicated chemical systems and the internal consistency of their application to a wide variety of geological problems indicates that exhumation of at least peridotites from depths in excess of 200 km must be accepted. The fact that these rocks are all found in subduction terranes further requires that at least the last stage of this exhumation occurred in a collision zone. Moreover, if the entire exhumation did not occur in the subduction zone in which the rocks are now found, at some earlier time the peridotites must have been carried up to the depths from which they were eventually exhumed to the surface and stored there under conditions capable of preserving the microstructural evidence of great depth for geologically significant time periods. One would also expect that some part of the chemical system would have re-equilibrated to the ‘storage’ conditions because mantle
xenoliths demonstrate the ability of peridotites to equilibrate under a wide variety of conditions. Moreover, if there has been a period of such storage, either the rocks currently surrounding the peridotites also must have been stored under those conditions, or the peridotites must have been intercalated with their surroundings during the final exhumation. That is, rejection of the idea that subduction has extended to the depths recorded by the exhumed rocks implies two or three complicated geological events for which direct evidence must be sought. It also means that this complicated process must be rather common given that to avoid the idea of direct exhumation from very great depths now requires this inference for at least three continental collision terranes, an implication itself worthy of considerable thought in terms of plate-tectonic processes involved in continental collisions. Is it more difficult to conceive (or model) exhumation from 150 km than 100 km? Is it more difficult to imagine a mechanism that could be responsible for exhumation from 200 km? Or 300 km? To our understanding, only buoyancy forces can achieve exhumation from such depths where rocks are hot and ductile. Thus, arrival at the surface of a peridotite from great depths requires that a larger body of lower density was subducted to depths in excess of the last resting place of the peridotite and rafted it to crustal levels.

References


