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Revised Thermobarometry of Alpe Arami and other Garnet Peridotites from the Central Alps

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*The four, currently best constrained, independent thermobarometers for garnet peridotites, namely Taylor's (*Neues Jahrbuch für Mineralogie, Abhandlungen ***172,*** *381–408, 1998) pyroxene solvus and Krogh's (*Contributions to Mineralogy and Petrology ***99,*** *44–48, 1988) clinopyroxene–garnet Fe–Mg exchange thermometers, and Taylor's (1998) Al-in-orthopyroxene and Nimis & Taylor's (*Contributions to Mineralogy and Petrology, ***139,*** *541–544, 2000) Cr-in-clinopyroxene barometers, have been applied to garnet lherzolites from the Central Alps. Analyses from the literature, as well as new in-house analyses, all pertaining to core compositions of first-generation, garnet lherzolite minerals, have been selected for thermobarometric calculations. The* P–T *data obtained for the three known garnet lherzolite occurrences in the Central Alps are tightly constrained, consistent with one another, and summarized as follows: Alpe Arami, 3·2 GPa and 840°C; Monte Duria, 3·0 GPa and 830°C; Cima di Gagnone, 3·0 GPa and 740°C. These values are consistent with experimental data on pargasite stability and composition in peridotitic systems. Our* P–T *estimates, along with microstructural and field geological observations, indicate that the garnet lherzolite parageneses form part of the prograde, Alpine, high-pressure metamorphic sequence of the Adula–Cima Lunga unit. Thermobarometry shows that the garnet lherzolites reached a maximum depth of subduction of ~100 km, leaving little room for an extraordinary high-pressure, Alpine metamorphism at Alpe Arami. The very high pressure estimates obtained by some other workers are ascribed to inaccuracies in thermometric evaluations and to the strong temperature dependence of the Al-in-Opx barometer. Uncertainties in the determination of garnet–olivine equilibrium compositions, combined with the small sensitivity of garnet–olivine chemical exchanges to temperature variations, and unreliability of garnet–orthopyroxene thermometry at low-temperature conditions may account for the observed inconsistencies. A discrepancy between thermobarometric data for the garnet lherzolites and the associated eclogites is non-existent in our example, and in other localities it may also be a thermobarometric artefact.*

KEY WORDS: *Adula–Cima Lunga nappe; Central Alps; high-pressure metamorphism; garnet peridotites; thermobarometry*

INTRODUCTION

The detection of metamorphic rocks of increasingly deeper origin (Chopin, 1984; Smith, 1984; Sobolev & Shatsky, 1990; Xu *et al.*, 1992; Van Roermund & Drury, 1998) is of considerable importance for the understanding of geodynamic processes. Amongst the lithologies that permit monitoring of deep pressure–temperature paths, garnet lherzolites have received a great deal of attention. These rocks have a relatively simple mineralogical composition and their stability has been investigated in numerous experiments. Perhaps the most famous occurrence of garnet lherzolite in a metamorphic terrain is that at Alpe Arami, Central Swiss Alps. For this garnet lherzolite, an exhumation after subduction to extraordinary depths of 300–650 km, i.e. to the mantle transition zone or

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deeper, has recently been postulated by Dobrzhinetskaya *et al.* (1996). This hypothesis has received support from the thermobarometric data of Brenker & Brey (1997), who estimated minimum P–T conditions of 5 GPa and 1120°C for Alpe Arami, and from the finding of lamellae of possible former $C2/c$ clinoenstatite in diopside (Bozhilov *et al.*, 1999). The arguments put forth by Dobrzhinetskaya *et al.* and Bozhilov *et al.* have, however, been strongly disputed by Risold *et al.* (1996, 1997), Hacker *et al.* (1997), Ulmer *et al.* (1998), Arlt *et al.* (2000), and Trommsdorff *et al.* (2000), leaving the very deep origin of Alpe Arami open to question. It is the purpose of this paper to present new thermobarometric data for the three garnet lherzolite occurrences in the Central Alps—Cima di Gagnone, Monte Duria and Alpe Arami—which contribute to solving the ultrahigh-pressure enigma of the Alpe Arami rocks. It will be shown that at all three localities maximum Alpine metamorphic pressures were of the order of 3 GPa, which is in serious conflict with the 10–24 GPa proposed by Dobrzhinetskaya *et al.* (1996) for Alpe Arami.

GEOLOGICAL OUTLINE

The known outcrops of garnet lherzolite in the Central Alps all occur in the same tectonic zone, namely the upper part of the Adula–Cima Lunga unit (Fig. 1). Palaeogeographically, this unit has been considered part of the former European continental margin (Schmid *et al.*, 1990) and thus it forms the uppermost nappe of the lower Penninic system. Heinrich (1982, 1986) mapped mineral assemblages in mafic and pelitic lithologies of the Adula nappe and established isograds of a regional high-pressure metamorphism pre-dating the classic, Lepontine isograd belt (Trommsdorff, 1966; Jäger *et al.*, 1967) of the Central Alps. The peak conditions of the high-pressure metamorphism increase from north to south from 1 GPa and 500°C to over 2·5 GPa and 800°C, based on thermobarometry of eclogites, metapelites and meta-ophicalcite rocks (Heinrich, 1982, 1986; Partzsch, 1996; Meyre *et al.*, 1997, 1999; Pfiffner, 1999).

Garnet lherzolites were found in the highest P–T region of the Adula–Cima Lunga nappe at three localities. These are, from east to west (Fig. 1): Monte Duria (MD; Fumasoli 1974), Alpe Arami (AA; Grubenmann, 1908; Möckel 1969), and Cima di Gagnone (CdG; Evans & Trommsdorff, 1978). At all three localities, the garnet lherzolites show a more or less pronounced layering determined by variations in content of olivine and pyroxene. At MD, garnet lherzolites occur in >20 individual ultramafic bodies, which form 10–100 m boudins within migmatitic gneisses. Garnet lherzolites and gneisses are folded around a steeply plunging megafold having an amplitude of at least 4 km. At AA, garnet lherzolites form the core of a 1 km × 400 m chlorite peridotite boudin surrounded by steeply south-dipping migmatitic gneisses. A discontinuous layer of eclogitic rocks separates the peridotite body from the country gneisses. At CdG, garnet lherzolite occurs in one of numerous ultramafic lenses of tens to hundreds of metres in size, surrounded by pelitic and semipelitic, in part migmatitic, gneisses, kyanite eclogites, marbles, and meta-ophicalcite rocks (Pfiffner & Trommsdorff, 1998). The peridotites show transitions to eclogite. Within the ultramafic rocks, the presence of metarodingite boudins, interpreted as former mid-ocean ridge basalt (MORB) dykes, testifies to an early serpentinite stage of the metaperidotites (Evans *et al.*, 1979, 1981). The ultramafic–mafic–carbonate suite at CdG has been interpreted as derived from an ocean basin near a continental margin (Pfiffner & Trommsdorff, 1988), with the ultramafic rocks representing former subcontinental mantle that had been exhumed during oceanic rifting.

The chemical compositions of the garnet lherzolites from the Central Alps (O'Hara & Mercy, 1966; Fumasoli, 1974; Rost *et al.*, 1974; Ernst, 1978; Evans & Trommsdorff, 1978; Pfiffner, 1999) are all remarkably similar and close to that of fertile mantle, which is typical for the subcontinental lithosphere in the Alpine realm and in Liguria (Nicolas & Jackson, 1972; Piccardo *et al.*, 1990; Menzies & Dupuy, 1991; Müntener 1997). Isotopic geochemical investigations of minerals from AA and CdG garnet lherzolites yielded consistent, Eocene ages of ∼40 Ma, based on garnet–clinopyroxene–whole-rock Sm–Nd isochrons (Becker, 1993), and of 43–35 Ma in zircon, based on the U–Pb method and SHRIMP analysis (Gebauer *et al.*, 1992; Gebauer, 1996). These ages are consistent with the modern dating of late-Eocene high-pressure metamorphism in the Western Alps (Froitzheim *et al.*, 1996; Gebauer *et al.*, 1997). In agreement with the prograde character of the eclogite sequence mapped in the Adula–Cima Lunga unit by Heinrich (1982, 1986), most of the garnet peridotites of the Central Alps show evidence of prograde metamorphism, as reported by Evans & Trommsdorff (1978) for CdG and as further documented in this paper.

PETROGRAPHY: POIKILOBLASTIC AND PORPHYROCLASTIC GARNET LHERZOLITES

Detailed petrographic descriptions of garnet lherzolites from MD, AA, and CdG have been given by Fumasoli (1974), Möckel (1969), and Evans & Trommsdorff (1978), respectively. Additional information regarding the textures is given here. The garnet lherzolites in the Central Alps exhibit two main textural types. The first type, a poikiloblastic garnet peridotite, occurs at CdG and, locally, at MD. It is weakly foliated, sometimes folded, with

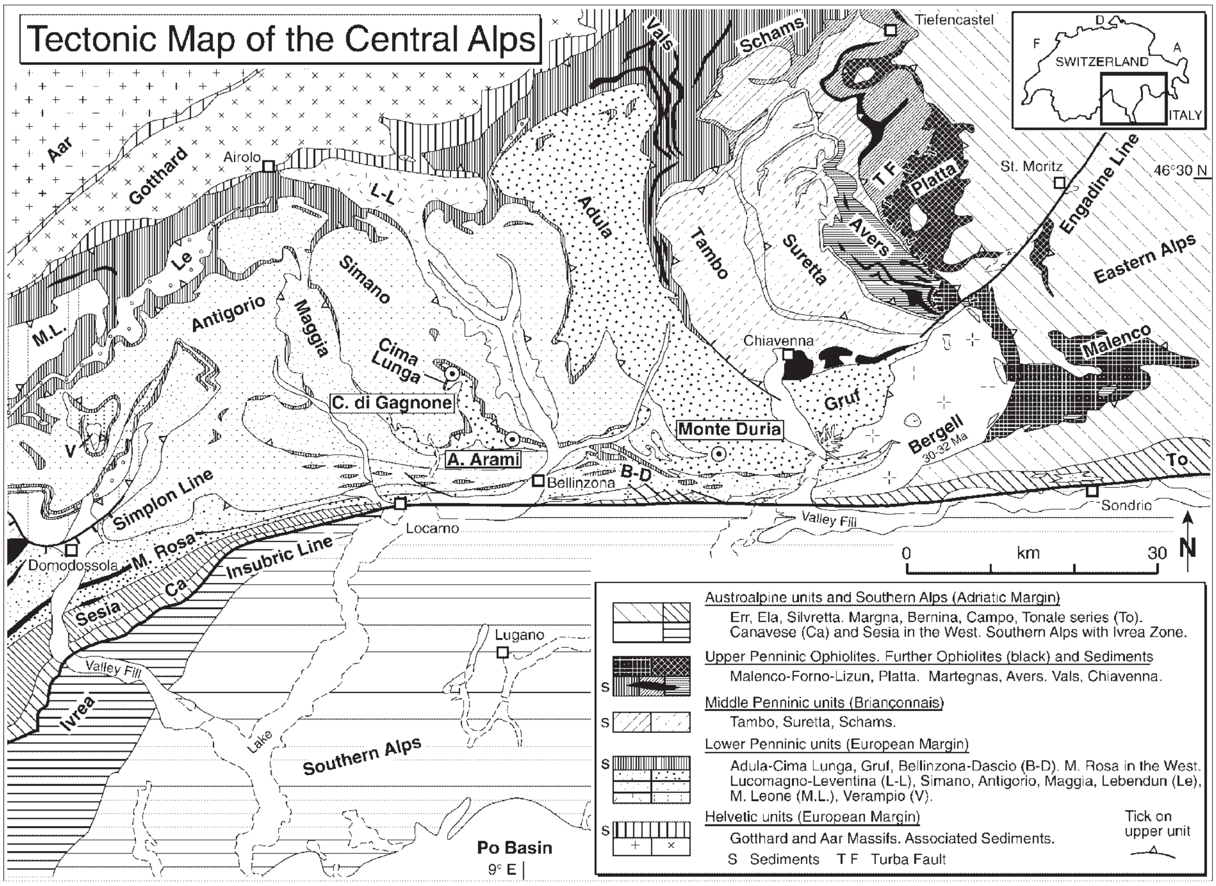

Fig. 1. Tectonic map of the Central Alps compiled by P. Nievergelt. [Modified from Spicher (1980).]

completely anhedral, often elongate garnet concentrated along pyroxene-rich layers. Garnet is poikiloblastic and often stuffed with abundant, sometimes folded inclusions of orthopyroxene and, less commonly, olivine, clinopyroxene and rare, pale brownish to greenish, rounded, magnesian Ca-amphibole. The matrix around the garnets is formed of an idioblastic mosaic of olivine, orthopyroxene, clinopyroxene and Ca-amphibole with a typical grain size of 0·5–1 mm. Pseudomorphs of olivine with inclusions of wormy ilmenite after titanian clinohumite (Evans & Trommsdorff, 1978) are commonly found along more pyroxenitic layers. Relicts of titanian clinohumite are rare. At CdG, garnet has overgrown pre-existing, isoclinal folds involving all the matrix minerals, including Ca-amphibole and pseudomorphs after titanian clinohumite (Pfiffner & Trommsdorff, 1998). The subrounded Ca-amphibole grains enclosed in garnet have a distinctly higher K_2O content (0·7 wt %; Table 1) than those in the matrix (0·15 wt %), probably as a result of residual K enrichment of Ca-amphibole during its partial resorption and growth of garnet. Thus, there is a large body of evidence that, in the poikiloblastic type, garnet formed during prograde metamorphism. Evans & Trommsdorff (1978) interpreted the poikiloblastic peridotites from CdG as 'subduction zone garnet peridotites' derived from a partially serpentinized, hydrous protolith. At CdG, prograde chlorite–amphibole peridotites are commonly found to be isofacial with the garnet peridotites. The simultaneous occurrence of both rock types can be explained in terms of variable bulk rock composition and/or H_2O activity (Trommsdorff, 1990; Pfiffner, 1999).

The second textural type, a porphyroclastic garnet lherzolite, dominates at AA and MD. It is characterized by more or less equant, anhedral porphyroblasts of garnet up to 1 cm in diameter in a porphyroclastic matrix with a mortar texture. Garnet is relatively free of inclusions and rounded, but in some cases it developed embayed grain boundaries around olivine and pyroxene porphyroclasts. At AA, primary inclusions of spinel are rarely found that pre-date garnet (Möckel, 1969; pp. 76 and 85). Here we report the finding of an inclusion of K-enriched Ca-amphibole, associated with a brown spinel with Cr/(Cr + Al) = 0·24. Despite some retrograde

Table 1: Compositions of amphiboles in selected samples from Alpe Arami and Cima di Gagnone

	Alpe Arami 9-5-2					Cima di Gagnone	
	incl. in Grt (core)	incl. in Grt (rim)	matrix* (near kel.)	matrix* (interm.)	matrix* (rim)	160-96-2 inclusion in Grt	160-4-8 matrix†
SiO_2	44·07	44·03	44·53	46·05	45·00	47·38	46·50
TiO_2	0·60	0·51	0·27	0·41	0·53	0·53	0·46
Al_2O_3	14·49	14·93	16·37	13·51	12·67	11·73	10·86
Cr_2O_3	1·07	1·12	1·28	1·29	1·24	0·39	1·37
FeO_{tot}	2·82	2·71	4·10	4·17	4·30	3·46	3·19
MnO	0·03	0·08	0·01	0·07	0·04	0·08	0·04
MgO	19·08	18·98	17·65	18·69	20·32	19·93	19·40
CaO	11·13	11·84	11·92	11·55	10·87	11·98	12·00
Na_2O	3·35	3·30	3·34	3·02	2·63	1·91	2·15
K_2O	0·25	0·17	0·05	0·08	0·18	0·68	0·15
Total	96·89	97·67	99·52	98·84	97·78	98·05	96·12
Cations per 23 oxygens							
Si	6·25	6·21	6·18	6·42	6·35	6·61	6·64
Ti	0·06	0·05	0·03	0·04	0·06	0·06	0·05
Al	2·42	2·48	2·68	2·22	2·11	1·93	1·83
Cr	0·12	0·13	0·14	0·14	0·14	0·04	0·16
Fe	0·33	0·32	0·48	0·49	0·51	0·40	0·38
Mn	0·00	0·01	0·00	0·01	0·01	0·01	0·01
Mg	4·04	3·99	3·65	3·89	4·28	4·15	4·13
Ca	1·69	1·79	1·77	1·73	1·64	1·79	1·84
Na	0·92	0·90	0·90	0·82	0·72	0·52	0·60
K	0·05	0·03	0·01	0·01	0·03	0·12	0·03

*Zoned amphibole growing at the outer margin of a partially kelyphitized garnet.
†After Evans & Trommsdorff (1978).

re-equilibration at the inclusion–host interface, testified by depletion in garnet of Cr and in amphibole of alkalis, the amphibole cores still preserve a high-K composition (K_2O = 0·25 wt %; Table 1). By analogy with those found at CdG, this K-enriched Ca-amphibole inclusion is believed to pre-date garnet. At MD, in several cases garnet overgrows pre-existing, folded strings of brown spinel and Ca-amphibole grains. These observations indicate that garnet, probably at AA and definitely at MD, was prograde and developed after an amphibole + spinel peridotite stage. Consistent with a prograde history, titanian clinohumite, partially replaced by olivine + ilmenite symplectites, has been found in a loose block of garnet peridotite near AA (Möckel 1969) and in garnet pyroxenite veins crosscutting the main garnet peridotite at MD (Heinrich, 1983). The matrix is dominated by two generations of olivine, a porphyroclastic generation of large, irregular crystals up to 2 mm in diameter and a recrystallized generation of small, subhedral olivine that forms the mortar between the porphyroclasts. Less abundant orthopyroxene is up to 2 mm in size and locally kinked. Chromian clinopyroxene, up to 5 mm, is frequently concentrated near the garnet (Dobrzhinetskaya *et al.*, 1996), which is in calcium-rich domains of the rock. In many instances it contains exsolution lamellae of pargasitic to edenitic amphibole (Yamaguchi *et al.*, 1978) and, at AA, of $P2_1/c$ clinoenstatite. The latter contains antiphase domains adjacent to diopside, which may reflect exsolution of a former $C2/c$, high-pressure clinoenstatite polymorph (Bozhilov *et al.*, 1999). This has been offered as a proof for original pressures exceeding 8 GPa, but the strong effect of foreign cations on the stability of $MgSiO_3$ polymorphs (Arlt *et al.*, 2000) renders this argument inconclusive. On the basis of existing experimental data, Arlt *et al.* concluded that 'it remains highly speculative to suggest a very high pressure origin of Alpe Arami peridotite'. According to these workers 'the possibility of a metastable formation of a high-*T* *C*2/*c*-phase within the orthopyroxene stability field must also be considered and appears to be the easiest explanation'. A second, smaller and less chromian generation of clinopyroxene forms part of the matrix. At AA, a textural type transitional to the porphyroblastic type characterized by amoeboid garnets elongated parallel to the main foliation is also rarely found.

At all three occurrences (i.e. CdG, AA, and MD), the garnet peridotite assemblages underwent partial to complete local retrogression under amphibolite-facies conditions. The extent of the retrogression was dependent on the availability of H_2O and involved partial kelyphitization (spinel + amphibole) of garnet, growth of sapphirine (at CdG and MD) and of various, progressively less sodic, low-K (K_2O <0·2 wt %) Ca-amphibole generations (Table 1) and, finally, chloritization of garnet. The various retrograde stages were described in detail by Möckel (1969), Fumasoli (1974), Evans & Trommsdorff (1978), and Trommsdorff (1990), and are not further considered in this paper.

MINERAL COMPOSITIONS

Chemical analyses of core and rim compositions of garnet (Grt), orthopyroxene (Opx), clinopyroxene (Cpx) and olivine (Ol) from CdG, AA and MD were given by Evans & Trommsdorff (1978). Numerous additional microprobe analyses have been published for samples from AA (Ernst, 1978; Yamaguchi *et al.*, 1978; Hacker *et al.*, 1997; Brenker & Brey, 1997; Paquin *et al.*, 1999). These data indicate pronounced mineral zonations towards the outer parts of the first-generation porphyroclasts and well-developed plateaux in their inner parts. None the less, both garnets and olivines tend to show conspicuous grain-to-grain compositional variations in terms of Fe/Mg ratios and, for the garnets, Ca and Cr contents (Evans & Trommsdorff, 1978), indicating the existence of distinct compositional domains at the thin-section scale. Trace-element profiles in garnets, including W-shaped Cr patterns [with Cr contents in the range 0·075–0·085 atoms per formula unit (p.f.u.)] and approximately bell-shaped Ni patterns (with Ni contents up to 65–70 ppm), indicate minor zonations within some garnet cores (Paquin *et al.*, 1999). These profiles have been interpreted as the result of a complex *P*–*T* history, but, alternatively, they could simply reflect local equilibrium with pre-existing Cr–Ni-rich, aluminous metamorphic minerals on which garnet porphyroblasts preferentially nucleated (e.g. spinel, chlorite). These features could be preserved because of low diffusivity of Cr and Ni in garnet. Evans & Frost (1975) have shown that spinel in metamorphic peridotites has high NiO contents (0·3–0·9 wt %) and becomes progressively more magnesian with increasing metamorphic grade. Nucleation of garnet on pre-existing spinel sites may therefore account for both the high-Cr, high-Ni cores and the large variations in Fe/Mg ratios between different garnet grains. The rims of the pyroxene porphyroclasts have compositions similar to those of the smaller, second-generation, matrix minerals (Evans & Trommsdorff, 1978; Paquin *et al.*, 1999). The relative changes in mineral compositions towards their rims reflect re-equilibration under retrograde, hydrous conditions corresponding to the kelyphite stage. For our thermobarometric study, we have selected analyses from the literature and from our own database that clearly pertain to the cores of first-generation porphyroclasts characterized by virtually flat major-element profiles (Table 2).

To minimize errors in thermobarometric calculations, we have determined by Mössbauer spectroscopy Fe^{3+}/Fe_{tot} ratios in garnet and clinopyroxene separates from two garnet lherzolites from CdG and AA [samples 160-4-8 and 9-5-2 of Evans & Trommsdorff (1978)]. Garnets from CdG and AA yielded Fe^{3+}/Fe_{tot} ratios of 0·04 and 0·03, respectively. These values are consistent with those obtained by Fett (1989) for other garnet peridotites from these two localities using the same technique. Clinopyroxenes yielded higher ferric iron fractions, resulting in Fe^{3+}/Fe_{tot} ratios of 0·05 and 0·14 for CdG and AA, respectively. Because significant zonation in minerals only involves the outermost, narrow rims (~100 μm for garnet, <50 μm for pyroxenes; Brenker & Brey, 1997, fig. 4), the values obtained can be considered as representative within uncertainties of the core portions of the grains. In view of the strict textural and mineralogical similarity between MD and AA garnet peridotites, we will assume Fe^{3+}/Fe_{tot} ratios for MD to be equal to those obtained for AA. Ferric iron contents in Opx were estimated on the basis of Cpx–Opx Fe^{3+} systematics in garnet peridotites ($Fe^{3+}{}_{Opx} = 0{\cdot}55Fe^{3+}{}_{Cpx}$; Canil & O'Neill 1996). Recalculated Fe_2O_3 and FeO wt % proportions are given in Table 2.

THERMOBAROMETRY

Numerous thermobarometers have been proposed for the evaluation of the pressures and temperatures of equilibration of garnet peridotites. Most available thermometers are based on Fe–Mg exchange between Grt and Cpx, Opx, or Ol (e.g. O'Neill & Wood, 1979; Harley, 1984; Krogh, 1988), and on the pyroxene solvus (e.g. Taylor, 1998). Aluminium exchange between Opx and Grt and Cr exchange between Cpx and Grt are strongly dependent on pressure and provide the basis for two independent barometers (e.g. Taylor, 1998; Nimis &

Table 2: Core compositions of olivine, orthopyroxene, clinopyroxene, and garnet in selected samples from Alpe Arami, Monte Duria and Cima di Gagnone

	Alpe Arami 636-1				Mt. Duria 264·3				Cima di Gagnone 160-4-8			
	Ol*	Opx*	Cpx*	Grt*	Ol†	Opx†	Cpx‡	Grt‡	Ol†	Opx†	Cpx†	Grt†
SiO_2	41·18	57·93	54·98	42·06	41·00	56·90	53·36	41·56	41·20	57·50	54·80	41·90
TiO_2	0·00	0·05	0·20	0·18	0·00	0·05	0·14	0·20	0·01	0·06	0·09	0·05
Al_2O_3	0·01	0·93	2·36	22·21	0·00	0·72	2·31	22·52	0·00	0·54	1·55	22·80
Cr_2O_3	0·00	0·29	1·25	1·82	0·00	0·13	1·09	1·44	0·00	0·10	0·33	1·18
Fe_2O_3§	n.a.	0·21	0·37	0·30	n.a.	0·23	0·39	0·32	n.a.	0·07	0·12	0·44
FeO§	9·80	5·78	2·04	8·35	9·47	6·05	2·12	8·87	9·05	5·73	1·99	9·60
MnO	0·14	0·14	0·01	0·38	0·13	0·12	0·05	0·36	0·08	0·09	0·05	0·42
MgO	48·03	34·62	15·41	20·00	49·40	35·71	16·65	20·30	49·70	36·10	16·90	19·10
NiO	0·00	0·00	0·00	0·00	0·38	0·03	0·05	0·02	0·45	0·07	0·03	0·01
CaO	0·00	0·23	21·41	4·93	n.a.	0·26	21·97	4·99	n.a.	0·28	23·00	5·23
Na_2O	0·00	0·00	1·47	0·03	0·00	0·00	1·19	0·03	0·00	0·00	0·95	0·00
K_2O	0·00	0·00	0·00	0·03	n.a.	0·00	0·00	0·00	n.a.	0·00	0·00	n.a.
Total	99·15	100·15	99·50	100·25	100·38	100·17	99·28	100·61	100·49	100·53	99·81	100·73
Cations p.f.u.												
Si	1·015	1·987	1·995	2·994	1·000	1·960	1·951	2·957	1·002	1·969	1·987	2·983
Ti	0·000	0·001	0·005	0·010	0·000	0·001	0·004	0·011	0·000	0·002	0·002	0·003
Al	0·000	0·038	0·101	1·863	0·000	0·029	0·100	1·888	0·000	0·022	0·066	1·913
Cr	0·000	0·008	0·036	0·102	0·000	0·004	0·032	0·081	0·000	0·003	0·009	0·066
Fe^{3+}	0·000	0·005	0·010	0·016	0·000	0·006	0·011	0·017	0·000	0·002	0·002	0·024
Fe^{2+}	0·202	0·166	0·062	0·497	0·193	0·174	0·065	0·528	0·184	0·164	0·060	0·572
Mn	0·003	0·004	0·000	0·023	0·003	0·004	0·002	0·021	0·002	0·003	0·002	0·025
Mg	1·765	1·770	0·833	2·122	1·796	1·833	0·907	2·153	1·802	1·843	0·913	2·027
Ni	0·000	0·000	0·000	0·000	0·007	0·001	0·001	0·001	0·009	0·002	0·001	0·001
Ca	0·000	0·008	0·832	0·376	0·000	0·010	0·861	0·380	0·000	0·010	0·893	0·399
Na	0·000	0·000	0·103	0·004	0·000	0·000	0·084	0·004	0·000	0·000	0·067	0·000
K	0·000	0·000	0·000	0·000	0·000	0·000	0·000	0·000	0·000	0·000	0·000	0·000
Total	2·985	3·989	3·978	8·008	3·999	4·022	4·016	8·041	2·998	4·018	4·005	8·013

n.a., not analysed.
*After Brenker and Brey (1997).
†After Evans & Trommsdorff (1978).
‡Analyst E. Reusser.
§Based on Mössbauer Fe^{3+}/Fe^{2+} determination (garnet and clinopyroxene) and Opx–Cpx Fe^{3+} systematics in garnet peridotites (Canil & O'Neill, 1996) (orthopyroxenes); for Alpe Arami and Monte Duria Fe^{3+}/Fe^{2+} in garnet and clinopyroxene was assumed identical to that measured on A. Arami sample 9-5-2 of Evans & Trommsdorff (1978); FeO corresponds to FeO_{tot} for olivine.

Taylor, 2000). The choice of the thermometer–barometer combination is, however, crucial for obtaining reliable P–T estimates. Many widely used calibrations may yield grossly erroneous results for some bulk compositions and/or for certain P–T ranges. The recent assessment of available thermobarometers by Taylor (1998) has shown that none of the previous formulations of the pyroxene solvus thermometer and of the Opx–Grt barometer are sufficiently robust against all possible compositional factors (see also Finnerty & Boyd, 1984, 1987; Carswell & Gibb, 1987; Brey & Köhler 1990). Amongst Grt–Cpx thermometers (Fe–Mg exchange), that of Krogh (1988) best reproduces experimental temperatures for peridotitic compositions (Brey & Köhler, 1990; Taylor, 1998). Temperatures obtained by means of this thermometer should, however, be considered with caution, because the Fe^{3+}

contents of the experimental garnets and clinopyroxenes used for its calibration are unknown and, possibly, significant (Canil & O'Neill 1996). As Fe^{3+}/Fe_{tot} ratios are typically higher in Cpx than in Grt, the use of measured Fe^{2+} atomic fractions in thermometric calculations will provide minimum temperature estimates. Amongst alternative barometers, the Cr-in-Opx barometer of Nickel (1989) was calibrated on a simple system and lacks adequate testing on natural compositions, and the Ca-in-Ol barometer of Köhler & Brey (1990) is problematic because of its strong temperature dependence, the low Ca contents, and the fast diffusion of Ca in olivine. These alternative barometers will not be used in the present study.

Evaluation of previous *P–T* estimates

Brenker & Brey (1997) calculated pressures and temperatures for a sample from Alpe Arami by using various thermometers in combination with the Al-in-Opx barometer of Brey & Köhler (1990). Their temperature estimates span a wide range, from over 1100°C to ~800°C, depending on which thermometer is used. This has been interpreted as the result of a retrograde metamorphic history starting from very high *P–T* conditions (≥5 GPa and 1120°C). The very great depth of origin (≥150 km) implied by these estimates essentially stems from the very high temperatures indicated by the O'Neill & Wood (1979; OW79) and Harley (1984; HA84) thermometers ($T_{OW79} \approx 1150°C$; $T_{HA84} \approx 1050°C$). However, Brey & Köhler (1990) and Taylor (1998) showed that the O'Neill & Wood (1979) formulation is characterized by large uncertainties, which probably arise from the small $\Delta G°$ of the Fe–Mg exchange reaction between garnet and olivine (Ganguly & Saxena, 1987). Small errors in the determination of garnet and olivine equilibrium compositions therefore propagate large errors on temperature estimates. Taking into account the significant grain-to-grain variations in FeO and MgO contents of olivine (~±1 wt % absolute; Evans & Trommsdorff, 1978) and the strong pressure dependence of the O'Neill & Wood (1979) thermometer, we calculate that the O'Neill & Wood estimates can be biased by as much as 200°C, in addition to the intrinsic uncertainty of their calibration (±60°C; O'Neill & Wood, 1979). Errors in chemical analyses of garnet would produce additional uncertainties (e.g. −30°C for +1% and −1% relative variations in FeO and MgO contents, respectively). Moreover, if redox conditions in the calibration experiments were similar to those in our samples, Fe_{tot}—rather than measured Fe^{2+}—contents should be used in thermometric calculations and temperature estimates would be lowered by some 40°C. Furthermore, high-MgO garnet cores, which are responsible for the very high calculated temperatures, may reflect local equilibrium with pre-existing magnesian spinel rather than olivine (see the section on Mineral Chemistry). From all these observations, it appears that, in our case, O'Neill & Wood (1979) estimates cannot provide reliable indications of peak metamorphic conditions. It is worth noting that O'Neill & Wood (1979) estimates for AA, MD, and CdG, assuming $P = 3$ GPa and using chemical analyses reported in Evans & Trommsdorff (1978) and our measured Fe^{2+}/Fe_{tot} ratios for garnets, give concordant results at 844, 836, and 773°C, respectively, whereas Ernst's (1978) analyses for AA yield values in the range 933–1060°C. In spite of this apparent discrepancy, which may reflect propagation of errors in the determination of Grt and Ol equilibrium compositions, temperatures obtained by means of Krogh's (1988) Grt–Cpx Fe–Mg exchange thermometer are 790 ± 48°C for both datasets, suggesting a greater robustness of Grt–Cpx thermometry to analytical errors.

As for the Harley (1984) thermometer, the evaluation by Brey & Köhler (1990, fig. 2d) showed that this thermometer overestimates temperatures at *T* significantly lower than 1100°C, the bias being progressively greater with lower *T*. Extrapolation of Brey & Köhler's data to $T = 800°C$ predicts deviations of some 100°C. The additional uncertainty of the Harley (1984) calibration (standard error of estimate = ±33°C), that related to Fe^{3+} determinations, and the effect of the pressure dependence of the thermometer should also be considered. For $P = 3$ GPa, 'corrected' Harley (1984) estimates based on Brenker & Brey's (1997) Opx and Grt analyses would be lowered to ~850–950°C.

Brenker & Brey (1997) utilized a new thermobarometric formulation based on Ca/Cr relations in garnet as a test for their extraordinary high *P–T* estimates. We have expressed their equation (1) as a thermometer and as a geobarometer, and have calculated temperatures and pressures for experiments on peridotitic compositions (Nickel, 1989; Brey *et al.*, 1990; Taylor, 1998). The results (Fig. 2) suggest that Brenker & Brey's formulation contains a strong compositional dependence, which would render their test totally inconsistent. Because only in the Brey *et al.* experiments have garnet compositions been reversed, the apparent compositional effect may be an artefact of incomplete garnet equilibration in Nickel's and Taylor's experiments. If we consider the Brey *et al.* data alone, the compositional effect is no longer evident, yet the uncertainties remain very large, with calculated standard errors of estimate of 1·2 GPa for *P* (mean error = +0·5 GPa) and 159°C for *T* (mean error = +68°C). Given these uncertainties, the Brenker & Brey formulation is of dubious validity. Furthermore, as the Ca and Cr contents of garnets are similar at AA and MD (Table 2), the *P–T* conditions obtained for AA should apply to MD as well. An extraordinary high-*P–T*

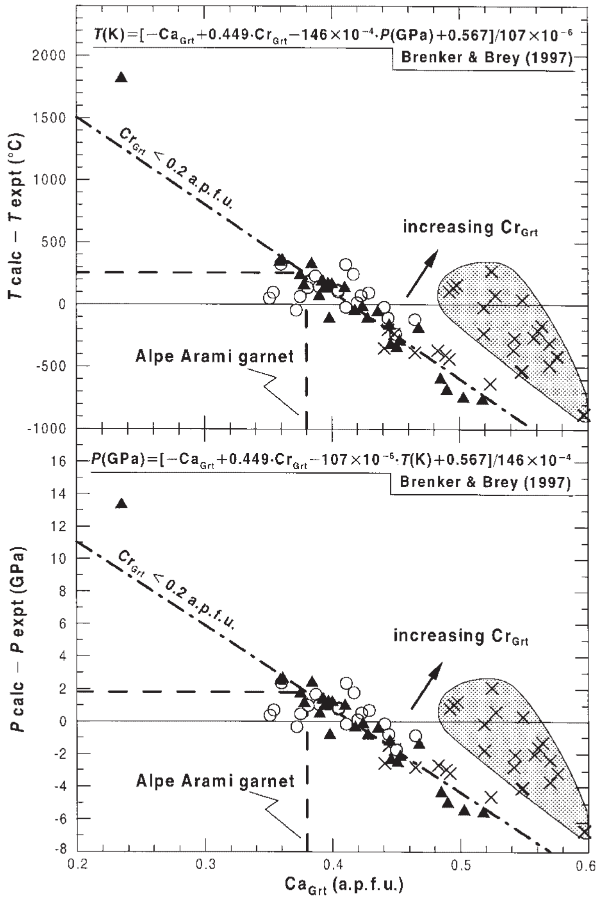

Fig. 2. Test of the Brenker & Brey (1997) garnet Ca/Cr thermobarometer against experiments on peridotitic compositions. Dash–dot lines show linear regressions for garnets with Cr contents comparable with those of Alpe Arami garnets (i.e. excluding data with Cr >0·2 atoms p.f.u. enclosed in shaded fields). For Ca contents equal to those in Alpe Arami garnets (dashed line) *T* and *P* can be greatly overestimated. Source of data: Brey *et al.* (1990; ○); Nickel (1989; ×); Taylor (1998; ▲).

metamorphism at MD is, however, excluded on the basis of both textural observations (see the section on Petrography) and other thermobarometric data (see below).

Paquin *et al.* (1999) recently obtained very high temperatures for AA using the Ni-in-Grt thermometer (Ol–Grt Ni-exchange) of Ryan *et al.* (1996) and a thermometer based on Ni–Mg partitioning between Ol and Opx (Podvin, 1988). Using the more recent version of the Ni-in-Grt thermometer proposed by Canil (1999) and Paquin *et al.*'s analyses for Ni, we calculate a temperature of ∼1100°C. This temperature is similar to that obtained by Brenker & Brey (1997) using the Fe–Mg Ol–Grt thermometer (O'Neill & Wood, 1979) and would correspond to pressures of ∼4–4·5 GPa (see below). However, as is the case for the O'Neill & Wood (1979) thermometer, the high temperature estimates yielded by Ni-based methods are of dubious significance, as: (1) high Ni contents in some garnets may reflect local equilibrium with pre-existing Ni-rich spinel (see the section on Mineral Chemistry); (2) the Podvin (1988) thermometer was calibrated on a simple system at $T \geq 1075$°C and its applicability to natural compositions has never been tested. Therefore, we believe that the extraordinary high *P–T* values estimated for Alpe Arami by Brenker & Brey (1997) and Paquin *et al.* (1999) have no solid basis.

New *P–T* estimates

We have calculated the pressures and temperatures of equilibration of garnet peridotites from our three localities (AA, MD, and CG) by using the mineral compositional data reported in Table 2 (porphyroclast core compositions), including those utilized by Brenker & Brey (1997) for their thermobarometric study of AA. We have taken care to use only the most up-to-date thermobarometers that have been calibrated on experimental datasets covering compositional fields relevant to natural peridotites and *P–T* ranges overlapping those of metamorphism in orogenic belts. Our preferred thermobarometer combination includes the two independent thermometers of Taylor (1998; pyroxene solvus) and Krogh (1988; Fe^{2+}–Mg exchange between Cpx and Grt) and the two independent barometers of Taylor (1998; Al-in-Opx) and Nimis & Taylor (2000; Cr-in-Cpx). The Ca-in-Opx version of the pyroxene solvus thermometer proposed by Brey & Köhler (1990) is complementary to the Cpx–Opx thermometer and can therefore provide some indications of the degree of equilibration between the pyroxenes. Because of the small temperature sensitivity of the pyroxene solvus at the Opx limb and because of the low Ca contents of our orthopyroxenes (CaO <0·3 wt %; Table 2), temperatures obtained by means of the Ca-in-Opx thermometer should be taken with caution. Therefore, this thermometer will not be used here to estimate peak metamorphic conditions and will only be considered for comparison with the other methods.

The results of our thermobarometric calculations are shown in Fig. 3. For AA and MD, the four independent thermobarometers yield concordant values and intersect one another at 3·2 ± 0·3 GPa and 844 ± 23°C (AA) and 3·0 ± 0·2 GPa and 829 ± 33°C (MD), respectively. Uncertainties given here refer to the *P–T* ranges for which the different thermobarometric methods yield results that are consistent within their respective uncertainties (1σ). Using Fe_{tot} rather than Fe^{2+} in calculations raises Cpx–Grt temperatures for AA and MD by 55°C, but the corresponding increase in pressure estimates is insignificant (<0·3 GPa). The agreement amongst the four independent thermobarometers indicates that a very good

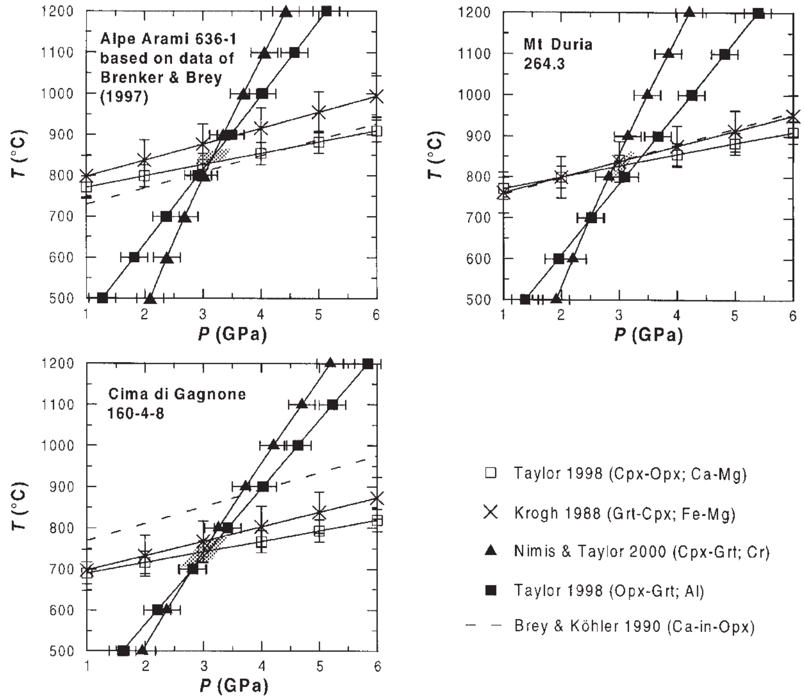

Fig. 3. Thermobarometry of garnet peridotites from Adula–Cima Lunga nappe based on mineral chemistry data reported in Table 2 (core compositions), including Brenker & Brey's (1997) data for Alpe Arami. The agreement amongst the four independent thermobarometers (continuous lines) suggests a very good degree of equilibration. Error bars are 1σ uncertainties as derived from thermobarometer calibrations, except for the Krogh (1988) thermometer (σ = 33°C), for which an uncertainty of 50°C has been assumed to account for errors in Fe^{3+}/Fe_{tot} determinations. It should be noted that the latter thermometer yields minimum temperature estimates because of unknown Fe^{3+} in experiments used for its calibration (see text for further explanation). Shaded area in each plot marks the *P–T* field for which the different thermobarometers yield results that are consistent within their respective uncertainties. Estimates based on the less precise Ca-in-Opx thermometer are shown for comparison as dashed lines.

degree of equilibration was achieved amongst the different minerals. [Cpx porphyroclasts at AA may contain exsolution lamellae of $(Mg_{1·57}Fe_{0·25}Na_{0·03}Al_{0·02}Ca_{0·14})$-$Si_{2·00}O_6$ clinoenstatite (~3 vol. %; Bozhilov *et al.*, 1999). Brenker & Brey (1997) did not check for the presence of such lamellae in their sample. Reintegration of 3% clinoenstatite into Brenker & Brey's Cpx would raise both Cpx–Opx and Cpx–Grt temperature estimates to 930°C and, accordingly, raise pressure estimates to 3·6 GPa. The four thermobarometers would still be in very good agreement with one another, but the change in pressure estimates would be minor (+ 0·4 GPa).] This is in total disagreement with the conclusions of Brenker & Brey (1997), but consistent with data on trace-element partitioning between pyroxenes and garnets (Rampone *et al.*, 1999) and with the results of Ca-in-Opx thermometry (Fig. 3). For CdG, the four independent thermobarometers tend to converge at a comparable pressure but slightly lower temperature (3·0 ± 0·4 GPa and 742 ± 38°C). The somewhat higher temperatures yielded by the Ca-in-Opx thermometer suggest a slightly poorer degree of equilibration, which is again consistent with trace-element data (Rampone *et al.*, 1999). It is worth noting that the effect of significant Fe^{3+} contents on pressure estimates is predictably negligible (Nimis & Taylor, 2000), although this has not been investigated experimentally. Assuming Fe^{3+} behaves like Cr (i.e. assuming $Cr = Cr + Fe^{3+}$ in pressure calculations), as can be expected from their similar ionic radii, the pressures calculated by means of the Cr-in-Cpx barometer remain within 0·04 GPa of the above estimates.

Thermobarometric results for AA based on additional mineral compositional data reported by Evans & Trommsdorff (1978) and Ernst (1978) are in good agreement with the above estimates. An exception is Ernst's sample F52-c, for which the two-pyroxene thermometer yields much higher temperatures (~1000°C). The high temperature values led Ernst (1981) to estimate a pressure of equilibration of 4–5 GPa for this sample. This anomalous high temperature is, however, supported neither by Grt–Cpx thermometry nor by the low CaO content of the orthopyroxene (0·17 wt %). The low Na_2O content of the clinopyroxene (0·35 wt %) also conflicts with a very high pressure estimate. Furthermore, at such high temperatures, the Cr-in-Cpx and Al-in-Opx barometers would yield significantly diverging results (3·5 GPa and

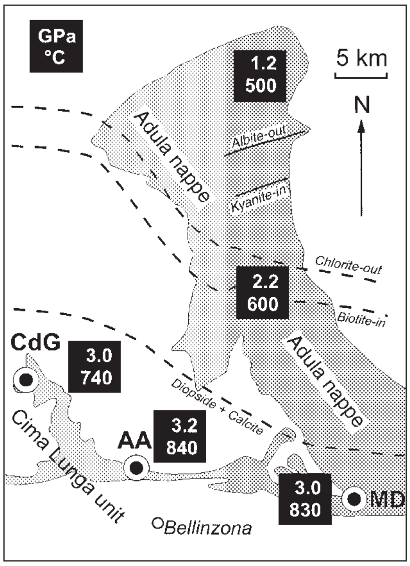

Fig. 4. Peak *P–T* conditions for the Eocene high-pressure metamorphism in the Adula–Cima Lunga nappe, based on thermobarometry of eclogites (Heinrich, 1986; Partzsch, 1996; Meyre *et al.*, 1997), metapelites (Heinrich, 1982; Meyre *et al.*, 1999), meta-ophicarbonate rocks (Pfiffner 1999) and garnet peridotites (this work). The metamorphic grade increases continuously to the south. Isograds for the Oligocene thermal metamorphism (dashed lines) after Trommsdorff (1966), Fox (1974) and Klein (1976).

4·3 GPa, respectively). The two barometers and the Ca-in-Opx and Grt–Cpx thermometers are instead in good agreement with one another and with our *P–T* estimates of 3·2 GPa and 844°C. We suspect that Ernst's analysis for clinopyroxene in sample F52-c does not reflect an equilibrium composition. The high *P–T* values proposed for this sample should be dismissed.

Our *P–T* estimates are in good agreement with data on pargasite stability and composition in MORB pyrolite (Niida & Green, 1999), which show an increase of Na with increasing *P–T*. Comparison of Niida & Green's data with experimental data for pargasite in various synthetic peridotitic systems (Mengel & Green 1989; Wallace & Green, 1991) indicates that Na_{Amph} vs *P–T* relations are little dependent on bulk composition as long as K is added to Na. Extrapolation of experimentally determined Na isopleths to $T = 850°C$ yields a pressure of ∼3·6 GPa based on the composition of a pargasite inclusion in an AA garnet (Na + K = 0·97 atoms per 23 oxygens; Table 1). Our pressure estimate of 3·2 GPa for AA corresponds to the maximum pressure of stability of K-free, MORB pyrolite pargasite, consistent with the almost total absence of prograde amphibole at AA. The non-negligible K contents (Table 1) may have extended the stability of pargasite inclusions to slightly higher pressures (Niida & Green, 1999). The compositions of CdG Ca-amphiboles, both included in garnet and in the matrix, (Na + K = 0·62–0·64 atoms per 23 oxygens; Table 1) are also consistent with our *P–T* estimates. Assuming $T = 750°C$, Na isopleths (Niida & Green, 1999) yield a pressure of ∼3·1 GPa.

DISCUSSION AND CONCLUSIONS

Thermobarometric estimates represent closure *P–T* conditions for chemical reactions amongst the peridotite minerals and, hence, provide minimum estimates for the peak conditions of metamorphism. However, the presence of prograde amphibole at CdG, MD and, probably, AA places an upper limit for pressure that is very close to the thermobarometric results. Therefore, our *P–T* estimates can be taken as representative of the metamorphic climax. The thermobarometric data are in very good agreement with geological evidence in the Adula–Cima Lunga unit. Our results match the regional pattern of *P–T* values obtained by Heinrich (1986) for associated eclogites of the whole Adula nappe, by Partzsch (1996) and Meyre *et al.* (1997) for those of the middle Adula, by Heinrich (1982) and Meyre *et al.* (1999) for metapelites of the middle Adula, and by Pfiffner (1999) for meta-ophicarbonate rocks of CdG (Fig. 4). The progressive southward increase of *P–T* values fits well the prograde, high-pressure metamorphic sequence of ultramafic rocks, which comprises from low to high grade and from north to south the assemblages antigorite + olivine → talc + olivine → enstatite + olivine (all with chlorite, clinoamphibole, and an increasingly chromian spinel phase) → garnet + diopside + enstatite + olivine ± Ca-amphibole (Trommsdorff *et al.*, 2000). The garnet peridotite at CdG is isofacial with prograde chlorite + amphibole peridotite. The slightly higher-*T* garnet peridotites at AA and MD are close to the stability limit of chlorite + amphibole peridotite and rarely retain relicts of prograde amphibole. The *P–T* estimates for AA and MD are identical within uncertainties, consistent with their similar location in the southernmost Adula–Cima Lunga nappe and their similar mineralogical and textural features. In contrast to the poikiloblastic garnet peridotite at CdG, the porphyroclastic garnet peridotites at AA and MD are better equilibrated both texturally and for trace elements (Rampone *et al.*, 1999), consistent with their higher temperatures of equilibration. The revised *P–T* values for CdG, AA, and MD (∼3 GPa and 750–850°C) are also in good agreement with the maximum *P–T* values derived for the Eocene high-pressure metamorphism in deeply subducted parts of the Western Alps (2·5–3·0 GPa and 600–750°C) and Liguria (2·5 GPa and 600°C). In general, the high-pressure metamorphism is younger in palaeogeographically more external parts of the Alpine edifice (Rubatto 1998) and maximum recorded *P–T* conditions increase in the same direction.

Despite the internal consistency amongst our *P–T* estimates for CdG, MD and AA, there still exists a remote possibility that some garnet cores at AA have preserved a memory of ancient, higher-temperature regimes (~1100°C), as suggested by garnet–olivine Fe–Mg–Ni thermometry. If correct, these thermometric estimates would place the AA peridotite at an original depth of ~140 km (4–4·5 GPa), which would still be a 'normal' depth for rocks exhumed during collisional processes. This hypothesis, however, implies that: (1) all spinel and amphibole inclusions in AA garnets, including those described by Möckel (1969) as 'primary', either were retrograde or survived metastably to 1–1·5 GPa above their stability limit; (2) Fe–Mg exchange between Cpx and Grt and Ca–Mg exchange between Cpx and Opx ceased after closure of garnet–olivine Fe–Mg–Ni reactions; (3) the AA peridotite followed a *P–T* path different from that shared by the other garnet peridotites in the Adula–Cima Lunga unit during Alpine high-pressure metamorphism. An alternative to point (3) is that the AA peridotite experienced a high-*P–T* regime in the mantle before its being involved in the Alpine subduction process, a scenario similar to that proposed for garnet peridotites from Western Norway and the Eastern Alps (Van Roermund & Drury, 1998; Nimis & Morten, 2000). In the absence of unambiguous proof (e.g. relics of majoritic garnet and subcalcic clinopyroxene) the ultrahigh-*P–T* hypothesis for Alpe Arami still lacks adequate documentation. In any case, maximum pressure estimates based on Cr-in-Cpx and Al-in-Opx barometers would not exceed 4–4·5 GPa. There appears to be no evidence for any of the Adula–Cima Lunga garnet peridotites being subducted to depths of 150 km or more.

The thermobarometric data presented in this study are corroborated by field geological, petrographic and mineralogical observations, and leave little room for the extraordinary high-*P–T* metamorphism scenario invoked by Dobrzhinetskaya *et al.* (1996), Bozhilov *et al.* (1999), Brenker & Brey (1997) and Paquin *et al.* (1999) for AA. The arguments put forth by Dobrzhinetskaya *et al.* and Bozhilov *et al.* to vindicate the hypothesis of retrogression and exhumation from great depths (e.g. exsolution of titanate in olivine and of clinoenstatite in diopside) were shown elsewhere to be inconclusive (e.g. Ulmer *et al.*, 1998; Arlt *et al.*, 2000). The data of Brenker & Brey and Paquin *et al.* (e.g. occurrence of high-Mg, high-Ni garnet cores) are interpreted by us as indicative of prograde metamorphism to depths of ~100 km.

We emphasize the importance of using appropriate thermobarometers for the estimation of *P–T* conditions of equilibration of metamorphic garnet peridotites. Thermometers characterized by low sensitivity to temperature variations and high sensitivity to uncertainties in the determination of equilibrium mineral compositions can produce errors greater than 200°C. Moreover, the use of a single barometer, which has been common practice in previous thermobarometric studies, can lead to grossly erroneous pressure estimates if the temperatures are incorrect or in case of disequilibrium (e.g. Nimis & Morten, 2000). Only a combination of well-tested thermometers and barometers based on independent chemical exchanges can indicate whether the minerals in the rock are in equilibrium and provide reliable *P–T* estimates. A discrepancy between thermobarometric data of eclogites and associated garnet lherzolites is nonexistent in our example, and in other localities it may also be a thermobarometric artefact.

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