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Mantle-Derived, UHP Garnet Pyroxenite and Eclogite in the Moldanubian Gföhl Nappe, Bohemian Massif: A Geochemical Review, New P-T Determinations, and Tectonic Interpretation

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Abstract

The Gföhl nappe, the uppermost structural unit in the Moldanubian Zone of the Variscan Bohemian Massif, contains a distinctive association of HP crustal granulite (900–1000°C, 15–18 kbar) and UHP mantle garnet peridotite (875–1150°C, 33–60 kbar). Ultrahigh-pressure (UHP) garnet peridotite is host to layers and lenses of garnet pyroxenite and eclogite, which formed by high-pressure crystal accumulation of garnet and pyroxene (± trapped melt) from transient melts in subcontinental lithosphere. The source of such melts was subducted, hydrothermally altered oceanic crust. New analyses of garnet websterite, orthopyroxene eclogite, and kyanite eclogite yield temperatures of 840–950°C and pressures of 34–43 kbar, comparable to those of enclosing peridotite, although kyanite eclogite at one locality (Úhrov) yields significantly different values of 1030–1200°C and 17–22 kbar. Most petrological and geochemical features of the Gföhl crustal and mantle association can be explained in terms of Devonian (Emsian to Famennian) convergence and subduction of Moldanubia beneath Tepla-Barrandia, culminating in Early Carboniferous (Tournaisian) continental collision. However, this tectonic scenario fails to account for a pressure gap of ~20 kbar between HP granulite and UHP peridotite-pyroxenite-eclogite, which remains problematic.

Introduction

THE BOHEMIAN MASSIF, the easternmost Variscan massif in Europe north of the Alpine front, is a tectonic collage resulting from the Devonian convergence and Carboniferous collision of Laurussia, Gondwana, and intervening amalgamated plates (Franke, 2000; Matte, 2001; Winchester et al., 2002). The Massif is divided into five main segments—Saxothuringia, Tepla-Barrandia, Moldanubia, Sudetes, and Moravo-Silesia (Fig. 1)—each of which is a composite terrane. Despite generally poor exposures and widespread high-temperature (HT) and medium-pressure (MP) to low-pressure (LP) overprinting, eclogite and other high-pressure (HP) rocks have been recognized in many tectonostratigraphic units in the massif (Fig. 1; Massonne and O'Brien, 2003, and citations therein). Devonian (Emsian to Famennian) bilateral subduction of Saxothuringia and Moldanubia beneath Tepla-Barrandia and subsequent Carboniferous (Tournaisian) dextral transpression produced two ages of HP metamorphism, one at 400–370 Ma and another at 340 Ma. The older HP metamorphism occurs in the Mariánské Lázně complex, the Münchberg klippe, the Zone of Erbendorf-Vohenstraus, and Góry Sowie, and the younger HP metamorphism is recorded in the Granulitgebirge, Ergebirge, Snieznik dome, and Gföhl nappe (Fig. 1). Eclogite also occurs in the Moldanubian Monotonous and Varied Groups (Medaris et al., 1995b), but except for a Sm-Nd age of 424 Ma for a single sample of eclogite from the Monotonous Group in the Oberpfalz (von Quadt and Gebauer, 1993), this HP event remains undated.

Most HP rocks in the Bohemian Massif yield pressures ≤25 kbar, but minerals indicative of ultrahigh pressure (UHP) have been discovered in several localities. In the Saxothuringian Erzgebirge, coesite occurs as inclusions in garnet in eclogite of the Eclogite-Gneiss Unit (Massonne, 2001), and microdiamonds are found as inclusions in garnet,

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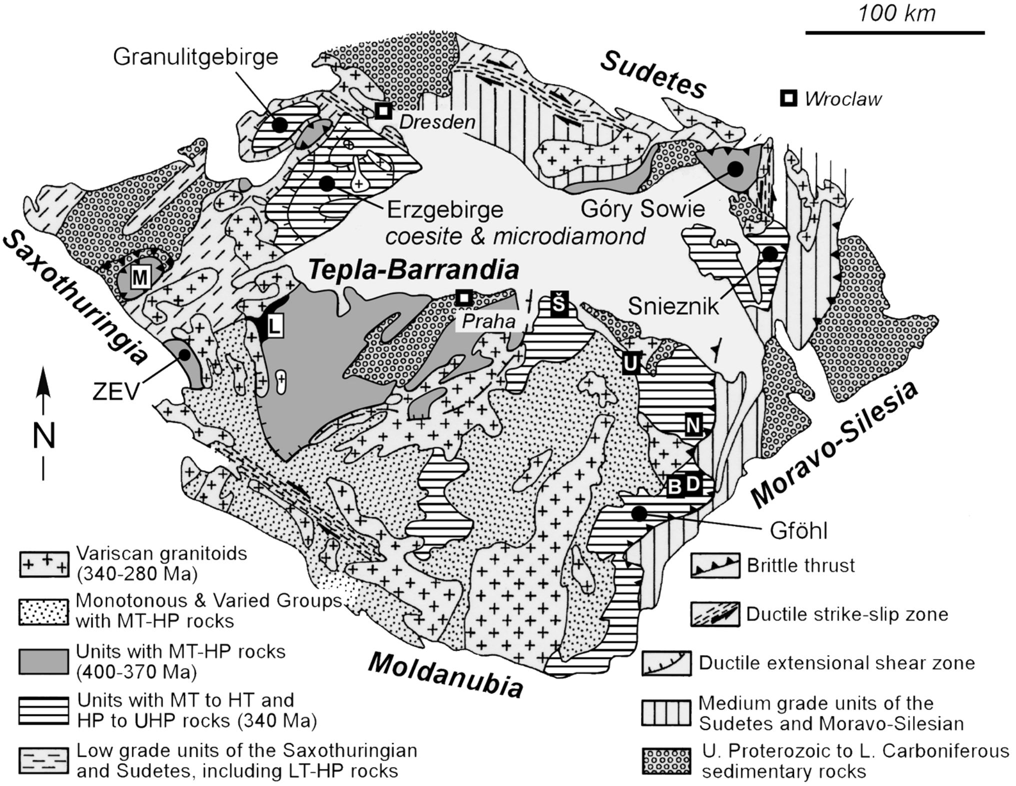

FIG. 1. Tectonostratigraphic map of the Bohemian Massif, illustrating the distribution and ages of HP and UHP metamorphism (modified from Willner et al., 2002). Abbreviations: L = Mariánské Lázně Complex; M = Münchberg Klippe; ZEV = Zone of Erbendorf-Vohenstraus. Peridotite localities: B = Biskupice; D = Nové Dvory; N = Níhov; Š = Dobešovice; U = Úhrov.

kyanite, and zircon in migmatitic gneiss, as well as in the Eclogite-Gneiss Unit (Massonne, 1998). In the Moldanubian Gföhl nappe, three occurrences of UHP minerals have been identified, including: (1) potassium feldspar lamellae in clinopyroxene in calc-silicate marble, for which the potassium content of the reconstructed clinopyroxene indicates pressures of 30 to 40 kbar (Becker and Altherr, 1992); (2) Na-bearing, grossular-rich garnet in a garnetite boudin in migmatitic gneiss, for which a pressure of ~40 kbar has been estimated (Vrána and Frýda, 2003); and (3) the association of wagnerite and kyanite in a pyropic garnetite boudin (Novák and Povondra, 1984).

In the Gföhl nappe several bodies of garnet peridotite yield calculated pressures of 30 to 60 kbar (Medaris et al., 2005), and two layers from peridotite (one a garnet websterite and another a kyanite eclogite) both equilibrated at a pressure of ~42 kbar (Medaris et al., 1995b; Nakamura et al., 2004). These UHP, mantle-derived ultramafic rocks are of interest because they are associated with HP crustal rocks (granulite), whose maximum pressure did not exceed 20 kbar (Carswell and O'Brien, 1993; Cooke et al., 2000).

Although P-T conditions of peridotites and associated HP granulite in the Gföhl nappe have been well established (Carswell and O'Brien, 1993; Cooke et al., 2000; Medaris et al., 2005), few P-T estimates have been made for garnet pyroxenite and eclogite layers in peridotite. Hereafter, such layers in Gföhl garnet peridotite will be referred to as the GGPE suite (Gföhl Garnet Pyroxenite Eclogite). In this investigation, we summarize the chemical composition and origin of the GGPE suite, present new P-T determinations for garnet websterite and kyanite eclogite members of the GGPE suite, and discuss mechanisms for juxtaposition of UHP mantle rocks and HP crustal rocks in the Gföhl Nappe.

Ultramafic Rocks in the Gföhl Nappe

The Gföhl nappe, which is the uppermost tectonic unit in Moldanubia and whose root zone is located at the boundary between Moldanubia and Tepla-Barrandia, contains two principal types of peridotite (Medaris et al., 1990). The Mohelno type consists largely of spinel peridotite, has high-Al_2O_3 orthopyroxene, contains subordinate spinel pyroxenite layers, equilibrated at 20–29 kbar, and has concordant contacts with surrounding HP granulite. Garnet occurs in Mohelno-type peridotite only within a few meters of contacts with granulite. In contrast, the Nové Dvory type consists of garnet peridotite, has low-Al_2O_3 orthopyroxene, contains relatively abundant layers of garnet pyroxenite and eclogite (the GGPE suite), records pressures of 30–60 kbar, and has discordant contacts with the surrounding migmatitic Gföhl gneiss. Based on lithological characteristics, elemental composition, radiogenic isotope composition, and P-T conditions, the Mohelno type peridotite is thought to represent suboceanic asthenospheric mantle, and the Nové Dvory type, subcontinental lithospheric mantle (Medaris et al., 2005).

The GGPE suite contains two major lithologies, garnet pyroxenite and eclogite. Although both these rock types consist primarily of garnet and clinopyroxene, they are distinguished on the basis of chemical composition, e.g., the Mg# for garnet pyroxenite is ≥80, and for eclogite is ≤80 (see the following section for additional chemical distinctions). Each rock group is further subdivided according to mineralogical composition. Garnet pyroxenite includes garnet clinopyroxenite (grt + cpx) and garnet websterite (grt + cpx + opx), and eclogite includes common eclogite (grt + cpx), kyanite eclogite (grt + cpx + ky), and orthopyroxene eclogite (grt + cpx + opx). In general, garnet pyroxenite layers and lenses range in thickness from millimeters to centimeters (Fig. 2A), whereas eclogite layers tend to be much larger, ranging from layers centimeters thick to bodies as large as 30 × 600 meters (Fig. 2B). Variations in proportions of garnet and pyroxene impart a pronounced layering to many eclogites (Fig. 2B), and nearly monomineralic layers of orthopyroxene, clinopyroxene, or garnet occur in some samples of orthopyroxene eclogite (Fig. 2C).

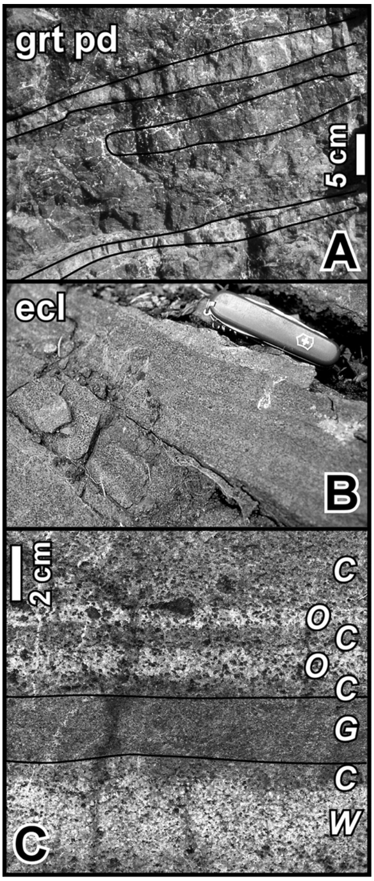

FIG. 2. A. Outcrop of thin garnet pyroxenite layers and lens in the Bečváry garnet peridotite (these features are outlined for clarity). B. Outcrop of large eclogite layer in the Nové Dvory garnet peridotite. Note the mineralogical layering. C. Hand sample of strongly layered orthopyroxene eclogite in the Biskupice garnet peridotite; abbreviations: C, garnet clinopyroxenite; G, garnetite; O, garnet orthopyroxenite; W, garnet websterite.

Geochemistry

A substantial geochemical database is available for the GGPE suite (Scharbert and Carswell, 1983; Carswell and Jamtveit, 1990; Brueckner et al., 1991; Beard et al., 1992; Medaris et al., 1995a; Becker, 1996). Although eclogite and garnet pyroxenite are mineralogically similar, in that both consist largely of garnet and clinopyroxene, they can readily be separated on a chemical basis. Eclogite is distinguished from garnet pyroxenite by an Mg# <80, Na_2O >0.8 wt%, Cr_2O_3 <0.15 wt%, and Ni

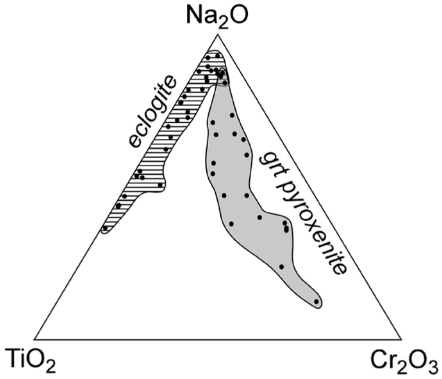

FIG. 3. Discrimination between Gföhl garnet pyroxenite and eclogite, based on variation in Na_2O, TiO_2, and Cr_2O_3 (wt%).

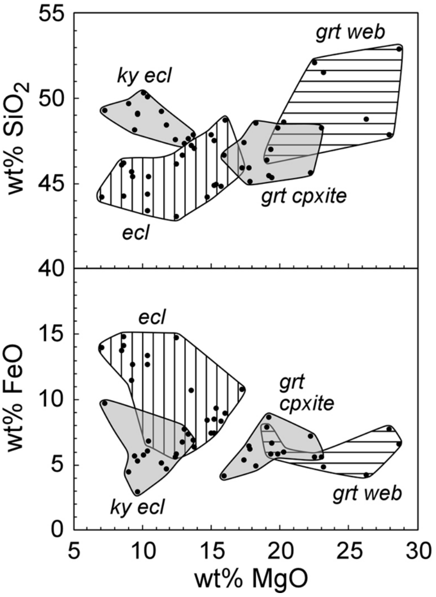

FIG. 4. Variation of wt% SiO_2 and wt% FeO_{total} with wt% MgO in Gföhl garnet websterite, garnet clinopyroxenite, eclogite, and kyanite eclogite, illustrating the large scatter in compositions. Eclogite and kyanite eclogite are effectively separated in these two plots, but largely overlap in terms of other elements, such as CaO, Al_2O_3, and Na_2O.

<400 ppm. A plot of Na_2O, TiO_2, and Cr_2O_3 (Fig. 3) provides an effective means of discriminating between the two rock types, with only a small overlap occurring near the Na_2O apex of the plot. Within each rock group, further distinctions can be made on a mineralogical basis, as described previously, e.g., the presence of orthopyroxene or kyanite.

A characteristic feature of the GGPE suite is a wide range in major-element composition and large degree of scatter in Harker plots (Fig. 4). As a whole, the suite shows a range in MgO from 7 to 29 wt%; SiO_2, 43 to 53%; FeO_T, 3 to 15%; Al_2O_3, 6 to 21%; and CaO, 5 to 17%. Considerable scatter in the data occurs within each of the four rock types in the entire dataset, as well as within samples from individual bodies, reflecting in large part the modal variation of garnet and clinopyroxene and layered aspect of many samples.

Normalized REE plots for members of the GGPE suite (not reproduced here) show a range of patterns, including those with negative slopes, positive slopes, and convex-upward shapes. This diverse range of REE patterns has been successfully reproduced by a model of garnet and clinopyroxene accumulation ± variable amounts of trapped melt (Medaris et al., 1995a). Over half the analyzed samples display an Eu anomaly, either positive or negative, which suggests the presence of plagioclase and influence of LP processes at an early stage in petrogenesis of the GGPE suite.

Clinopyroxene and garnet in the GGPE suite are in HT oxygen isotope equilibrium, with $\Delta_{Cpx-Grt} = -0.06 \pm 0.29‰$ (1σ, $n = 9$). Values of $\delta^{18}O_{SMOW}$ for clinopyroxene range from 4.8 to 5.7‰ in garnet pyroxenite and 3.8 to 5.8‰ in eclogite, with six samples clustering around values of 4.7 to 5.0‰. Such values are isotopically light in comparison to a typical value of 5.6‰ for mantle clinopyroxene, and may be inherited from an early stage of hydrothermal alteration of oceanic crust (Medaris et al., 1995a).

The ε_{Nd} values and $^{87}Sr/^{86}Sr$ ratios calculated at 335 Ma for the GGPE suite are negatively correlated, with ε_{Nd} values ranging from +6.0 to –6.0 and $^{87}Sr/^{86}Sr$ ranging from 0.7027 to 0.7145 (Fig. 5). This range in Nd and Sr isotope compositions plots along the oceanic mantle array extending from a depleted MORB mantle (DMM) component to an EMII component, although the GGPE suite extends to more radiogenic $^{87}Sr/^{86}Sr$ ratios compared to oceanic basalts (Fig. 5). The GGPE suite, except for two kyanite eclogites from Dobešovice that possess very radiogenic $^{87}Sr/^{86}Sr$ ratios, have the same range of Nd and Sr isotope compositions as Cr-diopside and Al-augite pyroxenites from Beni Bousera (Fig. 5; Pearson et al., 1993). Data for the pyroxenite group (clinopyroxenite and websterite) completely overlap those for the eclogite group (eclogite and kyanite eclogite), aside from the two radiogenic Dobešovice samples. The Nd and Sr isotope variations in the GGPE suite can be closely modeled by mixing between a DMM component and up to 15% oceanic clay (Medaris et al., 1995a). Two eclogite samples from the Úhrov locality (ε_{Nd} = +4.9; [^{87}Sr/^{86}Sr]$_i$ = 0.709) depart from the oceanic trend and have been attributed to mixing of DMM with 15–20% arc detritus (Medaris et al., 1995a).

The GGPE suite is thought to have formed by HP crystal accumulation of garnet, clinopyroxene, and accessory phases (± trapped melt) from transient melts in lithospheric mantle, based on the presence of modal layering, the large degree of scatter in oxide variation diagrams, the correlation between elemental compositions and modes, the results of REE modeling, and the absence of any prograde metamorphic features. The existence of Eu anomalies and Nd, Sr, and O isotope data strongly indicate that subducted oceanic crust contributed to the melts that crystallized to form the various members of the GGPE suite. A similar hypothesis has been advocated for the petrogenesis of equivalent rocks in the Beni Bousera (Pearson et al., 1993), Balmuccia (Shervais and Mukasa, 1991), and Ronda (Suen and Frey, 1987) peridotite massifs. Contrary to this hypothesis, it has been proposed that kyanite eclogite in the GGPE suite (Obata et al., 2006) and corundum-bearing garnet clinopyroxenite at Beni Bousera (Kornprobst et al., 1990) represent oceanic crust that was subducted, accreted to lithospheric mantle, and recrystallized at high pressure. Further investigation of kyanite eclogites in the GGPE suite is under way to explore this intriguing question.

Pressure-Temperature Conditions

The predominant mineral assemblage in the GGPE suite is a nonlimiting, bimineralic association of garnet and clinopyroxene (plus accessory phases), from which only minimum pressure estimates may be obtained. Consequently, we have analyzed minerals in several samples of garnet websterite, orthopyroxene eclogite, and kyanite eclogite, which contain limiting assemblages, to determine P-T conditions for the GGPE suite, and to compare such results with those for the enclosing garnet peridotites and those for granulite and eclogite in the surrounding country rocks.

For orthopyroxene-bearing samples, P-T estimates can be obtained by a combination of Fe-Mg exchange geothermometers and the Al-in-Opx geobarometer (Fig. 6A), assuming that all reactions equilibrated at the same time. For Fe-Mg exchange between garnet and orthopyroxene, we have used calibrations by Brey and Köhler (1990) and Harley (1984), and for garnet-clinopyroxene, that by Powell (1985). Because of the generally small amounts of iron in clinopyroxene and the difficulty in estimating Fe^{2+}/Fe^{3+} ratios by stoichiometry, all iron is treated as ferrous in the calculations. Results from the Brey and Köhler (1990) calibration are 20–40°C less than those from the Harley (1984) formulation, and results from the garnet-clinopyroxene geothermometer are comparable to those from the two garnet-orthopyroxene methods, with no consistent difference, either higher or lower, from the two. Two versions of the Al-in-Opx geobarometer have been used (Brey and Köhler, 1990; Nickel and Green, 1985), which yield closely similar pressures at 800–900°C and slightly increasing divergence at higher temperatures (Fig. 6A).

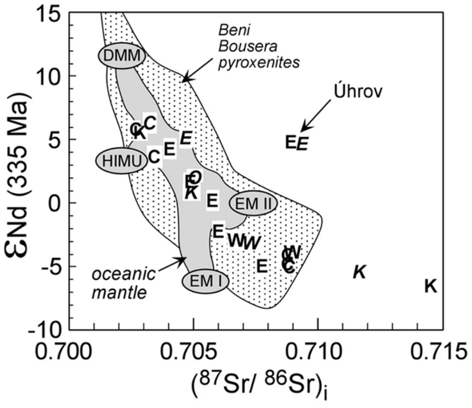

FIG. 5. Variation in ε_{Nd} and [^{87}Sr/ ^{86}Sr]$_i$ (at 335 Ma) in Gföhl garnet pyroxenite and eclogite. Abbreviations: C = garnet clinopyroxenite; E = eclogite; K = kyanite eclogite; O = orthopyroxene eclogite; W = garnet websterite (upright letters, clinopyroxene analyses; italicized letters, whole-rock analyses). End-member mantle compositions, DMM, HIMU, EMI, and EMII from Hart (1988). Compositional field for Beni Bousera pyroxenites from Pearson et al. (1993).

For kyanite eclogite, P-T estimates were made from a combination of the garnet-clinopyroxene Fe-Mg exchange geothermometer and the reaction:

$$\underset{\text{diopside}}{3CaMgSi_2O_6} + \underset{\text{kyanite}}{2Al_2SiO_5} = \underset{\text{grossular}}{Ca_3Al_2Si_3O_{12}} + \underset{\text{pyrope}}{Mg_3Al_2Si_3O_{12}} + \underset{\text{coesite/quartz}}{2SiO_2}, \quad (1)$$

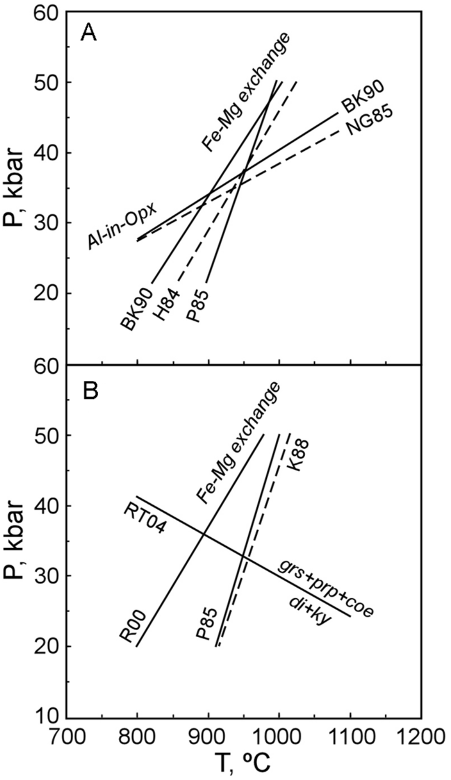

FIG. 6. A. Equilibrium curves for garnet-orthopyroxene and garnet-clinopyroxene Fe-Mg exchange geothermometers and Al-in-Opx geobarometers in sample CZ4A, garnet websterite, Níhov. Abbreviations: BK90 = Brey and Köhler (1990); H84 = Harley (1984); P85 = Powell (1985); NG85 = Nickel and Green (1985). B. Equilibrium curves for three calibrations of the garnet-clinopyroxene Fe-Mg exchange geothermometer and reaction [1] geobarometer for sample CZ12C, kyanite eclogite, Dobešovice. Abbreviations: K88 = Krogh (1988); P85 = Powell (1985); R00 = Ravna (2000); RT04 = Ravna and Terry (2004).

as illustrated in Figure 6B. Calibrations of the garnet-clinopyroxene geothermometer by Powell (1985), Krogh (1988), and Ravna (2000) have been used, with the Powell and Krogh methods giving comparable results, and the Ravna formulation yielding temperatures about 90° lower at 30 kbar (Fig. 6B). The equilibrium position for reaction (1) was calculated by the method of Ravna and Terry (2004), which utilizes the thermodynamic database of Holland and Powell (1988), the clinopyroxene activity model of Holland (1990), and the garnet activity model of Ganguly et al. (1996).

Minerals were analyzed at the University of Wisconsin on a Cameca SX 50 electron microprobe, operating in a wavelength dispersive mode and using an accelerating voltage of 15 kV, a beam current of 20 nA, a suite of analyzed natural minerals as standards, and the $\phi(\rho Z)$ data reduction program (Armstrong, 1988). Commonly, garnet and pyroxene in the GGPE suite are compositionally zoned, due to the effects of arrested retrograde re-equilibration. The most significant compositional variations that may affect P-T estimates are an increase in Fe/Mg ratio at garnet margins, a core-to-rim increase in Al_2O_3 in orthopyroxene, and a decrease in jadeite content at clinopyroxene rims. Mineral core compositions were used in the P-T calculations, and representative analyses are listed in Tables 1 and 2.

P-T results for 11 samples are shown as polygons in Figure 7; each polygon outlines the combination of equilibrium curves calculated for that sample. Two samples of garnet websterite and two of orthopyroxene eclogite yield temperatures from 850 to 950°C and pressures from 34 to 43 kbar, consistent with pressures >30 kbar for the enclosing garnet peridotites. However, temperatures for these samples are appreciably lower than those for garnet peridotite, which may be an artifact arising from different thermobarometric approaches to the two rock types. Fe-Mg exchange temperatures for peridotite were obtained from the core compositions of garnet, a slowly diffusing phase, and the composition of olivine, which, although a fast diffusing phase, is the predominant mineral in peridotite and whose composition will change little as a result of Fe-Mg exchange with garnet during cooling. In contrast, Fe-Mg exchange temperatures for garnet websterite and orthopyroxene eclogite were obtained from garnet cores and coexisting orthopyroxene and clinopyroxene, whose abundances are subequal to that of garnet and whose compositions are likely to change to some degree by cation exchange with garnet during cooling. Consequently, garnet peridotite is more likely to preserve peak temperatures than are garnet websterite and orthopyroxene eclogite, which probably record post-peak, blocking temperatures, depending on their cooling histories.

Three samples of kyanite eclogite from Dobešovice yield values of 840–940°C and 34–37 kbar, similar to those for the orthopyroxene-bearing rock types. Kyanite eclogite from Nové Dvory, which was analyzed by Nakamura et al. (2004), gives 1100°C, 41 kbar by our calculation method, compared to 1050–1150°C, 45–49 kbar by their method,

TABLE 1. Analyses of Minerals in Garnet Websterite and Orthopyroxene Eclogite

Locality:	Níhov						Biskupice					
Sample:	CZ4A			CZ4B			CZ14I			BIS85		
Rock:		grt web			grt web			opx ecl			opx ecl	
Mineral:	grt	opx	cpx	grt	opx	cpx	grt	opx	cpx	grt	opx	cpx
						wt%						
SiO_2	41.27	56.60	54.10	40.97	56.70	53.63	40.20	55.90	54.80	40.95	56.17	54.41
TiO_2	0.00	0.00	0.00	0.00	0.00	0.00	0.12	0.08	0.11	0.16	0.06	0.09
Al_2O_3	22.33	0.86	2.41	22.20	0.66	1.84	22.20	0.71	2.21	21.91	0.71	1.83
Cr_2O_3	1.12	0.14	0.59	1.43	0.12	0.68	0.03	0.08	0.53	0.90	0.06	0.33
FeO	10.20	6.73	2.73	9.86	6.35	2.60	15.90	10.90	4.49	13.19	10.04	3.72
MnO	0.37	0.15	0.08	0.39	0.09	0.07	0.38	0.11	0.05	0.38	0.15	0.07
MgO	19.67	34.40	16.80	19.73	35.50	16.93	16.30	31.90	15.90	16.84	32.15	16.13
CaO	4.95	0.40	22.30	5.08	0.17	23.13	4.64	0.38	21.00	4.88	0.30	21.68
Na_2O			0.88			0.63			1.54			1.29
Sum	99.90	99.36	99.89	99.65	99.59	99.53	99.77	100.22	100.63	99.20	99.66	99.55
					No. of cations per oxygen atoms							
O	12	6	6	12	6	6	12	6	6	12	6	6
Si	2.969	1.969	1.964	2.956	1.964	1.960	2.974	1.966	1.983	3.003	1.976	1.987
Ti	0.000	0.000	0.000	0.000	0.000	0.000	0.007	0.002	0.003	0.009	0.002	0.003
Al	1.893	0.035	0.103	1.888	0.027	0.079	1.918	0.029	0.094	1.894	0.030	0.079
Cr	0.063	0.004	0.017	0.082	0.003	0.020	0.001	0.002	0.015	0.052	0.002	0.010
Fe	0.614	0.196	0.083	0.595	0.184	0.080	0.986	0.321	0.136	0.809	0.295	0.114
Mn	0.023	0.004	0.002	0.024	0.003	0.002	0.025	0.003	0.002	0.023	0.004	0.002
Mg	2.109	1.784	0.909	2.123	1.833	0.922	1.772	1.673	0.858	1.841	1.686	0.878
Ca	0.382	0.015	0.867	0.392	0.006	0.905	0.375	0.014	0.814	0.383	0.011	0.848
Na			0.062			0.045			0.108		0.000	0.091
Sum	8.053	4.014	4.007	8.059	4.021	4.013	8.059	4.021	4.013	8.015	4.007	4.012

TABLE 2. Analyses of Minerals in Kyanite Eclogite

Locality:	Dobešovice						Úhrov					
Sample:	CZ12B		CZ12C		CZ12D		CZ9C		CZ9D		CSUR1	
Mineral:	grt	cpx	grt	cpx	grt	cpx	grt	cpx	grt	cpx	grt	cpx
						wt% oxides						
SiO_2	40.12	54.83	40.25	55.50	40.62	54.33	40.65	52.25	40.70	51.60	40.90	52.28
TiO_2	0.00	0.18	0.00	0.00	0.00	0.09	0.00	0.00	0.00	0.00	0.16	0.18
Al_2O_3	22.36	10.06	22.46	9.37	22.58	10.34	22.48	11.13	22.50	7.97	22.88	9.01
Cr_2O_3	0.23	0.10	0.00	0.07	0.04	0.09	0.29	0.06	0.06	0.08	0.10	0.12
FeO	13.40	2.13	12.48	2.45	12.05	2.40	8.37	2.83	9.59	3.01	9.06	2.70
MnO	0.31	0.00	0.31	0.04	0.28	0.00	0.35	0.04	0.23	0.05	0.24	0.03
MgO	12.13	11.82	11.99	11.79	12.62	11.67	11.43	11.59	10.45	13.37	12.23	12.51
CaO	10.94	16.47	11.90	16.40	11.78	17.14	16.08	19.38	16.55	21.77	14.76	20.11
Na_2O		4.59		4.72		4.30		2.96		1.44		2.55
Sum	99.47	100.16	99.38	100.33	99.94	100.35	99.63	100.23	100.08	99.28	100.31	99.49
						No. of cations per oxygen atoms						
O	12	6	12	6	12	6	12	6	12	6	12	6
Si	2.985	1.943	2.991	1.965	2.992	1.928	2.994	1.871	3.001	1.881	2.985	1.891
Ti	0.000	0.005	0.000	0.000	0.000	0.002	0.000	0.000	0.000	0.000	0.009	0.005
Al	1.961	0.420	1.966	0.391	1.960	0.432	1.951	0.470	1.955	0.342	1.968	0.384
Cr	0.013	0.003	0.000	0.002	0.002	0.002	0.017	0.002	0.003	0.002	0.006	0.003
Fe	0.833	0.063	0.775	0.073	0.742	0.071	0.516	0.085	0.591	0.092	0.553	0.082
Mn	0.019	0.000	0.020	0.001	0.017	0.000	0.022	0.001	0.014	0.001	0.015	0.001
Mg	1.345	0.624	1.327	0.622	1.385	0.617	1.254	0.619	1.149	0.726	1.331	0.674
Ca	0.872	0.625	0.947	0.622	0.930	0.652	1.269	0.743	1.307	0.850	1.154	0.779
Na		0.315		0.324		0.296		0.206		0.102		0.179
Sum	8.028	3.998	8.026	4.000	8.027	4.001	8.022	3.996	8.020	3.997	8.020	3.999

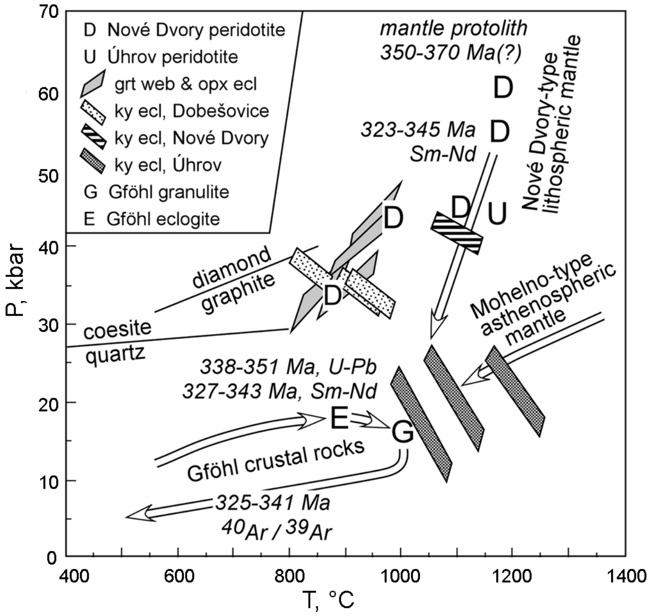

FIG. 7. Summary of P-T estimates and age determinations for various rock types in the Gföhl nappe. Arrows indicate the P-T paths for Gföhl mantle rocks and crustal rocks. See text for further discussion.

which is based on a different thermodynamic evaluation of reaction (1). In either case, such conditions are comparable to those of the enclosing Nové Dvory garnet peridotite.

Three samples of kyanite eclogite from Úhrov yield 1030–1200°C and 17–22 kbar (assigning quartz as the SiO_2 phase in reaction [1]). Such values are significantly different from those of the other eclogites, pyroxenites, and garnet peridotites, particularly with respect to pressure, and they lie intermediate between the maximum P-T conditions for eclogite and HP granulite in the Gföhl country rocks (900–1000°C, 16–18 kbar) and the P-T array for Mohelno-type, suboceanic asthenospheric mantle (1120–1335°C, 22–29 kbar). However, P-T conditions for the Úhrov garnet peridotite (1170°C, 43.8 kbar) are similar to those for the Nové Dvory garnet peridotite (Fig. 7). Thus, the P-T characteristics of the Úhrov body do not fall neatly into one or the other of the two principal Gföhl peridotite types. The Úhrov peridotite is located in the Běstvina Formation (thought to be a Gföhl correlative) of the Kutna Hora-Svratka complex, and it may represent an additional type of mantle with a different P-T-t history. Moreover, although the Úhrov kyanite eclogite has not yet been analyzed for Nd and Sr isotopes, two samples of eclogite from this locality are isotopically distinct from the oceanic trend of the other GGPE eclogites and garnet pyroxenites (Fig. 5).

Discussion

It has been proposed that the Mohelno type peridotite represents suboceanic asthenospheric mantle and the Nové Dvory type peridotite, which hosts the GGPE suite, was derived from subcontinental lithospheric mantle, based on bulk major and trace-element compositions, mineral compositions, Nd and Sr isotopes, and P-T regimes (Medaris et al, 2005). Sm-Nd mineral isochron ages for Gföhl garnet peridotites range from 329 to 354 Ma (mean = 339 ± 10 Ma, *n* = 7), although the Mohelno garnet peridotite *sensu stricto* gives a significantly older age of 371 Ma (Beard, 1992; Becker, 1997). Members of the GGPE suite yield Sm-Nd ages similar to those in the host garnet peridotites, ranging from 324 to 344 Ma (mean = 336 ± 7 Ma, *n* = 9), although older Sm-Nd ages of 370, 373, and 377 Ma were obtained from garnet pyroxenites at Mitterbachgraben, Níhov, and Bečváry, respectively (Carswell and Jamtveit, 1990; Brueckner et al., 1991; Beard et al., 1992; Becker, 1997). The mean Sm-Nd ages of 336 and

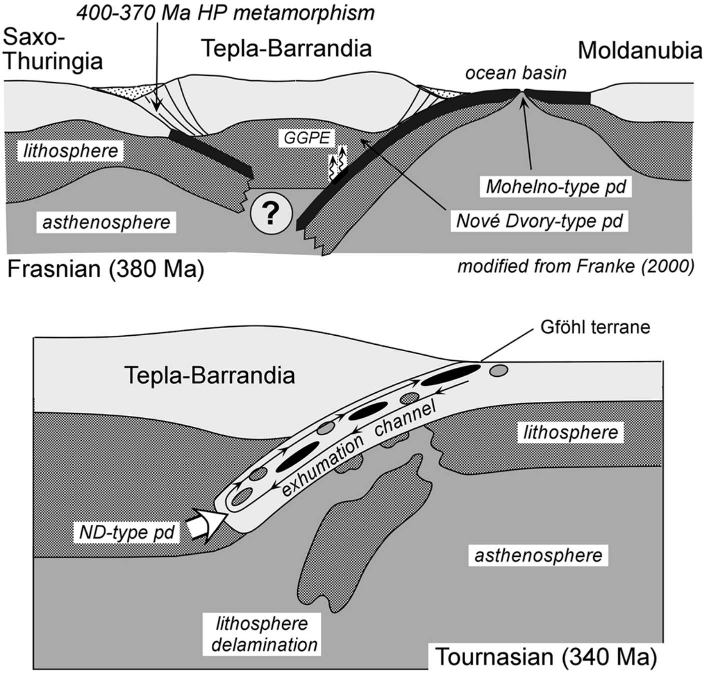

FIG. 8. Proposed tectonic scenario to account for the origin and evolution of Mohelno type peridotite, Nové Dvory type peridotite, and the GGPE suite in the Gföhl nappe. Patterns: black = oceanic crust; light grey = continental crust; heavy stipple = lithospheric mantle; dark grey = asthenospheric mantle.

339 Ma for the GGPE suite and Gföhl garnet peridotites are comparable to the U-Pb ages for metamorphic zircon from Gföhl granulite, which range from 338 to 347 Ma and are taken to represent the time of HT-HP metamorphism in the Gföhl Nappe (Kröner et al., 2000, and citations therein).

The provenance of peridotites in the Gföhl nappe can be envisioned in the context of the tectonic scenario proposed by Franke (2000). During Frasnian time, bilateral subduction was occurring beneath Tepla-Barrandia (Fig. 8, upper panel). Moldanubia, which was either a member of the Armorican terrane assemblage or a fragment of northern Gondwana, was separated from Tepla-Barrandia by a small ocean basin and spreading center, which allowed the shallow rise of asthenospheric mantle (Mohelno type peridotite). Fluids derived from dehydration and/or melting of subducted oceanic crust infiltrated the overlying lithospheric wedge (Nové Dvory type peridotite) to produce the GGPE suite. HP metamorphism was generated at this time in material that was subducted and accreted at the northern margin of Tepla-Barrandia, e.g., the Mariánské Lázně complex and Münchberg eclogite. Sm-Nd ages of 370 to 377 Ma, which have been preserved in a few Gföhl garnet pyroxenites, may record the effects of this Late Devonian convergence on the lithospheric mantle beneath Tepla-Barrandia.

With closure of the ocean basin and initial collision of Moldanubia and Tepla-Barrandia, imbrication of asthenospheric mantle (Mohelno type), oceanic crust, and continental crust may have occurred in the vicinity of the subducted spreading center, giving rise to most lithologic members of the Gföhl granulite unit and elevating temperatures in the crustal rocks. The 371 Ma Sm-Nd age for the Mohelno garnet peridotite may reflect the initial juxtaposition of hot mantle and cooler crust in Late Devonian time.

Continued convergence culminated in the Early Carboniferous collision of Moldanubia and Tepla-Barrandia, with subduction of the former under the latter (Fig. 8, lower panel). Penetration of Moldanubia to lithospheric depths may have allowed the subducted crustal slab to incorporate fragments of subcontinental lithosphere, which were then distributed through the crustal slab by exhumation channel flow and which now appear as Nové Dvory type peridotites and accompanying GGPE layers in the Gföhl nappe. At this stage in the tectonic scenario, attainment of HT-HP conditions by the Gföhl crustal rocks and extraction and quenching of Nové Dvory type peridotite from subcontinental lithosphere are recorded by numerous U-Pb and Sm-Nd ages in the range 340–335 Ma.

Diffusion modeling of compositionally zoned garnet (O'Brien and Vrána, 1995; O'Brien, 1997; Cooke et al., 2000; Medaris et al., 2005, 2006), $^{40}Ar/^{39}Ar$ ages of 341–329 Ma for hornblende (Dallmeyer et al., 1992), and the presence of Gföhl detritus in Lower Viséan (330–334 Ma) conglomerates of the Moravian foreland basin (Dvořak, 1982) indicate that the Gföhl granulite terrane resided at peak metamorphic conditions for a limited time and experienced rapid cooling (100–165°C/m.y.) and exhumation (7–11 mm/year). Such short-lived metamorphic culmination and rapid exhumation of the HT-HP Gföhl terrane can be attributed either to detachment of lithospheric mantle through slab break-off or delamination (Arnold et al., 2001; Willner et al., 2002), as shown in Figure 8, or to compression-driven extrusion of a softened orogenic root in a transpressive regime (Thompson et al., 1997; Schulmann et al., 2002; Svojtka et al., 2002).

Although the tectonic scenario described above is consistent with most of the petrological and isotopic features of the Gföhl terrane, it fails to account for the ~20 kbar gap in peak pressure between the UHP Nové Dvory type mantle rocks and the HP Gföhl crustal granulite (Fig. 7). It appears that the mantle and crustal rocks attained their maximum P-T conditions at about the same time, ~340 Ma, but at different depths, shortly after which they were tectonically juxtaposed and subsequently shared a common retrograde history during exhumation.

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