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Petrological and geochemical constraints on the origin of garnet peridotite in the North Qaidam ultrahigh-pressure metamorphic belt, northwestern China

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Abstract

In the North Qaidam ultrahigh-pressure (UHP) metamorphic belt, NW China, a large garnet peridotite massif has been exhumed together with felsic gneisses and eclogite during Early Paleozoic continental collision between North China Craton and Qaidam–Qilian Block. Four rock types are recognized in this massif: dunite without garnet, dunite with garnet, garnet lherzolite and garnet pyroxenite. The bulk compositions of these rocks retain depleted major element characteristics, but possess elevated abundances of incompatible trace elements, especially the fluid-mobile large ion lithophile elements (LILE, e.g., Rb, Cs, Ba, Sr, U, and Pb). The unradiogenic Nd ($\varepsilon_{Nd[460 Ma]} - 0.5$ to -6.8) isotopes of these rocks are consistent with their flat to light rare earth element (LREE)-enriched patterns. Petrological and geochemical data altogether suggest that the protoliths of the garnet peridotite massif represent rocks from a subarc system in the Early Paleozoic. These rocks were then further metasomatized by the ascending melts before being transported to depths of 150–200 km or more by the subducting slab-induced mantle wedge corner flow. Their subsequent exhumation to the upper crust is likely to have been accomplished during continental collision with a recovery rate of about 7.5–10 mm/year.

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Keywords: Garnet peridotite; Cumulate origin; Continental subduction; North Qaidam UHP belt; NW China

1. Introduction

Garnet peridotite is volumetrically small, yet a common component in ultrahigh-pressure (UHP) metamorphic terranes in many orogenic belts such as the well-studied West Gneiss Region of Norway (e.g.,

* Corresponding author. Tel./fax: +86 10 62767729. *E-mail address:* sgsong@pku.edu.cn (S. Song). O'Hara and Mercy, 1963; Carswell et al., 1983; Medaris and Carswell, 1990) and the Dabie–Sulu terrane of eastern China (e.g., Yang et al., 1993; Yang and Jahn, 2000; Zhang et al., 2000). The observation that the orogenic garnet peridotite is exclusively associated with zones of continental collision, but is absent in zones of oceanic lithosphere subduction suggests its genetic affinity with continental subduction and subsequent continental collision. Three possible origins were

summarized for garnet peridotites (Medaris and Carswell, 1990; Liou and Carswell, 2000): (1) derivation from subducted lithospheric mantle; (2) derivation from the mantle wedge above subduction zones; and (3) subduction zone metamorphism of peridotites previously emplaced into the crust.

The Luliangshan garnet peridotite massif is the largest ultramafic body in the North Qaidam ultrahighpressure (UHP) metamorphic belt, a 400-km-long continental collision belt of the Early Paleozoic at the northern edge of the Tibetan Plateau (Fig. 1). Mineral exsolution lamellae, including rutile + two pyroxene + sodic amphibole in garnet and ilmenite + Al-chromite in olivine (Song et al., 2004, 2005a), and diamond inclusion in zircon crystal (Song et al., 2005b) suggest that this massif has been undergone ultrahigh-pressure metamorphism at depths greater than 200 km. In this paper, we present and discuss in detail the petrological and geochemical data on the garnet peridotite massif. The data suggest that the protoliths of the peridotite include subarc lithospheric harzburgite and a cumulate assemblage precipitated from mantle wedge-derived melts. The garnet peridotite massif thus provides insights into the process of mantle wedge material recycling in response to oceanic lithosphere subduction and its ultimate exhumation as a result of continental collision.

2. Geological setting

The North Qaidam UHP belt extends from Dulan northwestward to Yuka for about 400 km, and is offset for up to 400 km by the Altyn Tagh Fault, a large NE-striking sinistral strike-slip fault in western China (Fig. 1). On both sides of the fault are blocks with Late Precambrian crystalline basement rocks of similar ages (e.g., Wan et al., 2001; Song et al., 2006). The Qilian Terrain is bound further to the north by an Early Paleozoic Pacific-type (i.e., Ernst, 2001) or oceanic-type (i.e., Song et al., 2006) suture with well-exposed blueschists and ophiolites (Wu et al., 1993).



Fig. 1. (a) Geological map showing the locality of the Luliangshan garnet peridotite body. (b) Geological map of the garnet peridotite (after Song et al., 2004).

Rock assemblage in the North Qaidam UHP belt consists predominantly of granitic and pelitic gneisses intercalated with minor blocks of eclogite and, rarely, garnet peridotite. Coesite has been identified as inclusions in zircon separates from paragneiss at Dulan (Yang et al., 2002; Song et al., 2003a); its occurrence and P-T estimates of the enclosing eclogite and garnet peridotite establish the North Qaidam–Altun eclogite belt as an UHP terrane exhumed from depths in excess of 100 km (Song et al., 2003a,b).



Fig. 2. Photographs showing field occurrences of various rock types of the Luliangshan garnet peridotite massif. (a) Garnet lherzolite with layered structure; (b) garnet peridotite interlayered with garnet–olivine pyroxenites, note that the rhythmic variation of garnet content in the peridotite layers; (c) garnet-free dunite and garnet peridotite; (d) garnet lherzolite with subsequent foliations along which olivines were serpentinized; (e) garnet lherzolite interlayered with garnet dunite, olivine in dunite is strongly serpentinized and shows dark colored layers; (f) garnet pyroxenite dyke cutting garnet peridotite.

These characteristics suggest that the North Qaidam is a typical continental-type subduction/exhumation belt and is comparable to those well-studied UHP belts/ terranes in the Dabie–Sulu of eastern China and the West Gneiss Region of Norway. Geochemical data reveal that protoliths of most eclogites in North Qaidam UHP belt possess characteristics of depleted and enriched mid-ocean ridge basalts (i.e., N- and E-MORB) and ocean island basalts (OIB) (Song et al., 2003b, 2006).



Fig. 3. Photomicrographs of various rock types from the Luliangshan garnet peridotite. (a) Sample 2C40. Strongly serpentinized dunite interlayering with garnet lherzolite. (b) Sample 2C40. Garnet lherzolite with cumulate texture showing intercumulus orthopyroxene. (c) Sample 2C42. Garnet lherzolite with cumulate texture showing intercumulus orthopyroxene. (d) Sample 2C26. Garnet-bearing dunite with triple-junction texture. (e) Sample 2C15. Garnet-free dunite with Cr-spinel and orthopyroxene. (f) Sample 2C13. Garnet pyroxenite with assemblage of garnet, two pyroxenes and phlogopite.

3. Petrography

The Luliangshan garnet peridotite is the largest peridotite outcrop recognized so far along the entire Qaidam UHP belt. It occurs as lenses with a total exposed area of 500 m×1000 m within an eclogite-bearing quartzofeldspathic gneiss terrane (Fig. 1b). This garnet peridotite massif comprises a wide range of lithologies from rocks dominated by olivine to those dominated by pyroxene. On the basis of field and petrographic observations, we group the rocks into four types: (1) mostly garnet lherzolite with minor amounts of (2) garnet-bearing dunite, (3) garnet-free dunite and (4) garnet pyroxenite dikes/dikelets. No eclogite was found to be in direct contact with the peridotite. Photographs in Fig. 2 show field occurrences/relationships of the four rock types.

3.1. Garnet lherzolite

The garnet lherzolite is about 70–80 vol.% of the peridotite massif. Fig. 2a to c show that all peridotites in the field occur as layers with various thickness (from meters to centimeters) and sharp boundaries. As

pyroxenes vary in abundance on various scales, it could be termed as garnet lherzolite or olivinewebsterite (see below). Garnet lherzolite and dunite display apparent interlayering on different scales determined by modal variation of olivine and pyroxene (Figs. 2a and b, 3a). Such layering could well be inherited from primary igneous layering such as seen in layered cumulates but could also be of metamorphic origin.

Much of the garnet lherzolite shows massive coarsegrained inequigranular to granoblastic textures. Some of the garnet lherzolite and garnet–olivine pyroxenite display cumulate-like textures with orthopyroxene occurring as intercumulus between olivines (Fig. 3b and c). Serpentinization (Fig. 3 and also dark layers in Fig. 2d–e) apparently represents a rather late hydrothermal metamorphic event. The major minerals are garnet, olivine, orthopyroxene (Opx), clinopyroxene (Cpx) and minor Cr-rich spinel (Spl). Garnets are mostly porphyroblastic and vary in size (3 to 10 mm) and abundance (~5 to 15 vol.%). Olivine shows a wide range of Fo values (i.e., $Mg^{\#}$, $Mg/[Mg+Fe^{2+}]=0.84-$ 0.91) and constitutes 40–60 vol.% of the rock. Opx and Cpx take ~10–30 and 5–15 vol.% of the rock,



Fig. 4. Compositional variation of garnets from different rocks from the Luliangshan garnet peridotite.

respectively. Some samples show apparent layering defined by varying olivine and pyroxene proportions on local scales. Amphiboles are retrograde products overprinting garnet and pyroxenes. Fine-grained Cr-rich spinels ($Cr^{\#}$, [Cr/Cr+Al]=0.60–0.69) occur either as inclusions of other minerals or as a decompression phase of previous high-Cr garnet and pyroxenes.

High concentrations of decompression-induced exsolution textures have been observed in some porphyroblastic garnets and olivines in the lherzolite. Exsolution lamellae in garnets include densely packed rods of rutile, orthopyroxene, clinopyroxene and sodic amphibole (Song et al., 2004, 2005a). Exsolutions in olivine are needles of ilmenite and chromian spinel (Song et al., 2004). The pyroxene exsolutions suggest that their parental garnet host crystals originally possessed excess silicon, i.e., they were majoritic garnets that are only stable at depths well in excess of 150 km (Ringwood and Major, 1971; Irifune, 1987). Similar textures were also described in garnets from the Norwegian peridotite (Van Roermund and Drury, 1998). The exsolution of rutile and sodic amphiboles (Song et al., 2005a) further suggests that the inferred majoritic garnets also contain excess Ti, Na and hydroxyl that are only soluble at very high pressures. The exsolution of ilmenite and Alchromite needles from olivines is also consistent with the peridotite once being equilibrated at very high pressures (>300 km) (Dobrzhinetskaya et al., 1996). Estimation of Ti and Na contents in the inferred majoritic garnets gives an equilibrium pressure of \geq 7 GPa (Song et al., 2004, 2005a). Application of Alin-Opx geothermobarometry (Nickel and Green, 1985; Brey and Köhler, 1990) and Grt-Ol geothermometry (O'Neill et al., 1979) to matrix minerals yields P=5.0-6.5 GPa and T=960-1040 °C (Fig. 4a and b).

3.2. Garnet-bearing dunite

The garnet-bearing dunite occurs either as layers varying in thickness from 10 cm to up to 2 m within the garnet lherzolite or as rhythmic bands that vary gradually from garnet-bearing dunite to harzburgite to

Table 1

Average compositions of garnet from Luliangshan garnet peridotite massif

Sample	2C13	2C14	2C26	2C27	2C28	2C34	2C37	2C39	2C41	2C42	2C44	2C50	L382
	G-P	G-P	G-P	G-D	G-L	G-L	G-D	G-D	G-L	G-L	G-L	G-L	G-L
	<i>n</i> =3	<i>n</i> =6	<i>n</i> =3	<i>n</i> =3	n=4	<i>n</i> =2	<i>n</i> =2	n=5	<i>n</i> =3	n=21	<i>n</i> =3	<i>n</i> =2	<i>n</i> =3
SiO ₂	42.27	41.70	41.59	41.77	41.45	41.19	42.59	42.68	40.76	42.42	41.55	41.34	41.23
TiO ₂	0.00	0.04	0.03	0.05	0.09	0.04	0.09	0.06	0.07	0.16	0.04	0.08	0.10
Al_2O_3	23.58	23.50	21.74	22.64	22.74	22.72	22.49	22.34	22.24	22.28	22.79	21.83	22.19
Cr_2O_3	0.35	0.43	2.24	1.38	0.84	1.07	2.13	2.02	1.71	1.46	0.91	1.82	1.68
FeO	10.40	11.68	9.33	8.55	10.13	11.92	7.57	7.52	12.99	9.72	12.82	13.89	12.53
MnO	0.26	0.31	0.61	0.35	0.49	0.90	0.38	0.51	0.54	0.42	0.41	0.60	0.43
MgO	18.53	18.58	19.43	20.43	19.53	17.65	20.45	21.09	17.72	19.63	17.78	16.91	17.41
CaO	4.50	4.40	4.25	4.86	4.15	4.12	4.90	4.33	4.75	4.32	4.78	5.12	4.90
Na ₂ O	0.00	0.01	0.02	0.03	0.01	0.03	0.02	0.01	0.00	0.04	0.02	0.01	0.02
K ₂ O	0.00	0.00	0.00	0.00	0.01	0.00	0.00	0.00	0.00	0.01	0.00	0.00	0.00
Total	99.68	100.64	99.21	100.05	99.42	99.62	100.61	100.54	100.77	100.49	101.07	101.59	100.48
Si	3.019	2.974	3.003	2.965	2.976	2.999	3.006	3.010	2.933	3.018	2.972	2.970	2.976
Ti	0.000	0.002	0.001	0.003	0.005	0.002	0.005	0.003	0.004	0.009	0.002	0.004	0.006
Al	1.985	1.973	1.849	1.893	1.923	1.900	1.871	1.855	1.885	1.869	1.920	1.846	1.886
Cr	0.020	0.024	0.128	0.077	0.048	0.067	0.119	0.113	0.097	0.082	0.051	0.103	0.096
Fe ²⁺	0.621	0.697	0.563	0.507	0.608	0.733	0.447	0.442	0.782	0.578	0.767	0.835	0.756
Mn	0.016	0.019	0.037	0.021	0.030	0.055	0.023	0.030	0.033	0.026	0.025	0.037	0.026
Mg	1.973	1.975	2.091	2.162	2.090	1.904	2.152	2.219	1.901	2.082	1.896	1.811	1.874
Ca	0.344	0.336	0.329	0.370	0.319	0.339	0.371	0.327	0.367	0.330	0.366	0.394	0.379
Na	0.000	0.001	0.002	0.004	0.002	0.004	0.003	0.002	0.000	0.005	0.002	0.002	0.003
K	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.001	0.000	0.000	0.000
Sum	7.978	8.00	8.00	8.00	8.00	8.00	7.996	8.00	8.00	8.00	8.00	8.00	8.00
Alm	21.02	23.01	18.63	16.56	19.95	24.14	14.93	14.65	25.37	19.17	25.09	27.12	24.90
Gross	10.67	9.89	4.41	8.06	8.01	7.73	6.43	5.09	6.83	6.75	9.36	7.45	7.59
Pyrope	66.79	65.24	69.20	70.57	68.56	62.77	71.92	73.46	61.68	69.05	62.04	58.85	61.68
Spess	0.53	0.62	1.24	0.69	0.98	1.82	0.76	1.01	1.07	0.85	0.80	1.19	0.86
Uvaro	0.99	1.22	6.45	4.00	2.45	3.43	5.96	5.72	5.05	4.18	2.63	5.36	4.88

G-P: garnet pyroxenite; G-D: garnet-bearing dunite; G-L: garnet lherzolite.

Sample	2C15	2C26	2C27	2C28a	2C28b	2C34	2C36	2C37	2C39	2C41	2C42	2C44	2C50	L361	L382
Rock	D	G-D	G-D	G-L	G-D	G-L	G-D	G-D	G-D	G-L	G-L	G-L	G-L	G-L	G-L
Number	<i>n</i> =6	<i>n</i> =3	<i>n</i> =3	n=7	n=7	<i>n</i> =3	<i>n</i> =3	n=3	<i>n</i> =3	<i>n</i> =3	<i>n</i> =4	<i>n</i> =3	<i>n</i> =3	<i>n</i> =3	n=2
SiO_2	40.93	40.43	40.49	40.85	40.71	39.37	40.25	40.94	41.10	39.56	40.48	39.56	39.56	40.06	39.78
TiO ₂	0.00	0.00	0.00	0.02	0.01	0.01	0.00	0.00	0.02	0.02	0.01	0.03	0.01	0.02	0.04
Al_2O_3	0.01	0.00	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.02	0.01	0.01	0.00	0.00	0.06
Cr_2O_3	0.02	0.02	0.01	0.03	0.01	0.02	0.00	0.00	0.00	0.00	0.02	0.01	0.00	0.02	0.33
FeO	6.82	8.12	8.62	10.00	8.88	16.21	7.63	7.32	8.03	12.96	9.53	14.89	15.03	13.04	13.41
MnO	0.08	0.14	0.09	0.16	0.10	0.12	0.10	0.10	0.09	0.12	0.12	0.23	0.11	0.14	0.18
NiO	0.49	0.48	0.48	0.32	0.35	0.51	0.46	0.37	0.38	0.52	0.38	0.42	0.43	0.47	0.49
MgO	52.49	51.25	51.23	49.14	49.64	45.09	52.32	51.19	51.69	47.90	50.06	45.74	45.84	47.04	46.52
CaO	0.01	0.01	0.02	0.01	0.02	0.00	0.00	0.00	0.01	0.00	0.01	0.01	0.01	0.03	0.02
Na ₂ O	0.01	0.00	0.00	0.01	0.01	0.02	0.00	0.01	0.00	0.02	0.02	0.01	0.00	0.02	0.03
K ₂ O	0.00	0.01	0.00	0.01	0.01	0.01	0.00	0.01	0.01	0.00	0.00	0.00	0.01	0.01	0.03
Total	100.86	100.46	100.95	100.55	99.74	101.37	100.76	99.93	101.31	101.12	100.62	100.89	100.99	100.82	100.86
Si	0.984	0.983	0.981	0.997	0.997	0.984	0.974	0.994	0.988	0.977	0.987	0.987	0.986	0.991	0.987
Ti	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.001	0.000	0.000	0.001
Al	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.001	0.000	0.000	0.000	0.000	0.002
Cr	0.000	0.000	0.000	0.001	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.006
Fe	0.137	0.165	0.175	0.204	0.182	0.339	0.154	0.149	0.161	0.268	0.194	0.311	0.313	0.270	0.278
Mn	0.002	0.003	0.002	0.003	0.002	0.003	0.002	0.002	0.002	0.002	0.003	0.005	0.002	0.003	0.004
Ni	0.009	0.009	0.009	0.006	0.007	0.010	0.009	0.007	0.007	0.010	0.007	0.008	0.009	0.009	0.010
Mg	1.882	1.857	1.851	1.789	1.813	1.679	1.887	1.853	1.853	1.764	1.820	1.701	1.703	1.734	1.720
Ca	0.000	0.000	0.000	0.000	0.001	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.001	0.001
Na	0.001	0.000	0.000	0.000	0.000	0.001	0.000	0.000	0.000	0.001	0.001	0.000	0.000	0.001	0.001
Κ	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.001
Sum	3.016	3.017	3.019	3.002	3.003	3.016	3.026	3.006	3.012	3.023	3.013	3.013	3.014	3.009	3.010
Mg [#]	0.932	0.918	0.914	0.898	0.909	0.832	0.924	0.926	0.920	0.868	0.904	0.846	0.845	0.865	0.861

Table 2 Average compositions of olivines from Luliangshan garnet peridotite massif

D: garnet-free dunite; G-D: garnet-bearing dunite; G-L: garnet lherzolite.

Table 3 Average compositions of orthopyroxene from Luliangshan garnet peridotite massif

-	*				•									
Sample	2C12	2C13	2C14	2C26	2C28	2C36	2C34	2C37	2C39	2C41	2C42	2C44	2C50	L382
	G-P	G-P	G-P	G-D	G-L	G-D	G-L	G-D	G-D	G-L	G-L	G-L	G-L	G-L
	<i>n</i> =3	<i>n</i> =3	<i>n</i> =3	<i>n</i> =2	<i>n</i> =3	<i>n</i> =3	<i>n</i> =5	<i>n</i> =3	n=2	<i>n</i> =2				
SiO ₂	57.16	57.33	56.61	57.14	57.04	57.37	56.07	57.37	57.18	56.20	56.99	57.03	56.82	56.94
TiO ₂	0.00	0.02	0.02	0.07	0.05	0.03	0.03	0.00	0.00	0.08	0.05	0.03	0.03	0.01
Al_2O_3	0.97	0.99	1.07	0.68	0.59	0.64	0.56	0.67	0.63	0.60	0.46	0.58	0.61	0.64
Cr_2O_3	0.04	0.07	0.04	0.12	0.11	0.10	0.14	0.20	0.12	0.15	0.23	0.10	0.14	0.18
FeO	8.61	8.24	9.15	5.46	6.01	4.94	10.18	4.91	5.58	8.32	6.04	8.92	9.34	8.33
MnO	0.11	0.15	0.11	0.15	0.14	0.14	0.14	0.11	0.14	0.13	0.14	0.17	0.13	0.10
NiO	0.11	0.12	0.13	0.17	0.08	0.04	0.14	0.06	0.07	0.13	0.07	0.11	0.11	0.08
MgO	32.89	33.45	33.30	36.47	35.32	36.65	32.39	35.83	35.68	34.36	35.07	33.88	33.88	33.71
CaO	0.25	0.22	0.21	0.16	0.22	0.17	0.25	0.20	0.20	0.24	0.26	0.22	0.22	0.24
Na ₂ O	0.00	0.01	0.01	0.01	0.04	0.01	0.00	0.01	0.01	0.02	0.02	0.00	0.01	0.01
K_2O	0.00	0.00	0.03	0.00	0.00	0.00	0.00	0.00	0.01	0.01	0.01	0.03	0.01	0.00
Total	100.13	100.60	100.67	100.52	99.59	100.08	99.98	99.45	99.63	100.41	99.52	101.06	101.27	100.27
Si	1.988	1.980	1.957	1.944	1.963	1.956	1.965	1.975	1.968	1.942	1.973	1.962	1.952	1.972
Ti	0.000	0.001	0.000	0.002	0.001	0.001	0.001	0.000	0.000	0.002	0.001	0.001	0.001	0.000
Al	0.040	0.040	0.044	0.029	0.032	0.026	0.023	0.027	0.026	0.024	0.019	0.023	0.025	0.026
Cr	0.001	0.002	0.001	0.003	0.003	0.003	0.004	0.005	0.003	0.004	0.006	0.003	0.004	0.005
Fe3	0.000	0.000	0.041	0.075	0.040	0.059	0.042	0.019	0.036	0.086	0.029	0.050	0.067	0.026
Fe2	0.251	0.238	0.223	0.080	0.133	0.082	0.256	0.122	0.124	0.154	0.146	0.207	0.201	0.215
Mn	0.003	0.004	0.003	0.004	0.004	0.004	0.004	0.003	0.004	0.004	0.004	0.005	0.004	0.003
Ni	0.003	0.003	0.004	0.005	0.002	0.001	0.004	0.002	0.002	0.004	0.002	0.003	0.003	0.002
Mg	1.705	1.722	1.716	1.850	1.811	1.862	1.692	1.839	1.828	1.769	1.809	1.737	1.735	1.740
Ca	0.009	0.008	0.008	0.006	0.008	0.006	0.009	0.007	0.007	0.009	0.009	0.008	0.008	0.009
Na	0.000	0.000	0.000	0.000	0.003	0.000	0.000	0.001	0.000	0.001	0.001	0.000	0.001	0.001
Κ	0.000	0.000	0.001	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.001	0.000	0.000
Mg [#]	86.48	87.72	88.48	95.85	93.18	95.80	86.85	93.76	93.73	91.99	92.53	89.37	89.62	89.00

G-P: garnet pyroxenite; G-D: garnet-bearing dunite; G-L: garnet lherzolite.

garnet lherzolite (Fig. 2b). Garnet content varies widely in different layers. The rock is medium-grained and has an equigranular texture (Fig. 3d) dominated by olivine (Fo_{90.6–92.0}) (>90 vol.%), plus variable amounts of garnet, orthopyroxene (Mg[#]=0.90–0.92), and clinopyroxene (Mg[#]=0.94–0.96). Garnets are porphyroblastic and Mg-rich (69–75 mol% pyrope, 11–18% almandine, 3–8% grossular, 0.8–2.0% spessartine and 3–6% uvarovite). Fine-grained Cr-spinel (Cr[#]=0.61–0.65) occurs as product of decompression-induced breakdown of previous high-Cr garnet and pyroxenes. Al-in-Opx geobarometry of Brey and Köhler (1990) and Grt-Ol geothermometry of O'Neill et al. (1979) yield *P–T* conditions of *P*=4.6–5.3 GPa and *T*=980–1130 °C (e.g., Fig. 4c).

3.3. Garnet-free dunite

Garnet-free dunite occurs apparently as layers and/or lenses varying in thickness of up to 4 m with browncolored weathering surfaces distinguished from blackcolored garnet lherzolite and garnet-bearing dunite (Fig. 2c). It is medium-grained, dominated by strongly serpentinized olivine (>90 vol.%) with minor orthopyroxene and Cr-rich spinel (Fig. 3e). Magnetites are precipitated at the boundaries of these serpentinized olivine crystals and show a triple-junction texture. The olivine is Mg-rich (Fo_{92.4–93.7}) and the spinel is Cr-rich (i.e., $Cr^{\#}=0.66-0.73$) relative to those in the garnet lherzolite.

3.4. Garnet pyroxenite

As shown in Figs. 1c and 2f, the garnet pyroxenite, as a minor component, occurs as 2- to 5-m-thick dikes cross-cutting the apparent layering of the massif. Most samples are fresh with pink garnet and pale-green pyroxene conspicuous in the field. The constituent phases are garnet (20–30 vol.%), orthopyroxene (5– 10%), clinopyroxene (40–60%) and phlogopite (2–5%) without olivine. It shows a fairly uniform mediumgrained granular texture (Fig. 3f). Most garnets are rimmed with a kelyphitic Opx+Cpx+Spl assemblage and some break down to granular-textured high-Al Opx, Cpx and Al-Spl, which are characteristic decomposition features. These observations suggest that these rocks have also once equilibrated at high pressures. Their occurrences as dikelets, lack of olivine, low MgO (~18%), low Mg[#] (~0.81), and the presence of minor phlogopite are all consistent with the garnet pyroxenite being cumulate from more evolved mantle wedge melts (see below). The Al-in-Opx geobarometer and two pyroxene thermometer (Brey and Köhler, 1990) yield P=2.6-3.0 GPa and T=800-900 °C (Fig. 4d), which are much lower than that of garnet peridotite.

4. Mineral chemistry

Mineral analyses were done on a JEOL JXA-8100 Electron Probe Microanalyzer (EPMA) at Peking University. Analytical conditions were optimized for standard silicates and oxides at 15 kV accelerating voltage with a 20 nA focused beam current for all the elements. Routine analyses were obtained by counting for 30 s at peak and 10 s on background. Repeated analysis of natural and synthetic mineral standards yielded precisions better than $\pm 2\%$ for most elements. Representative electron microprobe analyses of major minerals are given in Tables 1–4.

4.1. Garnet

Most garnet crystals in garnet peridotite and garnetbearing dunite are porphyroblasts with varying size (3-10 mm across): those in the matrix are fine-grained and equigranular. Almost all the garnet crystals exhibit kelyphitized rims of clinopyroxene, orthopyroxene and spinel aggregates interpreted as resulting from decompression. In some samples, garnet is completely replaced by the kelyphitic Opx + Cpx + Spl. Pargasitic amphiboles appear at the outer circle of the kelyphite, suggesting a late retrograde metamorphic event. EPMA analyses (Table 1) show that some garnet crystals in garnet peridotite contain relatively high TiO₂ (up to 0.16 wt.%) (Table 1). Pyrope content of garnets increases from 60.1 mol% in lherzolite to 74.5 mol% in garnet-bearing dunite, and the almandine decreases from 25.7 mol% to 11.6 mol%, with relatively narrow

Table 4

Average compositions of clinopyroxene and spinel from Luliangshan garnet peridotite massif

Sample	2C12	2c13	2C28	2C44	2C42	2C42	2C37	2C39	2C15	2C41	2C36	2C26	2C37	2C39
Mineral	Срх	Срх	Срх	Cpx*	Срх	Cpx*	Cpx*	Cpx*	Sp	Sp	Sp	Sp	Sp-L	Sp-L
	<i>n</i> =2	<i>n</i> =3	n=2	<i>n</i> =2	<i>n</i> =3	<i>n</i> =3	<i>n</i> =2	<i>n</i> =3						
SiO ₂	54.73	55.37	54.19	54.67	54.94	54.05	53.89	54.25	0.03	0.00	0.07	0.00	0.10	0.03
TiO ₂	0.01	0.03	0.24	0.01	0.20	0.05	0.05	0.12	0.03	0.64	0.08	0.08	0.08	0.20
Al_2O_3	2.58	1.98	2.23	1.39	3.10	1.33	1.11	1.52	13.97	19.63	23.85	24.88	20.15	19.47
Cr ₂ O ₃	0.20	0.28	0.63	0.72	1.25	0.80	0.64	1.39	56.31	44.74	44.12	43.51	47.36	47.05
FeO	2.09	1.82	1.75	2.53	1.65	1.83	1.35	1.62	19.52	28.65	20.67	20.68	11.89	9.18
MnO	0.00	0.08	0.00	0.05	0.07	0.06	0.03	0.09	0.62	0.61	0.60	0.54	0.47	0.00
NiO	0.03	0.10	0.03	0.04	0.05	0.05	0.01	0.06	0.01	0.12	0.00	0.01	0.66	0.57
MgO	16.74	16.46	16.73	16.89	15.89	17.24	17.77	17.11	10.63	6.12	11.01	10.94	16.80	22.26
CaO	21.93	22.48	23.49	22.71	21.50	23.73	24.15	23.41	0.00	0.00	0.01	0.06	0.05	0.06
Na ₂ O	1.29	0.99	0.91	0.94	1.34	0.62	0.42	0.83	0.00	0.02	0.07	0.05	0.00	0.03
K ₂ O	0.01	0.01	0.00	0.00	0.00	0.00	0.00	0.01	0.01	0.00	0.00	0.00	0.00	0.01
Total	99.60	99.60	100.20	99.94	99.98	99.77	99.41	100.41	101.13	100.51	100.48	100.74	97.56	98.85
Si	1.978	2.003	1.955	1.981	1.978	1.961	1.96	1.96	0.001	0.000	0.002	0.000	0.003	0.001
Ti	0.000	0.001	0.007	0.000	0.005	0.001	0.00	0.00	0.001	0.015	0.002	0.002	0.002	0.004
Al	0.110	0.084	0.095	0.059	0.132	0.057	0.05	0.06	0.527	0.749	0.865	0.898	0.732	0.676
Cr	0.006	0.008	0.018	0.020	0.035	0.023	0.02	0.04	1.425	1.145	1.074	1.053	1.154	1.096
Fe3	0.018	0.000	0.027	0.024	0.000	0.018	0.04	0.03	0.045	0.077	0.058	0.049	0.104	0.219
Fe2	0.045	0.055	0.026	0.053	0.050	0.037	0.04	0.01	0.477	0.699	0.474	0.481	0.203	0.008
Mn	0.000	0.002	0.000	0.002	0.002	0.002	0.00	0.00	0.017	0.017	0.016	0.014	0.012	0.000
Ni	0.001	0.003	0.001	0.001	0.002	0.001	0.00	0.00	0.000	0.003	0.000	0.000	0.016	0.014
Mg	0.902	0.888	0.900	0.912	0.853	0.933	0.96	0.92	0.507	0.295	0.505	0.499	0.772	0.978
Ca	0.849	0.871	0.908	0.881	0.829	0.923	0.94	0.90	0.000	0.000	0.000	0.002	0.002	0.002
Na	0.091	0.069	0.064	0.066	0.094	0.044	0.03	0.06	0.000	0.001	0.004	0.003	0.000	0.002
Κ	0.000	0.000	0.000	0.000	0.000	0.000	0.00	0.00	0.000	0.000	0.000	0.000	0.000	0.000
	4.000	3.985	4.000	4.000	3.980	4.000	4.00	4.00	3	3.00	3	3	3	3
$Mg^{\#}$	0.95	0.94	0.97	0.95	0.94	0.96	0.96	0.98	49.3	27.54	48.7	48.5	71.6	81.2
Cr [#]									73.0	60.46	55.4	54.0	61.2	61.8

Cpx*: clinopyroxene with amphibole + quartz (or spinel) lamellae. Sp-L: spinel lamella in clinopyroxene.



Fig. 5. Compositional profiles of garnets from garnet lherzolite samples 2C42 (a) and 2C44 (b).

ranges of grossular + and radite (5-10%), spessartine (0.9-1.8%) and uvarovite (2-5%).

Porphyroblastic garnet crystals in all rock types are rather homogeneous with no obvious compositional zoning. As shown in Fig. 5, all oxides in garnet crystals from lherzolite samples 2C42 and 2C44 display flat patterns from core to rim; sharp changes of MgO and FeO only appear at the two edges in garnet of 2C42 (Fig. 5a), which is affected by the subsequent kelyphitisation.

4.2. Olivine

Olivine shows a large compositional variation in dunite and garnet peridotite (Table 2 and Fig. 6). Olivine



Fig. 6. Fo (=100[Mg/(Mg + Fe)]) vs. NiO content of olivines from the Luliangshan garnet peridotite.

in garnet-free dunite (2C15) has the highest Fo content of 0.927–0.937, which is slightly higher than that of garnet-bearing dunite (0.906–0.926). Fo content of olivine in garnet lherzolite, on the other hand, varies from 0.830 to 0.906 with a compositional gap between 0.871 and 0.895, and shows positive correlation with the modal percentage of olivine. NiO concentrations vary from 0.34 to 0.55 wt.% and have no significant difference among garnet lherzolite, garnet-bearing dunite and garnet-free dunite.

Ilmenite and Al-chromite rod- and needle-shaped exsolutions have been observed in some large olivine crystals, and are interpreted as resulting from decompression of the same phase previously equilibrated at great depths (Song et al., 2004).

4.3. Orthopyroxene

Orthopyroxene (Opx) in garnet peridotite, garnet– dunite and garnet pyroxenite show a large range of compositional variation (Table 3). Opx from garnetbearing dunite samples 2C26, 2C36, 2C37 and 2C39 has high $mg^{\#}$ (0.94–0.96) whereas Opx from garnet lherzolite has significantly lower $mg^{\#}$ (0.87–0.93). Opx $Mg^{\#}$ values are positively correlated with the Fo content of coexisting olivine.

All Opx crystals in garnet lherzolite and garnetbearing dunite are extremely Al-poor. Al₂O₃ contents in Opx from lherzolite are in the range of 0.38-0.66 wt.%, and from dunite, slightly higher in 0.55-0.73 wt.%. Opx crystals from garnet pyroxenite differ in composition



Fig. 7. mg-number vs. Al₂O₃ content of orthopyroxenes in Luliangshan garnet peridotite.

from those in garnet lherzolite and garnet-bearing dunite; they have lower Mg[#] (0.86–0.90) and higher Al₂O₃ (0.88–1.17 wt.%). Fig. 7 shows the co-variation of Opx Mg[#] and Al₂O₃ in the three type rocks.

4.4. Clinopyroxene

All clinopyroxene (Cpx) crystals in garnet lherzolite, garnet-bearing dunite and garnet pyroxenite are rich in diopside. Cr_2O_3 contents in Cpx from garnet lherzolite and garnet-bearing dunite vary from 0.6 to 1.6 wt.% (Table 4). However, those in garnet pyroxenite are Crpoor (Cr_2O_3 0.18–0.30 wt.%). Cpx Mg[#] varies from 0.901 in the garnet-bearing dunite to 0.916 in the garnet lherzolite and correlate positively with Mg[#] of olivine and Opx. Cpx in the garnet–phlogopite pyroxenites shows high yet a narrow range of Mg[#] (0.93–0.95). Al₂O₃ contents vary from 1.13 to 3.19 wt.%, and Na₂O from 0.43 to 1.32 wt.%. There is no Al₂O₃ correlation between Opx and Cpx.

High concentrations of decompression-induced exsolution products are also observed in some porphyroblastic clinopyroxene crystals from all garnet peridotite and garnet pyroxenite (Fig. 8). Fig. 8 shows that the exsolution lamellae in Cpx are amphiboles, quartz



Fig. 8. Photomicrographs showing exsolution lamellae in clinopyroxene from the Luliangshan garnet peridotite. (a) Amphibole + quartz lamellae in clinopyroxene in garnet pyroxenite (sample 2C14); (b) amphibole + Al-chromite lamellae in clinopyroxene in garnet lherzolite (Sample 2C27); (c) BSE image of amphibole + quartz lamellae (perpendicular to [100] of the host clinopyroxene) (sample 2C14); (d) BSE image of panel b.

Table 5
Whole-rock major and trace element abundance of Luliangshan garnet peridotite massif

Sample	2C11	2C12	2C13	2C15	2C26	2C27	2C34	2C36	2C37	2C39	2C41	2C42
	G-P	G-P	G-P	D	G-D	G-L	G-L	G-D	G-D	G-D	G-L	G-L
Maior e	lements (wi	t.%)										
SiO ₂	47.23	47.54	46.53	38.52	40.81	41.46	45.85	38.90	40.47	41.56	45.97	42.25
TiO ₂	0.05	0.05	0.05	0.00	0.04	0.04	0.07	0.02	0.02	0.01	0.06	0.17
Al ₂ O ₃	14.70	13.62	15.66	0.56	1.64	3.71	3.59	1.18	1.89	1.84	3.80	3.60
TFeO	7.37	7.40	7.57	6.10	6.43	6.77	12.22	6.44	6.86	6.52	9.64	7.60
MnO	0.17	0.17	0.17	0.09	0.10	0.11	0.16	0.10	0.11	0.09	0.14	0.12
MgO	17.78	17.64	18.11	41.62	39.67	35.26	32.01	40.66	39.72	39.46	33.35	35.14
CaO	10.72	10.52	9.24	0.14	1.13	2.70	2.81	0.35	1.20	1.57	2.23	2.76
Na ₂ O	0.56	0.77	0.59	0.08	0.28	0.10	0.12	0.06	0.06	0.13	0.18	0.22
K ₂ O	0.55	0.51	0.95	0.02	0.04	0.17	0.01	0.02	0.07	0.03	0.02	0.07
LÕI	0.54	1.35	0.64	12.50	9.69	9.29	2.83	11.76	9.44	8.39	4.15	7.70
Total	99.68	99.59	99.52	99.65	99.84	99.63	99.68	99.51	99.86	99.61	99.56	99.64
mg-no.	0.811	0.809	0.810	0.924	0.917	0.903	0.824	0.918	0.912	0.915	0.860	0.892
Trace el	ements (pp	m)										
Sc	44.8	34.1	51.6	5.88	9.4	9.1	20.4	6.70	10.2	8.1	21.2	14.6
V	167	167	166	12.4	25.9	49.9	74	20.1	29.1	31.6	85	72
Cr	1700	2157	1858	2605	2145	2341	1690	2505	2139	1725	5660	2290
Co	102	126	84	101	106	104	170	102	107	102	124	112
Ni	359	855	446	2283	2125	1864	1724	2207	2143	2153	1573	1870
Cu	2.47	34.1	3 99	0.72	1 83	2.45	67	2.03	2.71	2.87	13.6	35.0
Zn	18.6	20.0	21.8	33.9	32.3	154	77 7	37.5	33.5	34.6	88	39.4
Ga	5 46	5 17	5 79	0.30	0.82	1 66	2.61	0.67	1.02	1 16	3 37	2.50
Rh	30.63	32.08	49.7	0.59	0.57	7 46	0.41	0.58	2.10	0 375	0.310	1 41
Sr	319	265	330	4 71	20.7	128.1	43.2	6 66	34.0	28.5	16.3	43.6
Y	6.62	5 97	7 18	0.046	1.08	3.07	2.41	0.47	1 32	1 19	1 49	3.04
Zr	5.30	7.77	6.81	0.39	7.29	10.03	4.43	1.49	6.33	1.91	1.91	3.64
Nb	0.29	0.39	0.27	0.21	0.58	0.12	0.35	0.28	0.11	0.26	0.079	0.98
Cs	3.29	3.58	3.87	0.210	0.049	1.13	0.135	0.094	0.447	0.049	0.105	0.280
Ba	224.4	126.2	366.2	2.73	12.78	13.63	9.45	4.46	48.70	13.07	4.78	61.4
La	3.58	3.65	3.11	0.080	0.53	1.01	0.40	0.24	0.48	0.84	0.25	1.24
Ce	9.15	11.25	7.78	0.128	1.46	2.96	0.63	0.54	1.50	2.26	0.52	2.56
Pr	1.21	1.52	1.01	0.015	0.25	0.50	0.103	0.086	0.24	0.31	0.081	0.29
Nd	4.64	5.57	3.86	0.058	1.19	2.34	0.49	0.38	1.07	1.18	0.40	1.00
Sm	0.80	0.87	0.68	0.014	0.26	0.49	0.14	0.10	0.25	0.18	0.12	0.20
Eu	0.25	0.23	0.25	0.004	0.067	0.09	0.057	0.029	0.070	0.045	0.039	0.082
Gd	0.81	0.86	0.76	0.013	0.20	0.39	0.22	0.087	0.21	0.16	0.14	0.31
Tb	0.12	0.12	0.13	0.001	0.029	0.066	0.046	0.012	0.030	0.023	0.030	0.061
Dy	0.87	0.82	0.90	0.006	0.16	0.42	0.33	0.078	0.19	0.15	0.21	0.42
Но	0.23	0.21	0.25	0.001	0.036	0.10	0.08	0.015	0.044	0.038	0.052	0.11
Er	0.69	0.64	0.75	0.002	0.11	0.31	0.24	0.046	0.14	0.13	0.16	0.31
Tm	0.12	0.11	0.13	0.002	0.018	0.053	0.043	0.007	0.026	0.022	0.028	0.051
Yb	0.92	0.84	0.99	0.004	0.14	0.39	0.30	0.065	0.18	0.17	0.19	0.34
Lu	0.16	0.14	0.17	0.002	0.026	0.064	0.049	0.013	0.031	0.028	0.034	0.058
Hf	0.30	0.44	0.30	0.012	0.24	0.23	0.12	0.043	0.14	0.065	0.084	0.15
Та	0.02	0.02	0.017	0.013	0.059	0.007	0.021	0.017	0.007	0.020	0.005	0.067
Pb	32.41	29.46	25.40	0.271	0.612	1.63	1.33	1.99	4.40	0.746	0.912	3.21
Th	0.48	0.326	0.404	0.021	0.050	0.029	0.018	0.018	0.094	0.036	0.022	0.051
U	0.57	0.53	0.66	0.010	0.018	0.272	0.43	0.005	0.058	0.045	0.015	0.046

G-P=garnet pyroxenite; G-D=garnet-bearing dunite; G-L=garnet lherzolite; D=garnet-free dunite.

^a Samples were analyzed by ICP-MS for trace elements at Guangzhou Institute of Geochemistry, CAS, others at Northwest University, Xi'an, China.

(Fig. 8a and c) and Al-chromite (Fig. 8b and d). The lamellae are 50- to 300-µm-long spindle-shaped rods, and are strictly oriented in the [010] direction of the host Cpx. EPMA analysis suggests that the amphibole

exsolutions are edenitic to tremolitic in composition. Photomicrographs and BSE images show that amphibole lamellae in some Cpx crystals can reach up to 8.5 vol.% (Fig. 8a and c). The amount of exsolved

2C44	2C50	2C30 ^a	<u>2C30^a</u> <u>2C40^a</u> <u>2C43^a</u> <u>2C45^a</u> <u>2C46^a</u> <u>2C47^a</u> <u>2C49^a</u> <u>2C58^a</u>		2C59 ^a					
G-L	G-L	G-D	G-D	G-D	D	G-L	G-L	G-L	G-L	G-L
47.60	51.45	38.96	30 74	40.06	40.11	43 52	44 70	46.23	44 41	45 73
0.10	0.10	0.01	0.04	0.03	0.01	0.07	0.07	0.08	0.08	0.09
6.15	4 64	1 22	2 54	1 24	0.37	3.02	3.92	4 74	4 92	4 51
9.35	8.89	6.18	6.87	6 59	5 53	9.58	10.13	10.01	10.10	9.56
0.17	0.18	0.03	0.11	0.09	0.04	0.14	0.15	0.17	0.18	0.17
29.51	24 56	39.04	37.45	40.20	38.10	34 58	33.62	31.16	32 31	31.59
3.63	8.82	0.24	1 46	0.89	0.05	1.88	2 77	3 34	3 20	3 46
0.19	0.37	0.05	0.08	0.09	0.05	0.12	0.16	0.16	0.13	0.14
0.03	0.01	0.03	0.03	0.05	0.00	0.02	0.03	0.10	0.02	0.02
2.85	0.53	13.60	11 19	10.65	15 73	6.58	4 13	3 56	4 49	4 30
00 50	99.55	99.70	00.88	100.19	100.37	100.27	100.42	100.25	100.52	100.20
0.849	0.831	0.918	0.907	0.916	0.925	0.865	0.855	0.847	0.851	0.855
29.8	32.1	8.17	14.35	7.671	3.778	21.92	21.44	27.69	24.97	27.11
114	172	26.4	43.8	22.0	14.7	93.2	94.1	115	103	123
4794	2855	2442	2487	2001	2359	5104	4975	5120	4560	4836
112	126	84.04	90.48	105.4	93.67	113	112.8	107.1	106.3	101.5
1307	1049	2188	1876	2446	2438	1593	1472	1485	1522	1351
47.9	58.3	1.579	13.99	4.619	2.075	10.13	17.45	16.74	15.38	40.18
61.2	44.4	34.29	35.11	40.04	41.42	60.48	64.91	55.4	53.38	52.86
4.39	4.32	0.672	1.398	0.803	0.603	3.611	3.968	4.429	3.967	4.607
0.70	0.317	0.062	0.155	0.79	0.322	2.085	0.196	0.321	0.35	1.228
26.5	79.2	3.946	23.87	16.65	13.81	18.91	19.99	24.6	28.84	42.18
2.22	5.04	0.526	2.135	0.745	0.331	1.175	2.009	2.849	2.979	1.829
3.49	4.65	0.369	2.232	2.298	1.017	2.987	2.983	3.754	3.465	2.624
0.14	0.41	0.085	0.413	0.557	0.161	0.121	0.092	0.137	0.124	0.122
0.257	0.239	0.008	0.068	0.156	0.153	0.401	0.096	0.133	0.17	0.225
6.38	2.70	1.208	11.96	8.911	4.154	7.756	3.675	4.767	3.063	13.73
0.34	1.10	0.185	0.781	0.389	0.231	0.275	0.311	0.389	0.328	0.331
0.70	2.43	0.637	2.572	1.1	0.46	0.635	0.692	0.897	0.754	0.75
0.113	0.34	0.094	0.342	0.177	0.066	0.095	0.109	0.133	0.116	0.119
0.57	1.56	0.425	1.303	0.815	0.278	0.453	0.46	0.615	0.555	0.534
0.15	0.39	0.115	0.181	0.182	0.061	0.142	0.146	0.193	0.176	0.156
0.054	0.13	0.019	0.05	0.056	0.019	0.026	0.057	0.081	0.075	0.054
0.19	0.49	0.106	0.192	0.116	0.056	0.193	0.192	0.248	0.259	0.178
0.038	0.10	0.017	0.031	0.018	0.01	0.029	0.042	0.055	0.058	0.038
0.31	0.72	0.088	0.265	0.114	0.063	0.193	0.31	0.42	0.445	0.282
0.075	0.18	0.018	0.067	0.024	0.012	0.042	0.074	0.101	0.105	0.067
0.23	0.51	0.056	0.23	0.084	0.0315	0.13	0.227	0.306	0.334	0.206
0.042	0.09	0.0092	0.041	0.015	0.005	0.023	0.038	0.053	0.058	0.038
0.31	0.61	0.075	0.289	0.112	0.033	0.171	0.272	0.39	0.427	0.286
0.057	0.10	0.015	0.05	0.02	0.006	0.031	0.044	0.069	0.078	0.047
0.12	0.19	0.011	0.076	0.072	0.023	0.095	0.11	0.145	0.138	0.1
0.008	0.03	0.006	0.027	0.087	0.006	0.006	0.004	0.005	0.006	0.006
0.774	4.45	4.321	1.732	1.685	1.366	0.332	0.414	0.57	0.445	0.782
0.026	0.021	0.035	0.033	0.017	0.062	0.023	0.025	0.026	0.021	0.042
0.018	0.028	0.308	3.835	0.009	0.091	0.053	0.015	0.023	0.05	0.415

quartz rods can be as much as ~ 4 vol.% in the same grain in Fig. 8c.

The recognition of amphibole + quartz exsolutions in Cpx is significant: (1) the parental Cpx is supersilicic

that is only stable at very high pressures; and (2) the presence of a significant amount of hydroxyl in the parental Cpx host, perhaps in the form of structural effects could be an important source of retrograde fluid



Fig. 9. Diagrams of major elements vs. MgO.

for amphibolite-facies overprinting during exhumation (Zheng et al., 2003). The latter suggests that the protoliths of the garnet peridotite and garnet pyroxenite may have formed or experienced an environment in which a fluid phase was present in its history before exsolution. The maximum 8.5 vol.% of exsolved amphibole lamellae suggests the presence of >0.2 wt.% (>2000 ppm by weight) hydroxyl in their parental Cpx crystals. Similar results have been reported in the ultrahigh-pressure (>6 GPa) Cpx of eclogites from the

Kokchetav UHP terrane, Kazakhstan (Katayama and Nakajima, 2002) and of garnet peridotites from the Sulu terrane, eastern China (Chen and Xu, 2005).

4.5. Spinel

Cr-rich spinel occurs as fine-grained $(5-100 \ \mu m)$ euhedral crystals scattered fairly uniformly both inbetween and as inclusions of major silicate minerals in garnet lherzolite and garnet-bearing dunite, or occurs as exsolved lamellae in porphyroblastic olivine (Song et al., 2004) and Cpx (Fig. 8b and d). We thus infer that these spinel crystals are secondary and are not in equilibrium with garnet+pyroxene+olivine of peak metamorphic conditions. $Cr^{\#}$ (Cr/[Cr+Al]) of the spinel varies from 0.54 to 0.69, and the Mg[#] varies from 0.84 to 0.25.

Cr-rich spinel in garnet-free dunite (2C15) is compositionally in equilibrium with the coexisting

olivine and orthopyroxene (Fig. 3e), and has relatively higher but narrow range of $Cr^{\#}$ (0.66–0.73) and Mg[#] (0.49–0.51).

4.6. Phlogopite

Phlogopite mainly occurs in garnet pyroxenite and coexists with garnet and the two pyroxenes (Fig. 3f). However, phlogopite is not a primary phase of the fresh



Fig. 10. Diagrams of Ni, V, La, Sm, Yb, Tb/Yb vs. MgO and Yb, Tb/Yb vs. Al₂O₃.

Table 6 Sm-Nd isotopic analyses for garnet peridotite and pyroxenite

Sample	Rock type	Sm (ppm)	Nd (ppm)	147Sm/144Nd	¹⁴³ Nd/ ¹⁴⁴ Nd (2σ)	$\varepsilon_{\rm Nd}(t)~(t=460~{\rm Ma})$	$T_{\rm DM}$ (Ma)
2C12	Grt pyroxenite	0.80	5.28	0.0920	0.512179 ± 13	-2.8	1216
2C13	Grt pyroxenite	0.57	3.23	0.1061	0.51216 ± 10	-4.0	1400
2C15	Grt-free dunite	0.06	0.22	0.1716	0.512536 ± 89	-0.5	2215
2C39	Grt-bearing dunite	0.17	1.25	0.0833	0.511946 ± 7	-6.8	1405
2C42	Grt lherzolite	0.19	0.92	0.1259	0.512332 ± 12	-1.8	1418
2C44	Grt lherzolite	0.15	0.57	0.1621	0.512485 ± 25	-1.0	1960
2C50	Grt lherzolite	0.32	1.26	0.1538	$0.512371 \!\pm\! 12$	-2.7	1977

garnet peridotite. Very minor phlogopite is also observed in some strongly serpentinized garnet peridotite. Mg[#] of the phlogopite varies from 0.92 to 0.95 in the serpentinized garnet peridotites and from 0.87 to 0.89 in the garnet pyroxenites. TiO₂ concentrations in all phlogopite crystals are <1 wt.%, and Al₂O₃ ranges from 17.1 to 18.2 wt.%.

4.7. Amphibole

Amphiboles are the retrograde phase that overprints the primary mineral assemblage Grt + Opx + Cpx (+Ol) in garnet peridotites and garnet pyroxenite. They are pargasitic in composition with low TiO₂ (<0.8 wt.%), high Al₂O₃ (11.8–16.7 wt.%) and MgO (16.8–18.7 wt.%).

5. Whole-rock geochemistry

Whole-rock major elements (Table 5) were analyzed using a RIX-2100 X-ray fluorescence (XRF) spectrometer on fused glass disks at Northwest University of China. Calibrated against Chinese rock reference standard GBW07105, the reproducibility of major elements in garnet lherzolite and garnet pyroxenite is better than 2% (Si, Mg, Fe, Al, Ca), and that of minor and trace elements (Ti, Na, Cr, Co, Ni, V) better than 5%. The reproducibility of Ca and Al is better than 20% for dunite samples.

Whole-rock trace elements (Table 5) were analyzed using an Elan 6100-DRC inductively coupled plasma mass spectrometry (ICP-MS) at Northwest University and Guangzhou Institute of Geochemistry of China. 50 mg of



Fig. 11. Chondrite-normalized REE patterns and primitive mantle-normalized spidergrams for garnet peridotite from the North Qaidam UHP belt.

sample was dissolved in equal mixture of subboiling distilled superpure HF and HNO₃ with a Teflon digesting vessel on a hot-plate for 24 h. This procedure was repeated using smaller amounts of acids for a further 12 h. After digestion, the sample was evaporated to incipient dryness, refluxed with 6 N HNO₃, and heated again to incipient dryness. The sample was then dissolved in 2 ml of 3 N HNO₃ and diluted with Milli-Q water (18 M Ω) to a final dilution factor of 2000. Two USGS rock reference materials BCR-1 and BHVO-1 were used to monitor the analytical accuracy and precision. Analytical accuracy, as indicated by relative difference (RE) between measured and recommended values is better than 10% for most elements, ranging between 10 and 13% for Cu, Sc, Nb, Er, Th, and U, and between 11 and 20% for Ta, Tm, and Gd.

Sm–Nd isotopic analysis on representative samples (Table 6) was done at the Radiogenic Isotopes Laboratory in the Institute of Geology and Geophysics of Chinese Academy of Sciences in Beijing. The samples were dissolved in HClO₄ and HF in Teflon vessels at ~ 100 °C for 3 days and further dissolved at about 200 °C. Isotopic analyses were conducted using a Finnigan MAT-262 mass spectrometer. Blanks for the whole chemical procedures are 10^{-11} g for Sm and Nd. Mass fractionation was corrected against ¹⁴⁶Nd/¹⁴⁴Nd=0.7219. All ¹⁴³Nd/¹⁴⁴Nd values are given relative to 0.512643 ± 8 (2 σ) for USGS Rock Reference standard BCR-1. The constants are GBW04419 143 Nd/ 144 Nd=0.512728±9 (2 σ), λ^{147} Sm=6.54×10⁻¹² a⁻¹. $\varepsilon_{Nd}(t)$, the initial ¹⁴³Nd/¹⁴⁴Nd ratios, were calculated on the basis of present-day reference values for chondritic uniform reservoir (CHUR): $(^{143}\text{Nd}/^{144}\text{Nd})_{\text{CHUR}} = 0.512638, (^{147}\text{Sm}/^{144}\text{Nd})_{\text{CHUR}} =$ 0.1967.

5.1. Major element variations

Fig. 9 shows MgO variation diagrams of major element oxides of the four rock types. Except for the garnet pyroxenite, peridotite and dunite samples show coherent variations in major element oxides such as SiO₂, TiO₂, Al₂O₃, FeO, CaO and Na₂O. Deviations of a few individual samples from the main trend are largely due to mineral modal variations. Garnet-free dunite (e.g., samples 2C15 and 2C45) has higher $Mg^{\#}$ (0.924– 0.925) than garnet-bearing dunite (0.907-0.918), garnet lherzolite (0.824–0.903) and garnet pyroxenite samples (0.809-0.811). As expected, Al₂O₃, CaO and TiO₂ concentrations in dunite samples are extremely low. Garnet-bearing dunite and garnet lherzolite samples show a wide range of MgO (anhydrous) from 46.0 to 24.7 wt.%, Al₂O₃ 6.3 to 1.3 wt.%, and CaO 8.9 to 0.2. As expected from the mineral compositions, Al₂O₃



Fig. 12. IUGS rock classification (Streckeisen, 1976) of samples from the Luliangshan metamorphic garnet peridotite using the protolith mineralogy (Olivine, Opx, Cpx plus minor spinel) calculated from the bulk-rock major element compositions using the procedure developed for ultramafic rocks by Niu (1997) (see his Fig. 5 and Appendix C). The recovery for the totals is better than 99% for the garnet lherzolite and dunite (true ultramafic with ~25–48 wt MgO), but poor (94.3 to 96.9%) for garnet pyroxenites (not true ultramafic: MgO ~18 wt.%). The numerals next to the symbols are bulk-rock Mg[#]=[Mg/(Mg + Fe²⁺)], assuming total Fe as Fe²⁺ in their protolith stages).

contents tend to reflect the variation of modal percentage of garnet, and CaO contents are largely controlled by modal percentage of clinopyroxene in garnet lherzolites. In Ca/Al vs. MgO diagram (Fig. 9g), garnet-free dunite has low Ca/Al radios, whereas garnet pyroxenite, garnet lherzolite and garnet-bearing dunite show rather constant but subchondritic Ca/Al (~0.80–1.15), with the exception of one garnet lherzolite (2C50) and two garnet-bearing dunite samples (2C30 and 2C36).

Garnet pyroxenites have a narrow range of MgO contents from 18.0 to 17.5 wt.%, which are much lower than garnet lherzolite and dunite. As shown in Fig. 9, SiO₂, Al₂O₃, TiO₂ and K₂O (vs. MgO) do not follow the trend defined by garnet lherzolite and dunite. K₂O contents of garnet pyroxenite vary from 0.51 to 0.93, much higher than that of garnet lherzolite and dunite. K₂O in garnet lherzolite and dunite shows poor correlation with MgO contents (Fig. 9h) largely due to analytical uncertainties; some strongly serpentinized samples (2C42, 2C27, 2C37) contain higher K₂O, reflecting that phlogopite is probably not the original phase in most rocks.

5.2. Trace element variations

Fig. 10 shows MgO variation diagrams of some compatible and moderately incompatible trace elements. Well-defined positive Ni–MgO and negative V–MgO trends are obvious (Fig. 10a and b). On REE vs. MgO



Fig. 13. A tectonic modal that illustrates the forming process and environment for Luliangshan garnet peridotite. (a) Subduction of North Qilian Ocean and subarc cumulate chamber at about 460 Ma. (b) Corner flow of the mantle wedge during continental subduction has dragged the cumulate down to depths about 150–200 km at about 420 Ma and exhumed during subsequent continental collision (\sim 400 Ma) (Song et al., 2005b).

and Al_2O_3 plots (Fig. 10c-f), the better inverse correlations of MgO and positive correlations of Al_2O_3 with progressively heavier REEs (i.e., from La to Sm, to Yb) among the studied samples are consistent with MgO and heavy REEs reflecting a primary igneous process, whereas the light REEs may reflect the effects of metamorphism/metasomatism (Niu, 2004).

Chondrite-normalized rare earth element (REE) and primitive mantle-normalized (Sun and McDonough, 1989) trace element abundances are shown in Fig. 11. The garnet pyroxenite samples have uniform wholerock REE compositions and show identically U-shaped REE patterns with LREE-enrichment [(La/Yb)_N=2.45-3.10] and without obvious Eu anomalies (Table 5, Fig. 10a). The garnet lherzolite exhibits two groups of REE patterns. Samples 2C27, 2C42, 2C46 and 2C50 are LREE-enriched with $(La/Yb)_N > 1$ (1.15–2.61) similar to those of the garnet pyroxenites. Other garnet lherzolite samples are slightly depleted in LREE with $(La/Yb)_N < 1$ (0.77–0.96). All dunite samples, but 2C15 (too depleted in REEs), also show LREE-enriched and MREE-depleted patterns with (La/Yb)_N varying from 1.9 to 5.0. The slightly U-shaped REE patterns of these

rocks suggest their previous depletion overprinted by LREE enrichments.

The primitive mantle-normalized whole-rock trace element patterns of these rocks (Fig. 11b and d) show overall large ion lithophile element (LILE) enrichments, in particular the spikes of the fluid-mobile elements Cs, U, and Pb. Most garnet pyroxenite and garnet lherzolite samples show relative depletion in high field strength element (HFSE, i.e., Nb, Ta, Zr, Hf, Ti). The garnetbearing dunite samples show similar HFSE depletion, suggesting that these rocks are genetically related in some form. However, the garnet-free dunite samples (2C15 and 2C45) show different HFSE and HREE systematics, and may have different histories (Fig. 11).

The elevated abundances of LREE and LILE (Cs, Rb, Ba, U and Pb) in most samples are better interpreted as resulting from enrichments by a fluid-rich component. The freshness of some rocks with low LOI suggests that more recent alteration events, including serpentinization, may not be the cause of such enrichments. The high-level constancy of HREE patterns of these rocks suggests that they are not affected by the variation of garnet modal contents. It follows that the protoliths of these rocks already possessed such bulk-rock trace element systematics prior to the ultrahigh-pressure metamorphism. We thus incline to conclude that the banded structure of the massif is most likely inherited from their protoliths, probably magmatic layering.

5.3. Sm-Nd isotopes

Sm–Nd isotope data of unleached whole rocks of two garnet pyroxenite (2C12 and 2C13), three garnet lherzolite (2C42, 44 and 50), one garnet-bearing dunite (harzburgite 2C39) and one garnet-free dunite (2C15) samples are shown in Table 6. All ¹⁴³Nd/¹⁴⁴Nd initial ratios are corrected to 460 Ma based on magmatic ages (Song et al., 2005b).

All samples of the garnet peridotite massif yield negative ε_{Nd} values ranging from -0.5 to -6.8, which differ from sub-continental lithospheric mantle (e.g., Downes, 2001) and the depleted oceanic lithospheric mantle (e.g., Hart, 1988), but resemble terrigenous sediments and garnet peridotite from the Dabie-Sulu UHP terrane of eastern China (Jahn, 1999; Yang and Jahn, 2000). The garnet-free dunite has the highest ε_{Nd} value -0.5 and may represent fragments of subarc lithosphere weakly metasomatized by the ascending mantle wedge melt (see below), whereas the garnetbearing dunite (2C39) has the lowest ε_{Nd} value (-6.8) that may represent fragments of highly metasomatized subarc lithosphere. The garnet lherzolite samples have relatively higher ϵ_{Nd} value of -1.0 to -2.7 than garnet pyroxenites (-2.8 to -4.0), indicating that the ε_{Nd} values are closely associated with the degree of LREE enrichment in these rocks with the exception of the garnet-free dunite 2C15.

6. Discussion

The origin of garnet peridotites from orogenic belts has long been discussed (e.g., Ernst, 1978; Evans and Trommsdorff, 1978; Carswell et al., 1983; Medaris and Carswell, 1990; Zhang et al., 1994, 1995; Brueckner, 1998; Liou and Zhang, 1998; Medaris, 1999; Brueckner and Medaris, 2000; Zhang et al., 2000). Garnet peridotite of the Western Gneiss Region, Norway, for example, is interpreted to be a typical high-P-T residual peridotite of ancient mantle melt extraction events (Brueckner and Medaris, 1998, 2000), and is genetically unrelated to Caledonian UHP metamorphism (Medaris, 1999). Zhang et al. (2000) subdivided garnet peridotites from the Dabie–Sulu UHP terrane into two distinct types: type A, mantle fragments originated from the mantle wedge, footwall mantle, or mantle fragment emplaced at crustal depths prior to UHP metamorphism; type B, metamorphosed mafic–ultramafic complexes that were intruded into the continental crust as magmas prior to subduction. Petrological and geochemical data suggest that garnet peridotite massif from the North Qaidam UHP belt is not a simple mantle fragment, but an arc–subarc lithological association genetically associated with subduction zone magmatic processes before dragged down to deep mantle by the subducting oceanic/continental lithosphere. We interpret that this lithologic assembly was exhumed when the continental portion of the subducting oceanic lithosphere resisted subduction in response to continental collision.

6.1. Protoliths of the four rock types

Whole-rock and mineral compositions show that the garnet peridotite massif from the North Qaidam UHP belt resembles "Mg-Cr type" of Carswell et al. (1983). Field occurrences of these various rock types show layered structures largely defined by modal variations of major constituent minerals (garnet, olivine, orthopyroxene and clinopyroxene). Rhythmic crystallization bands of the protoliths can be convincingly inferred in some outcrops (e.g., Fig. 2b). In sample 2C28 (Fig. 2e), olivine Fo contents of olivine-rich layers are 90.6-91.2 (90.9 ± 0.2) , obviously higher than those of olivine-poor layers (Fo 89.5-90.1 with a mean of 89.8 ± 0.3). This observation allows us to infer a process of magmatic crystallization-differentiation for their protoliths before UHP metamorphism. The interlaying relationship between garnet-bearing dunite (2C26, 27, 36, 37 and 39) and garnet lherzolite (2C28, Fig. 2e) is also clear in the field. All these favor their being magmatic cumulate, consistent with trace element systematics (see Fig. 11).

Using bulk-rock major element compositions on an anhydrous basis, the standard CIPW normative minerals can be readily converted to norms of ideal peridotite minerals (Niu, 1997) as shown in Fig. 12. Clearly, the protoliths of the garnet lherzolite, garnet-bearing and garnet-free dunite encompasses a wide range of rocks equivalent to low-pressure (spinel peridotite stability depths or shallower) magmatic cumulate assemblage of websterite, olivine websterite, lherzolite to harzburgite. Such large variation could be due to large bulk-rock compositional variation as a result of sample size variation, large grain size and modal heterogeneity, but the overall increase in bulk-rock Mg[#] with increasing modal olivine is consistent with their protoliths being of cumulate origin.

Cumulus minerals must be in equilibrium with the melt from which they crystallized. Assuming that the

garnet peridotite massif was indeed a cumulate assemblage precipitated in an MgO-rich magma chamber at upper mantle level (crust-mantle boundary), then their crystallization pressures must be at depths of spinel/plagioclase peridotite stability fields ($\sim 1-$ 2 GPa). Their crystallization temperatures can thus be precisely estimated from experimentally established basalt phase equilibria (e.g., Roeder and Emslie, 1970; Bender et al., 1978; Walker et al., 1979; Langmuir and Hanson, 1981; Nielson and Dungan, 1983; Weaver and Langmuir, 1990; Niu, 1997; Niu et al., 2002), which states that the liquidus temperature of basaltic melt is proportional to the MgO content in the melt. The relationship between the two parameters is characterized by the well-known Fe-Mg exchange relationship of Roeder and Emslie (1970):

$$\mathrm{Kd} = (X_{\mathrm{Mg}}^{\mathrm{L}}/X_{\mathrm{Fe2+}}^{\mathrm{L}})/(X_{\mathrm{Mg}}^{\mathrm{ol}}/X_{\mathrm{Fe2+}}^{\mathrm{ol}}) = 0.3 \pm 0.03.$$

However, since the olivines in all rock types are recrystallized at ultrahigh-pressure conditions during continental subduction, the Fo contents cannot represent their primitive composition. In this case, *mg*-numbers of whole-rock compositions are used. Therefore, the garnet peridotite rocks were crystallized from relatively primitive melt of $Mg^{\#} = ~0.75$ at temperature of ca. 1250 °C (for garnet-bearing dunite sample of $Mg^{\#} = ~0.56$ at temperature of ca. 1170 °C (for garnet lherzolite sample of $Mg^{\#} = 0.824$).

Given the fact that the garnet porphyroblasts are clearly of metamorphic origin and that the olivine has varying Fo (Mg[#]=0.81–0.92), we can infer that the protolith of the lherzolite is neither fertile asthenospheric mantle material, nor mantle melting residues, but is most likely ultramafic cumulate from a cooling basaltic melt (Mg[#]=0.558–0.748) (Niu et al., 2002). Indeed, the reconstructed low-pressure (e.g., stable in the spinel peridotite field) (Fig. 12; Niu, 1997) modes of these samples from their bulk-rock compositions plot in the fields of lherzolite and olivine websterite (Fig. 12). The garnet-bearing dunite with high olivine modes and Mg[#] (0.912-0.918) could be interpreted as "fertile" source lherzolite to residual harzburgite, but their field relations with other samples suggest their being cumulate from the more primitive melt (Mg[#]=0.736 and 0.752 respectively) (Niu, 1997). Therefore, the protoliths of garnet lherzolite and garnet-bearing dunite are the most primitive cumulate and must have crystallized beneath the cold thermal boundary layer atop the mantle.

Reconstructed low-pressure modes (Fig. 12; Niu, 1997) from bulk-rock compositions suggest that the

protolith of the garnet-free dunite samples is variably/ highly depleted harzburgitic melting residues in equilibrium with mantle melts of $Mg^{\#}=0.75-0.77$ at relatively high pressures (>1 GPa).

The bulk-rock MgO values of 18.07 ± 0.22 wt.% suggest that the protoliths of the garnet pyroxenite dike/ dikelets are not of melt compositions because the latter would imply unrealistically hot melt (i.e., ~ 1420 °C). This inference plus the very low TiO₂ of ~ 0.05 wt.% favors that the protoliths of the garnet pyroxenites are also of cumulate origin and precipitated within the conduits of melt migration. These would then suggest some rather evolved parental melts (i.e., $Mg^{\#}=0.54$) with liquidus temperature of ~1162 °C (Niu et al., 2002). The relatively high Al_2O_3 (14.84±0.99 wt.%), CaO $(10.29\pm0.82 \text{ wt.}\%)$ and K₂O $(0.68\pm0.24 \text{ wt.}\%)$ yet rather low TiO₂ (0.05 wt.%) of the bulk-rock compositions suggest that the parental melts resemble subduction zone related mantle wedge melts. If the estimated liquidus temperature of ~1162 °C is reasonable, then the emplacement depth of the pyroxenite dikelets must be shallow, probably no deeper than spinel stability field. This would further suggest that these dikelets must also have experienced subsequent subduction zone metamorphism.

6.2. Tectonic settings of the protoliths

The foregoing discussion allows us to infer that the protoliths of the bulk garnet peridotite massif are of cumulate origin from mantle-derived basaltic melts. The major (Fig. 9) and trace (Figs. 10 and 11) element characteristics of these rocks favor that the mantle melts are subduction zone associated mantle wedge melts. The emplacement depth of the cumulate is hard to evaluate, but probably in the subarc lithosphere where the mantle wedge-derived melts experience cooling and crystallization. Some circumstantial lines of evidence in support our inference include: (1) all rocks show relative depletion of HFSEs (Nb, Ta, Zr, Hf and Ti) and enrichment of LILEs with spikes of fluid-mobile elements (e.g., Cs, U, Pb and Sr), which are typical geochemical signatures of subduction magmatism (e.g., Tatsumi et al., 1986; McCulloch and Gamble, 1991; Davidson, 1996; Elliott et al., 1997; Ewart et al., 1998; Zanetti et al., 1999). (2) The high-MgO parental melts estimated from Mg[#] of garnet-bearing dunite suggest that they may be sourced from more depleted mantle, probably the depleted subarc asthenosphere. (3) Magmatic zircon cores have been discovered in the garnet lherzolite samples and yielded 457±22 Ma (Song et al., 2005b), which represents subduction zone associated

mantle wedge magma genesis, and is consistent with metamorphic ages of the \sim 458 Ma (zircon SHRIMP and Sm/Nd bulk-rock mineral isochron ages) eclogites of MORB/OIB protoliths from the same UHPM belt (Song et al., 2003a,b, 2006). (4) All the three major rock types (except for garnet-free dunite) show weak to none negative Eu* anomaly, suggesting that plagioclase is originally absent during crystallization of their protoliths. Assuming that the peridotite massif was indeed a cumulate massif precipitated from a melt, the estimated liquidus temperature would be 1170-1250 °C with a mean of $\sim 1214 \pm 32$ °C (Niu et al., 2002). Such relatively low liquidus temperatures (low Mg[#] of the minerals and the bulk rock compositions) argue for basaltic (vs. ultramafic) parental melts. Given that the liquidus phases are dominated by Opx, Cpx and olivine without plagioclase (too low Al₂O₃ and CaO in the bulkrock compositions), we can infer that the crystallization pressures are significantly higher than 0.8 GPa (Langmuir et al., 1992) or the melts are rich in water that suppresses plagioclase but enhances Cpx on the liquidus (Gaetani et al., 1993). Such conditions are less likely beneath normal ocean ridges, but readily satisfied in a mantle wedge overlaying a subduction zone.

Most subduction-related ultrahigh-pressure peridotites are interpreted to have been derived either from the lithospheric mantle wedge above former subduction zones or have been exhumed in higher levels of the system while convergence continued (e.g., Brueckner and Medaris, 2000). These rocks often show a metasomatic enrichment of slab-derived elements (e.g., Zanetti et al., 1999; Paquin and Altherr, 2002). The enrichment of LILEs and LREEs and unradiogenic Nd (i.e., low ε_{Nd}) of the Luliangshan garnet peridotite is consistent with such enrichments. The terrigenous sediments may also contribute to the geochemical signatures of the peridotite.

6.3. Tectonic evolution of the garnet peridotite massif

The field observations and petrologic and geochemical data discussed above suggest that the Luliangshan garnet peridotite massif is of cumulate origin crystallized from Mg-rich melts generated in a mantle wedge overlaying a paleo-subduction zone. We further propose a three-stage tectonic evolution model for the garnet peridotite massif in the North Qaidam UHP belt as illustrated in Fig. 13. Stage I (~460 Ma): dehydration of subducted oceanic lithosphere (eclogite-facies metamorphic stage, Song et al., 2006) caused mantle wedge partial melting and high-Mg melt generation in a subarc environment; cooling-induced crystallization of these melts led to the formation of the cumulate assemblage in the subarc lithospheric mantle in the spinel peridotite stability field at about 460 Ma. Stage II: subducting slab-induced mantle wedge asthenospheric corner flow transported the cumulate peridotite body deep into the mantle in the subduction zone. Stage III: the subducted continental crust seized the peridotite body and carried it to depths in excess of 200 km at about 423 Ma before exhumed with the supracrustal felsic gneisses to the middle-crust level at about 400 Ma.

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