

Discovery of sapphirine-bearing Mg–Al granulites in the North China Craton: Implications for Paleoproterozoic ultrahigh temperature metamorphism

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Abstract

We report here for the first time the occurrence of Mg–Al granulites within the khondalite belt of the North China Craton and provide robust evidence for extreme crustal metamorphism at ultrahigh-temperature (UHT) conditions in this region. The UHT indicator assemblages in these rocks include equilibrium sapphirine+quartz, high alumina orthopyroxene+sillimanite+quartz and high temperature perthites. The surrounding khondalites also carry spinel+quartz equilibrium assemblage. The sapphirines occurring in various microstructural associations are markedly magnesian ($X_{\text{Mg}}=0.69\text{--}0.77$) with a composition varying from 7:9:3 to 2:2:1 end members. Spinel is a solid solution of Mg-spinel and hercynite ($X_{\text{Mg}}=0.41\text{--}0.54$) with ZnO content varying from 0.48 to 1.79 wt.%. Garnet shows a compositional range between pyrope and almandine ($X_{\text{Mg}}=0.39\text{--}0.48$) with minor grossular (2.6–3.8 mol%) and spessartine (0.5–1.8 mol%). The orthopyroxenes display marked compositional variations in terms of Al_2O_3 content, with the core of the medium-grained orthopyroxene adjacent to sillimanite in the matrix showing the highest Al_2O_3 content (8.4 wt.%). The history of prograde to peak metamorphism of these rocks is indicated by inclusions of sapphirine, spinel, quartz and sillimanite within garnet. Sapphirine formed during the peak stage through the reaction: $\text{Spl} + \text{Qtz} + \text{Sil} \rightarrow \text{Spr}$. Occurrence of sapphirine and quartz inclusions trapped in sillimanite suggests the reaction: $\text{Spr} + \text{Qtz} + \text{Grt} \rightarrow \text{Opx} + \text{Sil}$. A series of retrograde reactions are observed in these rocks as corona and symplectite textures around early coarse-grained minerals such as: $\text{Opx} + \text{Sil} + \text{Qtz} \rightarrow \text{Crd}$, $\text{Grt} + \text{Sil} + \text{Qtz} \rightarrow \text{Crd}$, $\text{Grt} + \text{Qtz} \rightarrow \text{Crd} + \text{Opx}$, $\text{Grt} + \text{O}_2 \rightarrow \text{Crd} + \text{Qtz} + \text{Fe}_2\text{O}_3$ in Mag or Ilm, and finally $\text{Opx} + \text{Kfs} + \text{H}_2\text{O} \rightarrow \text{Bt} + \text{Qtz}$. Whereas the stability of sapphirine+quartz and orthopyroxene+sillimanite+quartz in the matrix assemblage suggests $T > 1000^\circ\text{C}$ and $P > 10$ kbar peak metamorphism, conventional geothermobarometric estimates confirm the ultrahigh-temperature nature and also trace the retrograde conditions ($930\text{--}970^\circ\text{C}$ at $P > 10$ kbar — Al in Opx; $900\text{--}1000^\circ\text{C}$ — perthite; $930\text{--}990^\circ\text{C}$ — Grt–Opx; and $910\text{--}940^\circ\text{C}$ — sapphirine–spinel).

Electron microprobe dating of monazites in textural association with the UHT assemblages yield a sharply defined isochron age of 1917 ± 48 Ma, which correlates with the 1927 ± 11 Ma obtained in a previous study from monazites in the surrounding khondalites carrying spinel+quartz assemblage. An evaluation of the assemblages and textures using appropriate petrogenetic grids indicate that following peak ultrahigh temperature conditions, the rocks underwent initial isobaric cooling and subsequent isothermal decompression, and we interpret these trajectories to be part of an overall anti-clockwise P – T evolution. The study area, located within the northern margin of the North China Craton, defines a collisional orogenic belt where continental collision and deep subduction was possibly followed by extensional collapse of the orogen. Evidence for the extensional phase is provided by the several mafic magmatic intrusives of Paleoproterozoic age occurring in the region which suggest a possible link between the thermal anomaly associated with extreme crustal metamorphism in this region with mantle-derived magmas. The UHT metamorphism in North China Craton coincided with the assembly of the Paleoproterozoic supercontinent Columbia.

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1. Introduction

Metamorphism at extreme pressure–temperature conditions has emerged as one of the thrust areas of research in petrology during the recent years (Harley, 2004 and references therein). Characteristic mineral assemblages developed under ultrahigh-temperature (UHT) and ultrahigh-pressure (UHP) metamorphic events provide important clues on deep crust and crust–mantle processes as well as in formulating geodynamic models on collision–subduction–exhumation styles. Characterizing the nature and timing of episodic extreme metamorphism is also crucial in understanding the dynamics of the origin and evolution of continents and supercontinents as well as the thermal perturbations associated with major orogenies (Santosh et al., 2006a,b). Mineral assemblages typical of UHT rocks have been recorded from a number of granulite facies terrains, particularly from within Mg–Al rich pelitic lithologies, expanding the realm of granulite petrology and initiating several interesting approaches in constraining the stability fields, formation conditions, fluid regimes and tectonic implications of these rocks (e.g., Harley, 1998, 2004; Tsunogae et al., 2002; Das et al., 2003; Santosh et al., 2004; Kelsey et al., 2005; Santosh and Sajeew, in press, among others).

Granulite facies rocks that underwent extreme crustal metamorphism at $T=900\text{--}1150\text{ }^{\circ}\text{C}$ and $P=7\text{--}13\text{ kbar}$ are generally designated as UHT rocks (e.g. Harley, 1998, 2004). Some of the typical mineral assemblages that characterize the UHT rocks include sapphirine+quartz (Dallwitz, 1968; Ellis et al., 1980; Grew, 1980), spinel+quartz (Friedman, 1953; Morimoto et al., 2004), orthopyroxene+sillimanite+quartz (Hensen and Green, 1973; Tateishi et al., 2004; Sajeew et al., 2004), osumilite+garnet (Ellis et al., 1980; Grew, 1982), high alumina orthopyroxene (Harley and Motoyoshi, 2000; Sajeew et al., 2004), inverted pigeonite (Sandiford and Powell, 1986; Tsunogae et al., 2002), and high-fluorine biotite and pargasite (Motoyoshi and Hensen, 2001; Tsunogae et al., 2003).

Among the above key indicators of UHT metamorphism, the direct contact between sapphirine and quartz is regarded as the most robust evidence for UHT condition (e.g., Harley, 1998). The available phase equilibrium data suggest that sapphirine+quartz assemblage is stable at $>1030\text{ }^{\circ}\text{C}$ at 9.5 kbar (Hensen and Green, 1973) or $>1050\text{ }^{\circ}\text{C}$ at 11 kbar (Bertrand et al., 1991) at low oxygen fugacity conditions. Ever since Dallwitz (1968) first reported the coexistence of sapphirine and quartz in granulites from the Napier Complex, East Antarctica, there have been several reports of sapphirine+quartz assemblages from various regional UHT metamorphic terrains such as Napier and Lützow–Holm Complexes (East Antarctica), Wilson Lake (Canada), Labwor (Uganda), Eastern Ghats (India), Anapolis–Itaçu Complex (Brazil), In Ouzzal (Algeria), Cortlandt Complex (USA), Chogar Complex (Siberia), Highland Complex (Sri Lanka) and Madurai Block in southern India (Ouzegane and Boumaza, 1996; Yoshimura et al., 2003; Sajeew and Osanai, 2004; Tateishi et al., 2004, Harley, 2004 and references therein). Among the spinel-bearing textural equilibria reported from pelitic rocks throughout the world, the spinel+quartz assemblage is considered as a potential indicator of ultrahigh-

temperature crustal metamorphism. Spinel+quartz equilibrium has been reported from terrains of various ages and tectonic settings such as the Archean crustal fragments in Enderby Land, Antarctica (e.g. Sandiford and Powell, 1986) and Namaqualand, South Africa (e.g. Waters, 1989), the Grenville terrains like Eastern Ghats, India (e.g. Dasgupta et al., 1995), and the Pan-African terrains like southern Madagascar (e.g. Nicollet, 1990), Trivandrum and Madurai Blocks in southern India (Morimoto et al., 2004; Tsunogae and Santosh, in press), Highland Complex of Sri Lanka (Sajeew and Osanai, 2004), Lützow–Holm Complex, Antarctica (Kawakami and Motoyoshi, 2004), Anapolis–Itaçu Complex, Brazil (Moraes and Fuck, 2000) and In Ouzzal, Algeria (Ouzegane and Boumaza, 1996). Recently, spinel and quartz in textural equilibrium has also been documented from the Paleoproterozoic khondalite belt in the North China craton (Santosh et al., 2006).

In this study, we report for the first time a number of characteristic UHT assemblages, including the key indicators like sapphirine+quartz and orthopyroxene+sillimanite+quartz from Mg–Al granulites that we discovered within the granulite facies metapelite belt in the North China Craton. Our study provides clear evidence for extreme metamorphism at elevated temperatures in this terrain. From age data on monazites in textural association with the UHT assemblage, we demonstrate that the extreme crustal metamorphism in North China Craton coincided with Paleoproterozoic continental collision associated with the assembly of the Columbia supercontinent.

2. Geologic and tectonic framework of North China Craton

The North China Craton (Fig. 1) is considered to have formed through the suturing of an Eastern and a Western Block along the Central Orogenic Belt (Zhao et al., 2002; Kusky and Li, 2003). However, there is considerable debate surrounding the timing of this amalgamation. Some workers propose that the collision occurred at 2.5 Ga (Kusky and Li, 2003), whereas others believe that the collision occurred at ca. 1.85 Ga (Zhao et al., 2001). There is also considerable debate surrounding the question of 1.9–1.8 Ga collisional orogen along the northern margin of the North China Craton. Kusky and Li (2003) proposed the existence of this belt, but Zhao et al. (2006) came out with the idea that the granulites along the northern margin of the North China Craton were generated during multiple geological events and that the granulites along the northern margin of the Western and Eastern Blocks formed at 2.5 Ga with anticlockwise $P\text{--}T$ paths, whereas those along the northern margin of the Trans-North China Orogen formed during the collision between the Eastern and Western Blocks at 1.85 Ga with clockwise $P\text{--}T$ paths. Zhao et al. (2006) believe that North China Craton evolved independently and at different times through the amalgamation of three distinct granulite blocks.

Recently Santosh et al. (2006) reported monazite electron probe ages from granulite facies metapelites (khondalites) from the northwestern part of the North China Craton from rocks that carry spinel+quartz UHT association. They obtained two age peaks at $1927\pm 11\text{ Ma}$ and $1819\pm 11\text{ Ma}$, which were correlated

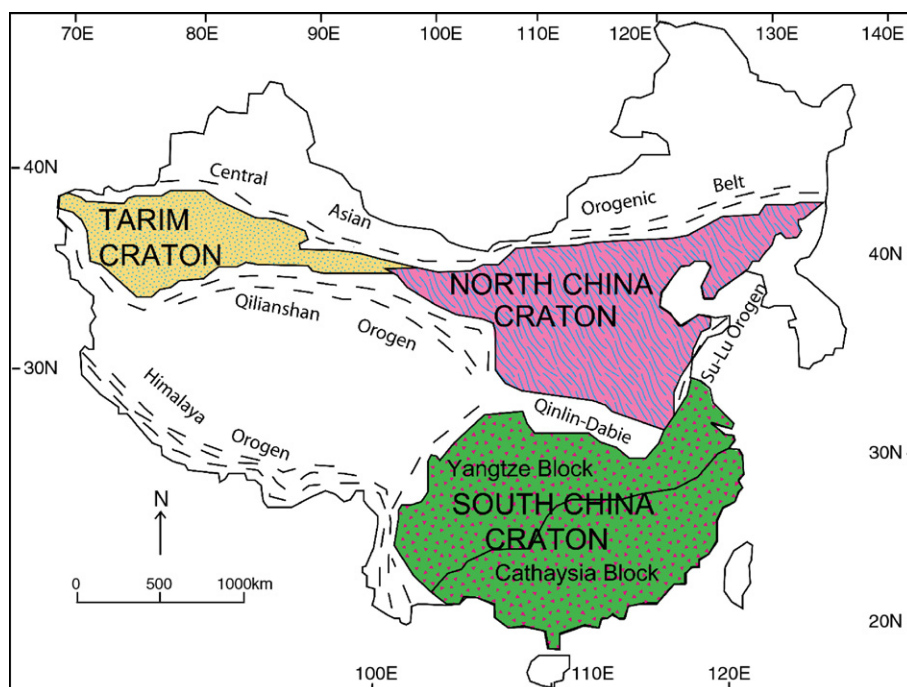


Fig. 1. Tectonic framework of China showing the major Precambrian cratons and younger orogens (after Zhao et al., 2001).

with the timing of a major collisional event and associated extreme crustal metamorphism in this belt. These results favour the ca. 1.9–1.8 Ga E–W collisional orogen in the northern margin of the North China Craton as proposed by Kusky and Li (2003).

3. The khondalite belt

The khondalite belt, comprising a vast sequence of metasediments, is distributed in a large area along the northwestern

segment of the North China Craton (Fig. 2). The rocks occur broadly within three terrains namely, the Jining–Liangcheng–Fengzhen, Daqingshan–Wulashan and Helanshan. The main rock associations in Jining–Liangcheng terrain are khondalites and S-type granites with minor metagabbro. The S-type granite belts are largely controlled by NNE trending sinistral strike-slip shear zones. To the west, the Daqingshan–Wulashan terrain is made up of reworked granulite–charnockite complex, khondalites, and granites. The Daqingshan–Wulashan terrain is dominated by nearly west- to east-trending structural grains.

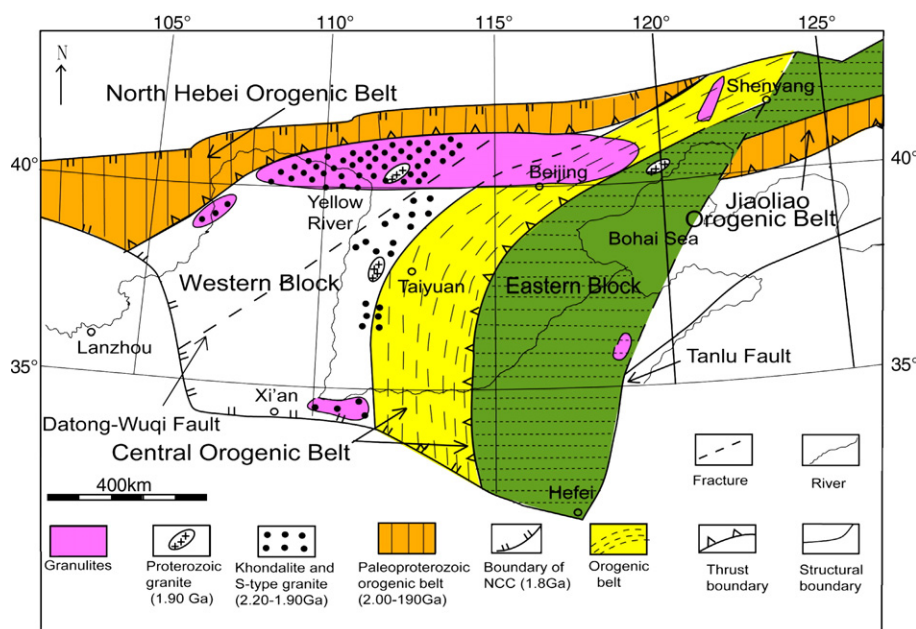


Fig. 2. Geologic and tectonic map of the North China Craton (after Kusky and Li, 2003), showing the distribution of the khondalite belt.

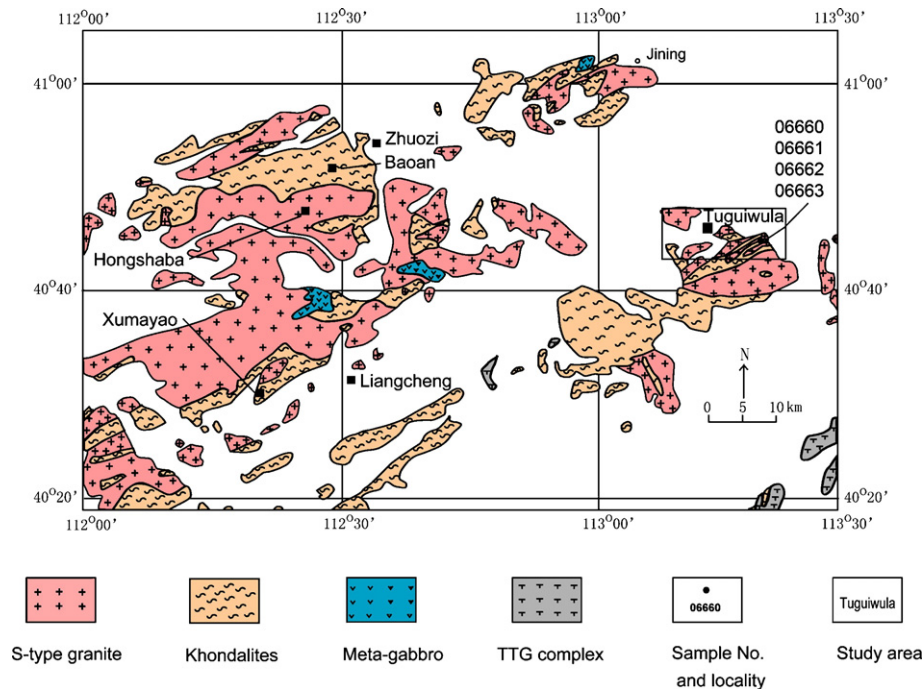


Fig. 3. Geological map of the khondalite belt showing the study area and sample locations. Various localities mentioned in the text are also shown.

Further to the west, the Helanshan terrain also consists of khondalites–S-type granite association. This terrain is characterized by NEE-trending foliation, associated with nearly EW- to WNW-trending structural dome. The common association of the khondalite terrains and synchronous emplacement of S-type granites with shear zone networks have been correlated with Palaeoproterozoic reworking episodes along the northern margin of the North China Craton. A molasse basin, older than 1.90 Ga lies to the northern margin of the belt. To the south, the khondalite belt is substituted by amphibolite-facies metasupracrustals and S-type granites. Mafic dyke intrusions of ca. 1.76–1.80 Ga occur in all the three terrains.

The major rock types in the khondalite belt comprise garnet- and sillimanite-bearing granulite grade metapelites, garnet- and biotite-bearing leptynites, marbles and calc-silicate rocks, broadly identical with the lithologic units in the khondalite belt of southern Kerala in India metamorphosed during the Late Neoproterozoic–Cambrian times (cf. Santosh et al., 2005). Condie et al. (1992) and Li et al. (2000) summarized the geochemical characteristics of the khondalites from North China Craton, and described features typical of post-Archean shales. These rocks are characterized by enrichment of large ion lithophile elements, low Zr–Hf–Sr contents, significant enrichment in LREE and negative Eu anomalies. The khondalite suite of lithologies is thought to be derived from the high-grade metamorphism of interlayered psammitic, psammopelitic, pelitic and carbonaceous sediments deposited in a shallow sea upon cratonic shelf. Graphite deposits commonly occur in many localities within the khondalite belt. The voluminous S-type granites containing garnet in most places may represent extensive partial melting during the high grade metamorphism of the thick sedimentary pile and melt movement. The metasedimentary

sequence overlies a high-pressure granulite belt, the Hengshan belt, which extends over 700 km (cf. Kusky and Li, 2003).

Zircons from the Jining Group khondalites have yielded two distinct age populations: 2.30–2.60 Ga and 1.85–2.00 Ga; the older ages are interpreted as the detrital zircon age and the younger ones as the metamorphic age (cf. Kusky and Li, 2003). A number of granite bodies with ages ranging from 2.2 to 1.9 Ga are emplaced within the Jining khondalite terrain, particularly the voluminous S-type granites which were possibly generated through anatexis of the pelitic sediments. These S-type granites from khondalite terrains of southern Inner Mongolia at the northwestern part of the North China Craton have yielded zircon U–Pb ages of 1971 ± 60 Ma and 1837 ± 40 Ma, Pb–Pb ages of 1975 ± 23 Ma and 1893 ± 14 Ma, and Rb–Sr isochron ages between 1970 and 1840 Ma (Guo et al., 1999). Among the granites, two types have been recognized namely garnet granite and leucogranite. The formation age of the garnet granite from Sanggan area has been estimated as 1836 ± 18 Ma from zircon U–Pb concordia; while the leucogranites gave an upper intercept age of 1912 ± 98 Ma (Guo et al., 1999). Detrital zircons from the garnet–sillimanite–plagioclase gneiss (khondalite) yielded an age of 2130 Ma, while the metamorphic zircons are dated at 1872 Ma (Wu et al., 1998). Monazite dating of khondalites with spinel + quartz UHT assemblage yielded two age peaks at 1927 ± 11 Ma and 1819 ± 11 Ma, correlated to distinct thermal perturbation during the prograde and peak stages of extreme crustal metamorphism in this belt (Santosh et al., 2006).

4. Geology of the study area

Detailed field investigations were carried out in the Bao'an, Hongshaba, Xumayao and Tuguiwula areas within the

khondalite belt (Fig. 3). In this study we focus on rocks collected from an outcrop in the Tuguiwula area. Interlayered aluminous metasediments in association with graphite-bearing garnet–biotite gneiss, garnet-bearing felsic gneiss and S-type granite are the dominant rock types in these areas. The rocks show prominent compositional layering with garnet–sillimanite±cordierite±spinel-rich dark bands and quartzo-feldspathic leucosome layers indicating partial melting. The khondalite layers commonly range in thickness from few meters to several tens of meters, and show polyphased deformational structures. An early NW trending and a later SW trending lineations are identified. The garnet–gneiss belts are strongly folded and the enclosing S-type granites also suffered strong ductile deformation with sinistral shear sense. At least four generations of deformation can be recognized with the khondalites involving D₁ early thrust; D₂ extension; D₃ strike-slip shear deformation (mainly sinistral) and D₄ intrusion of mafic dyke swarms.

The Bao'an area is characterized mainly by S-type granites in which melanocratic residual bands of metapelites occur pro-

fusely. At least four mafic dykes occur within the S-types in this area, ranging in width from 1.5 to 5 m. S-type granite–khondalite association is also exposed in the Honshaba area, where the khondalite band shows strong folding. Garnet bearing granites and khondalites underwent strong ductile shear deformation, and were partly mylonitized in the Xumayao area. The felsic bands show stretching, folding and development of S₂ foliation. Garnet-bearing dark grey coloured mylonitised gneiss appears in Zhaohaogou area, with large bands and relics of aluminous metapelites. Melanocratic residual bands rich in garnet and sillimanite occur within partially melted khondalites at Tuguiwula area. The sapphirine-bearing Mg–Al granulites examined in this study were collected from the Tuguiwula area, where they occur as several centimeter- to decimeter-sized concordant bands and layers within khondalites. Although the sporadic occurrence of sapphirine has been noted from the Inner Mongolia region in an earlier study (Liu et al., 1992), there has been no report as yet on the occurrence of ultrahigh temperature Mg–Al granulites in the North China craton, and this study is the

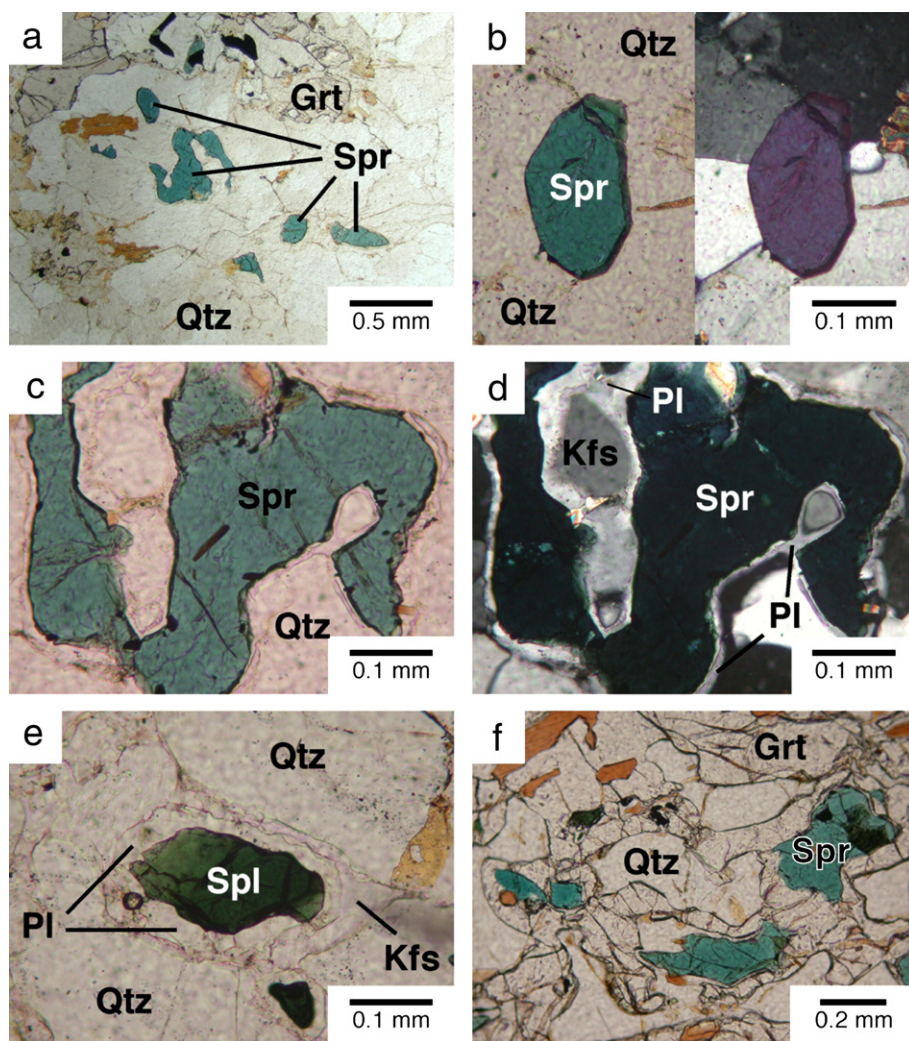


Fig. 4. Photomicrographs of sapphirine-bearing granulites from North China Craton. Mineral abbreviations are after Kretz (1983). (a) Bluish sapphirine scattered in quartzo-feldspathic layer of sample 06660. (b) Left: Equilibrium sapphirine + quartz assemblage in sample 06660. Right: same as the left photograph, crossed polars. (c) Sapphirine and quartz separated by K-feldspar and plagioclase in sample 06660. (d) Same as (c), crossed polars. (e) Spinel surrounded by plagioclase and K-feldspar in quartzo-feldspathic portion of the gneiss (sample 06660). (f) Poikiloblastic garnet with sapphirine, spinel, and biotite in melanocratic portion of sample 06660.

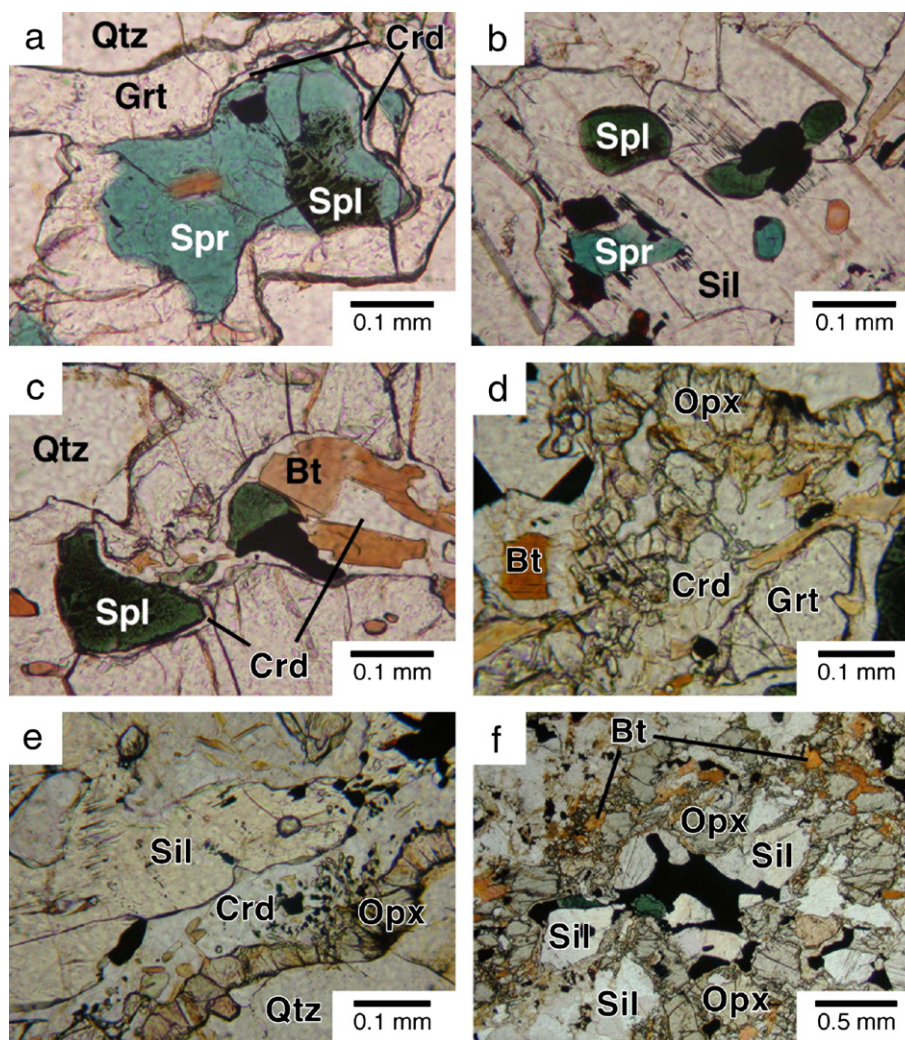


Fig. 5. Photomicrographs of sapphirine-bearing granulites from North China Craton. (a) Sapphirine and spinel included in garnet. Thin cordierite film is occasionally present between sapphirine and garnet (sample 06660). (b) Sillimanite with inclusions of sapphirine, spinel, ilmenite/magnetite, and biotite (sample 06660). (c) Retrograde biotite adjacent to rim of garnet. Biotite is rarely present as inclusion in garnet (sample 06660). (d) Corona of cordierite and vermicular orthopyroxene around garnet rim (sample 06660). (e) Cordierite + orthopyroxene corona between sillimanite and quartz (sample 06660). (f) Orthopyroxene + sillimanite + ilmenite/magnetite assemblage surrounded by retrograde biotite (brownish) in sample 06662.

first attempt to characterize the extreme crustal metamorphism in this region based on the sapphirine-bearing assemblages.

5. Petrography

We give below a brief description of the petrography of the different types of lithologic association in the Mg–Al granulite zone from an outcrop in the Tuguiwula the area. Representative assemblages and microstructures are shown in Figs. 4–7. The details of metamorphic reaction textures will be discussed in a subsequent section.

5.1. Garnet–sillimanite–sapphirine granulite (sample 06660)

The rock exhibits a weak foliation defined by the alternation of quartzo-feldspathic and garnet-rich layers. The quartzo-feldspathic layer is composed of quartz (50–60%), plagioclase (20–30%), K-feldspar (10–20%), biotite (5–10%), opaque minerals (2–5%), sapphirine (<2%), sillimanite (<2%), and

spinel (<1%), whereas the garnet-rich layer is composed of quartz (30–40%), garnet (10–20%), biotite (10–20%), sillimanite (5–10%), plagioclase (2–5%), K-feldspar (2–5%), opaque minerals (1–2%), sapphirine (<2%), spinel (<2%), orthopyroxene (<1%), and cordierite (<1%). Opaque minerals in all the examined samples are mostly ilmenite, but thin exsolution lamellae of magnetite are also present within the ilmenite grains (hereinafter ilmenite/magnetite). No sulphide mineral has been identified in the examined rocks. The sapphirine + quartz assemblage reported in this study was discovered in this rock type, within the quartzo-feldspathic domain. Sapphirine occurs as medium-grained (0.2–0.5 mm) pale bluish grains, and shows a weak pleochroism (Fig. 4a,b,c). The sapphirine grains are mostly anhedral, and are surrounded by medium-grained (0.2–0.4 mm) quartz. The sapphirine and quartz show direct grain contact without any reaction texture developed in between (Figs. 4b and 7a). In the same rock, sapphirine also occurs against quartz isolated by a thin film of plagioclase and K-feldspar (Figs. 4c,d and 7b). Dark greenish

and subhedral spinel is also present in the quartzo-feldspathic layer. Similar to the sapphirine shown in Fig. 4c, the spinel is always surrounded by plagioclase and K-feldspar (Fig. 4e). No direct contact of spinel and quartz has been found in this rock, although in a previous study from khondalites in a nearby locality, Santosh et al. (2006) reported equilibrium spinel+quartz in grain contact occurring both as inclusions within garnet and within matrix assemblage.

Garnet is the dominant mineral within the melanocratic domains of the rock (Fig. 4f). The mineral shows poikiloblastic texture with inclusions of quartz, sillimanite, spinel, sapphirine, ilmenite/magnetite, and biotite (Fig. 5a). Prismatic sillimanite in the layer also contains inclusions of spinel, sapphirine, and magnetite (Fig. 5b). Biotite occurs around both orthopyroxene and garnet as a retrograde phase (Fig. 5c). Rarely, biotite also occurs as minor inclusions in orthopyroxene as an early phase.

Garnet is occasionally surrounded by cordierite with or without fine-grained vermicular orthopyroxene, obviously de-

veloped during the retrograde stage (Fig. 5d). Medium grained discrete grains of orthopyroxene are surrounded by a corona of cordierite (Fig. 5d). Such cordierite–orthopyroxene association is also present around sillimanite (Fig. 5e). A thin film of cordierite isolates garnet and sapphirine (Fig. 5a).

5.2. Orthopyroxene–sillimanite granulite (sample 06662)

This lithology is composed of quartz (30–40%), K-feldspar (30–40%), garnet (5–10%), sillimanite (5–10%), biotite (5–10%), orthopyroxene (2–5%), ilmenite/magnetite (2–5%), spinel (1–2%), cordierite (<1%) and sapphirine (<1%). It is texturally heterogeneous, and consists of orthopyroxene- and sillimanite-rich domains within the matrix of quartz and feldspars. Orthopyroxene is medium-grained and shows intergrowth with sillimanite and ilmenite/magnetite as well as retrograde biotite (Figs. 5f and 6a). A thin cordierite film usually separates orthopyroxene and sillimanite. The cordierite is probably a retrograde product after the high-temperature event (Fig. 6a).

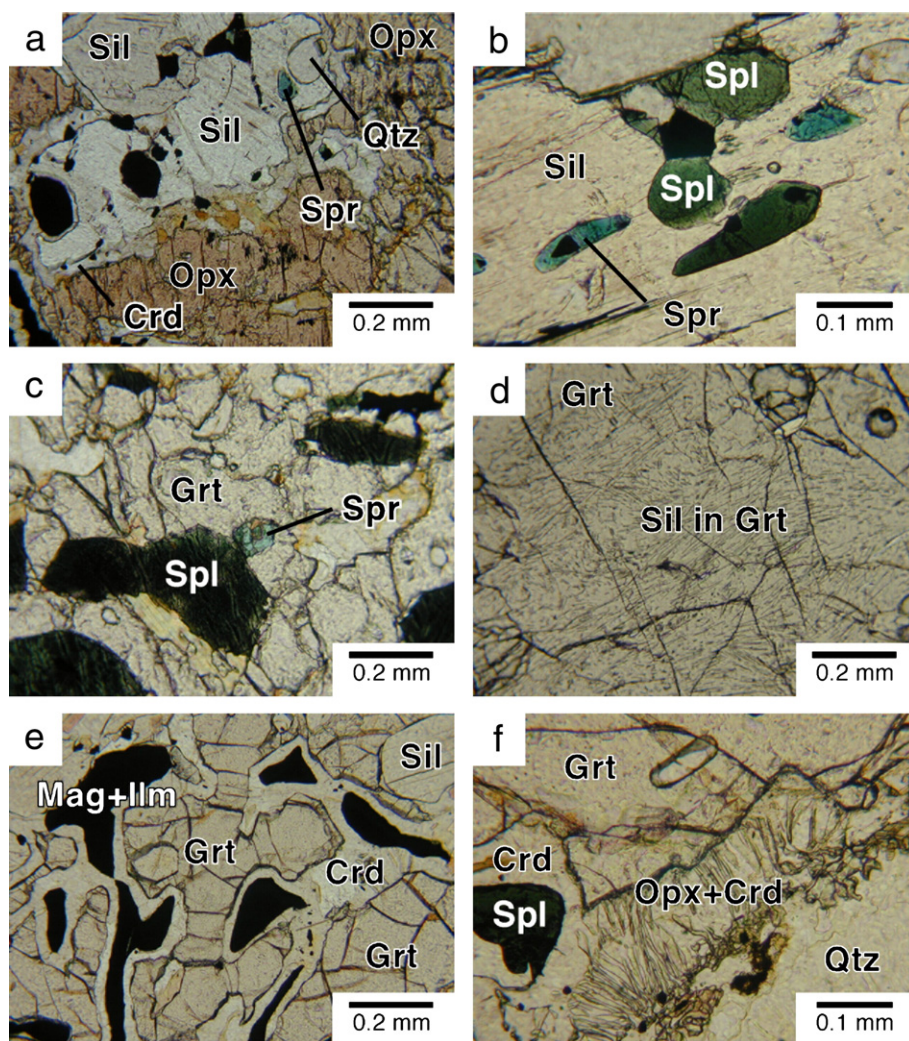


Fig. 6. Photomicrographs of sapphirine-bearing granulites from North China Craton. (a) Coarse-grained orthopyroxene and sillimanite assemblage. The two minerals are separated by a thin film of cordierite (sample 06662). (b) Sapphirine and spinel inclusions in sillimanite adjacent to orthopyroxene (sample 06662). (c) Poikiloblastic garnet containing inclusions of spinel and sapphirine in sample 06661. (d) Foliated fine-grained sillimanite inclusions in garnet in sample 06663. (e) Anhedronal garnet mantled by intergrowth of cordierite and ilmenite/magnetite (sample 06663). (f) Symplectitic corona of orthopyroxene and cordierite between garnet and quartz (sample 06663).

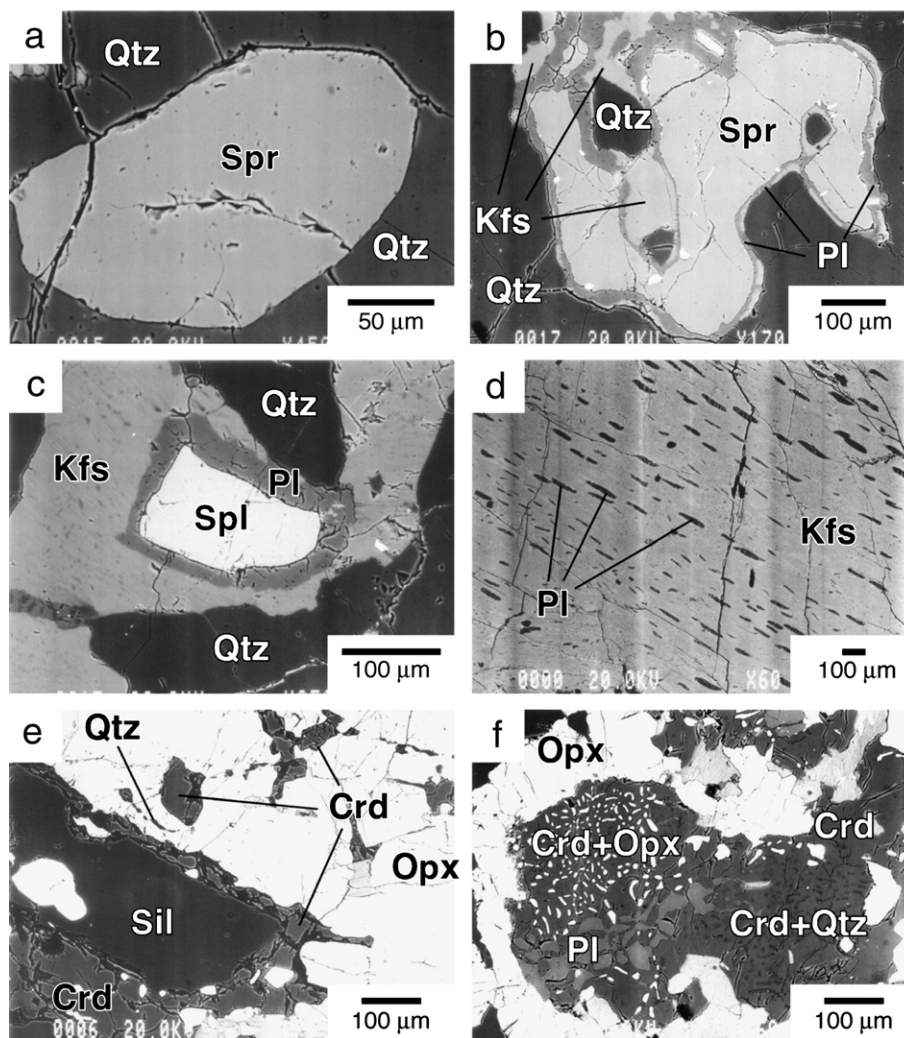


Fig. 7. Back scattered electron images showing detailed textures of sapphire-bearing granulites from North China Craton. (a) Direct grain contact relationship of sapphire and quartz without any reaction products (sample 06660). (b) Thin films of plagioclase and K-feldspar separating sapphire and quartz (sample 06660). (c) Spinel separated from quartz by K-feldspar and plagioclase (sample 06662). (d) Perthite in sample 06661. (e) Orthopyroxene and sillimanite assemblage separated by thin film of cordierite. Quartz occurs as rare inclusions in orthopyroxene (sample 06662). (f) Symplectitic intergrowth of cordierite+orthopyroxene and cordierite+quartz formed during retrograde metamorphism (sample 06663).

Sillimanite associated with orthopyroxene occasionally includes fine-grained sapphire and spinel (Fig. 6b). Rare quartz is present as fine-grained inclusions in orthopyroxene associated with sillimanite.

Garnet is rare in this sample and occurs mainly in the quartzo-feldspathic domain of the rock. No retrograde reaction corona can be seen around the mineral. Spinel occurs as isolated grains in the quartzo-feldspathic portion, or as fine-grained inclusions in sillimanite. Similar to the previous sample, spinel is separated from quartz by plagioclase, K-feldspar, and sillimanite (Fig. 7c).

5.3. Quartzo-feldspathic garnet–spinel granulite (sample 06661)

This lithology is composed of feldspathic layers and garnet-rich layers. The feldspathic layer is composed of K-feldspar (80–90%), quartz (5–10%), plagioclase (5–10%), biotite (1–5%), spinel (<1%), and ilmenite/magnetite (<1%). K-feldspar contains thin lamellae of plagioclase as perthite (Fig. 7d).

Granoblastic texture of the perthite and the presence of spinel along the grain boundaries of perthite suggest that this layer crystallized from a feldspathic melt probably derived by the partial melting of the surrounding garnet- and spinel-bearing pelitic rocks. It is obvious that the feldspathic layer also underwent high-grade metamorphism. We will discuss this aspect further in a later section while dealing with the geothermometry of perthites from this rock.

Garnet-rich layer of the rock is composed of quartz (40–50%), garnet (20–30%), plagioclase (5–10%), sillimanite (5–10%), spinel (2–5%), K-feldspar (2–5%), ilmenite/magnetite (2–5%), biotite (2–5%), and sapphire (<1%). Garnet is coarse-grained (~1.8 mm), poikiloblastic, and contains inclusions of spinel, quartz, sillimanite, ilmenite/magnetite, sapphire, and biotite (Fig. 6c). Quartz, plagioclase, sillimanite, spinel, and ilmenite/magnetite are matrix minerals. Spinel is separated from adjacent quartz grains by K-feldspar and plagioclase. Biotite occurs around the garnet as a retrograde phase.

5.4. Quartzo-feldspathic garnet–sillimanite granulite (sample 06663)

This lithology is composed mainly of K-feldspar (30–40%), quartz (30–40%), garnet (5–10%), sillimanite (5–10%), biotite (5–10%), orthopyroxene (2–5%), ilmenite/magnetite (2–5%), spinel (1–2%), and sapphirine (<1%). Garnet is poikiloblastic and contains numerous fine-grained oriented inclusions of sillimanite (Fig. 6d). In contrast, some garnet grains associated with cordierite contain no sillimanite, but have only quartz. Prismatic sillimanite is also present in the matrix where it shows intergrowth with spinel.

Several retrograde corona textures are present around garnet. In the sillimanite-rich portion of the rock, garnet is mantled by cordierite. Occasionally, irregular garnet grains are separated from ilmenite/magnetite by cordierite (Fig. 6e). In places, orthopyroxene and cordierite symplectite occurs between garnet rim and matrix quartz (Fig. 6f).

6. Mineral chemistry

6.1. Analytical procedure

Chemical analyses of minerals were performed by electron microprobe analyzer (JEOL JXA8621) at the Chemical

Analysis Division of the Research Facility Center for Science and Technology, University of Tsukuba, for all minerals. The analyses were performed under conditions of 20 kV accelerating voltage and 10 nA sample current, and the data were regressed using an oxide-ZAF correction program supplied by JEOL. Representative compositions of minerals in the analyzed samples are given in Tables 1–7. Fe^{3+} of sapphirine, spinel, garnet, and orthopyroxene was calculated based on stoichiometry.

6.2. Sapphirine

Sapphirine in the North China granulites occurs in three settings: (1) as subhedral crystals in grain contact with quartz within quartzo-feldspathic layer (e.g., Fig. 4b; sample 06660), (2) similar to type 1, but sapphirine and spinel are separated by plagioclase and/or K-feldspar (e.g., Fig. 4c; sample 06660), and (3) as fine-grained inclusions in sillimanite and garnet (e.g., Fig. 4f; sample 06660). All the above types of sapphirines are markedly magnesian ($X_{\text{Mg}}=0.70\text{--}0.80$) and their (Mg, Fe)O: $\text{Al}_2\text{O}_3\text{:SiO}_2$ ratio varies between 7:9:3 and 2:2:1 end member compositions depending on mineral assemblages (Fig. 8). The calculated $\text{Fe}^{3+}/(\text{Fe}^{2+} + \text{Fe}^{3+})$ ratio is low in type 1 sapphirine (0.12–0.17), although the ratio is slightly high in sample 06661 (type 3; 0.33–0.37). Other components such as TiO_2 , Cr_2O_3 ,

Table 1
Representative electron microprobe analyses of sapphirine in sapphirine-bearing granulites from North China Craton

Sample no.	06660								06661		06662		06663	
Remarks	Core, with Qtz	Rim, with Qtz	In Sil	In Sil	In Grt	In Grt	Core, no Qtz	Rim, no Qtz	In Grt	In Grt	In Sil	In Sil	With Spl	In Sil
O*	20	20	20	20	20	20	20	20	20	20	20	20	20	20
SiO ₂	14.96	14.97	14.61	14.31	14.70	14.07	15.08	14.76	12.62	12.74	13.64	12.98	13.19	12.87
Al ₂ O ₃	55.77	56.48	56.74	56.92	57.18	56.80	55.54	55.26	59.86	60.46	58.75	60.30	58.47	60.93
TiO ₂	0.07	0.05	0.07	0.05	0.06	0.04	0.06	0.03	0.03	0.03	0.02	0.02	0.02	0.22
Cr ₂ O ₃	0.01	0.09	0.17	0.13	0.20	0.15	0.04	0.02	0.05	0.05	0.12	0.03	0.15	0.07
Fe ₂ O ₃	2.28	2.43	2.40	2.28	1.88	2.78	1.81	3.03	4.58	4.24	2.30	3.15	3.75	2.61
FeO	10.69	10.40	10.21	9.95	10.05	9.00	11.20	10.45	7.03	7.59	9.12	8.47	9.12	8.51
MnO	0.08	0.00	0.04	0.02	0.02	0.00	0.08	0.04	0.02	0.06	0.05	0.04	0.03	0.06
MgO	14.89	15.20	14.96	14.79	15.23	15.22	14.63	14.77	15.51	15.34	15.02	15.06	14.68	15.10
ZnO	0.06	0.10	0.00	0.01	0.05	0.01	0.05	0.01	0.00	0.00	0.00	0.01	0.05	0.00
CaO	0.03	0.01	0.01	0.02	0.02	0.01	0.01	0.01	0.01	0.02	0.00	0.00	0.03	0.01
Na ₂ O	0.00	0.00	0.03	0.04	0.00	0.01	0.01	0.03	0.04	0.03	0.00	0.00	0.00	0.00
K ₂ O	0.00	0.00	0.00	0.02	0.00	0.02	0.00	0.01	0.00	0.02	0.00	0.01	0.00	0.00
Total	98.85	99.74	99.24	98.54	99.38	98.10	98.52	98.42	99.75	100.57	99.02	100.06	99.49	100.38
Si	1.842	1.825	1.788	1.761	1.792	1.735	1.864	1.828	1.526	1.529	1.665	1.566	1.611	1.546
Al	8.090	8.110	8.183	8.256	8.214	8.256	8.090	8.063	8.531	8.557	8.445	8.576	8.415	8.626
Ti	0.006	0.005	0.006	0.004	0.005	0.004	0.006	0.003	0.003	0.003	0.002	0.002	0.002	0.020
Cr	0.001	0.009	0.016	0.013	0.020	0.015	0.004	0.002	0.005	0.005	0.012	0.003	0.014	0.007
Fe ³⁺	0.211	0.223	0.221	0.211	0.172	0.258	0.168	0.282	0.417	0.383	0.211	0.286	0.345	0.236
Fe ²⁺	1.101	1.060	1.045	1.024	1.024	0.928	1.157	1.082	0.711	0.762	0.930	0.855	0.931	0.855
Mn	0.009	0.000	0.004	0.002	0.002	0.000	0.009	0.004	0.002	0.006	0.005	0.004	0.003	0.006
Mg	2.730	2.759	2.728	2.712	2.765	2.796	2.693	2.725	2.794	2.744	2.729	2.708	2.671	2.701
Zn	0.005	0.009	0.000	0.001	0.004	0.001	0.005	0.001	0.000	0.000	0.000	0.000	0.005	0.000
Ca	0.004	0.001	0.001	0.003	0.002	0.001	0.002	0.001	0.002	0.002	0.000	0.000	0.003	0.002
Na	0.000	0.000	0.007	0.009	0.000	0.003	0.002	0.008	0.010	0.007	0.001	0.000	0.000	0.000
K	0.000	0.000	0.000	0.003	0.000	0.003	0.000	0.001	0.000	0.002	0.000	0.001	0.000	0.001
Total	14.000	14.000	14.000	14.000	14.000	14.000	14.000	14.000	14.000	14.000	14.000	14.000	14.000	14.000
Mg/(Fe+Mg)	0.71	0.72	0.72	0.73	0.73	0.75	0.70	0.72	0.80	0.78	0.75	0.76	0.74	0.76
Fe ³⁺ /(Fe ²⁺ +Fe ³⁺)	0.16	0.17	0.17	0.17	0.14	0.22	0.13	0.21	0.37	0.33	0.18	0.25	0.27	0.22

*Number of oxygens.

Table 2
Representative electron microprobe analyses of spinel in sapphirine-bearing granulites from North China Craton

Sample no.	06660			06661		06662			06663	
Remarks	With Opq	In Sil	In Pl	In Grt	In Sil	In Sil	Core	Rim	With Spr	In Grt
O*	4	4	4	4	4	4	4	4	4	4
SiO ₂	0.03	0.24	0.03	0.00	0.05	0.00	0.01	0.05	0.03	0.00
Al ₂ O ₃	62.89	59.60	59.95	61.98	61.27	60.26	60.58	61.72	60.64	59.38
TiO ₂	0.05	0.00	0.00	0.03	0.00	0.02	0.00	0.00	0.00	0.02
Cr ₂ O ₃	0.52	0.75	0.46	0.37	0.17	0.09	0.16	0.28	0.58	0.83
Fe ₂ O ₃	0.98	2.49	2.94	2.57	3.18	3.59	2.93	2.37	2.51	3.77
FeO	21.40	18.98	21.54	20.50	22.24	23.37	24.56	24.25	25.00	25.24
MnO	0.04	0.03	0.00	0.08	0.00	0.07	0.08	0.11	0.18	0.05
MgO	12.49	11.13	9.78	13.23	12.17	10.98	10.20	10.60	10.28	9.99
ZnO	1.24	5.02	4.42	0.68	0.59	0.96	1.21	1.17	0.73	0.84
CaO	0.00	0.00	0.01	0.01	0.01	0.00	0.00	0.00	0.00	0.00
Na ₂ O	0.05	0.13	0.16	0.05	0.06	0.05	0.04	0.07	0.01	0.02
K ₂ O	0.02	0.00	0.02	0.00	0.00	0.00	0.00	0.03	0.00	0.00
Total	99.72	98.37	99.30	99.49	99.73	99.40	99.78	100.66	99.95	100.15
Si	0.001	0.007	0.001	0.000	0.001	0.000	0.000	0.001	0.001	0.000
Al	1.969	1.926	1.937	1.942	1.933	1.927	1.938	1.948	1.935	1.905
Ti	0.001	0.000	0.000	0.001	0.000	0.000	0.000	0.000	0.000	0.000
Cr	0.011	0.016	0.010	0.008	0.004	0.002	0.003	0.006	0.012	0.018
Fe ³⁺	0.020	0.051	0.061	0.051	0.064	0.073	0.060	0.048	0.051	0.077
Fe ²⁺	0.475	0.435	0.494	0.456	0.498	0.530	0.558	0.543	0.566	0.575
Mn	0.001	0.001	0.000	0.002	0.000	0.002	0.002	0.003	0.004	0.001
Mg	0.495	0.455	0.399	0.524	0.485	0.444	0.413	0.423	0.415	0.405
Zn	0.024	0.102	0.090	0.013	0.012	0.019	0.024	0.023	0.015	0.017
Ca	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000
Na	0.003	0.007	0.008	0.003	0.003	0.003	0.002	0.004	0.000	0.001
K	0.001	0.000	0.001	0.000	0.000	0.000	0.000	0.001	0.000	0.000
Total	3.000	3.000	3.000	3.000	3.000	3.000	3.000	3.000	3.000	3.000
Mg/(Fe+Mg)	0.51	0.51	0.45	0.53	0.49	0.46	0.43	0.44	0.42	0.41
Fe ³⁺ /(Fe ²⁺ +Fe ³⁺)	0.04	0.11	0.11	0.10	0.11	0.12	0.10	0.08	0.08	0.12

*Number of oxygens.

and ZnO total less than 0.35 wt.%. Below we illustrate in more detail the compositional variations of sapphirine in the Mg–Al granulites of the North China Block, depending on the different modes of occurrence in a single sample (06660).

Sapphirine in equilibrium with quartz (type 1) is slightly Si-rich and Al-poor (Si=1.83–1.86, Al=8.03–8.15 pfu) than that in the absence of quartz within the melanocratic layers. Type 2 sapphirine in quartzo-feldspathic layer is also Si-rich (Si=1.81–1.83 pfu). X_{Mg} (=Mg/[Fe²⁺+Mg]) of types 1 and 2 sapphirines in quartzo-feldspathic layer is generally low (0.70–0.73). In contrast, type 3 sapphirine enclosed in sillimanite with or without spinel in the melanocratic layer is Al-rich with Al=8.18–8.26 pfu, although the X_{Mg} value of this association is almost consistent with that of types 1 and 2. Sapphirine enclosed in garnet shows the highest Al content and X_{Mg} in the sample as 8.21–8.26 pfu and 0.73–0.75, respectively. Similar sapphirine inclusions in garnet in sample 06661 show the highest X_{Mg} of 0.78–0.80 among all the examined sapphirine grains in the different samples.

6.3. Spinel

Spinel in the North China granulites is a solid solution of Mg-spinel and hercynite (X_{Mg} =0.41–0.54), occasionally with small amounts of Cr₂O₃ (0.09–0.85 wt.%) and MnO (0–0.18 wt.%). Fe³⁺/(Fe²⁺+Fe³⁺) ratio varies from 0.03 to 0.12 for all

spinel, suggesting that Fe³⁺ content is low. ZnO content varies from 0.48 to 5.02 wt.% depending on mineral assemblage. For example, in sample 06660, spinel associated with magnetite/ilmenite shows higher X_{Mg} (0.51–0.52) and lower ZnO content (1.17–1.24 wt.%) than that in sillimanite or plagioclase (X_{Mg} =0.45–0.51, ZnO=4.42–5.02 wt.%) where opaque minerals are not associated.

6.4. Garnet

Garnet in all the examined samples is essentially a solid solution between pyrope and almandine (X_{Mg} =0.39–0.48) with minor grossular (2.6–3.8 mol%) and spessartine (0.5–1.8 mol%) components. All analyzed garnet grains show nearly homogeneous composition, although in general, the rim portions are slightly enriched in almandine. For example, the core of the garnet in sample 06663 has a composition of Alm_{54–55} Pyr_{40–42} Grs_{3–4} Sps_{1–2}, with a slight increase in the almandine component toward rim (Alm_{56–57} Pyr_{39–40} Grs_{3–4} Sps_{1–2}). The garnet rim adjacent to retrograde cordierite corona shows the highest Fe content of Alm_{57–59} Pyr_{37–38} Grs_{3–4} Sps_{1–2}.

6.5. Orthopyroxene

Orthopyroxene shows marked compositional variation, in terms of Al₂O₃ content even within a thin section scale,

depending on the mineral association and microstructure (Fig. 9). Medium-grained orthopyroxene adjacent to sillimanite in the matrix in sample 06662 shows the highest Al_2O_3 content of up to 8.2 wt.% ($\text{Al}=0.36$ pfu). Core of the orthopyroxene is slightly more enriched in Al_2O_3 (8.1–8.4 wt.%; $\text{Al}=0.35$ – 0.36 pfu) than the rim (7.0–7.8 wt.%; $\text{Al}=0.31$ – 0.34). In contrast, orthopyroxene intergrown with cordierite surrounding garnet in sample 06660 is less aluminous ($\text{Al}_2\text{O}_3=5.4$ – 5.9 wt.%, $\text{Al}=0.24$ – 0.26 pfu). Fine-grained orthopyroxene aggregates around cordierite in the same sample show low Al_2O_3 content of 4.6–5.3 wt.% ($\text{Al}=0.20$ – 0.24 pfu). X_{Mg} ratios of all the above orthopyroxene varieties are almost consistent, showing a range of 0.66–0.68, despite the variations in Al_2O_3 contents.

6.6. Feldspars

K-feldspar and plagioclase are dominant minerals in the quartzo-feldspathic layers of the rocks. The dominant feldspar is perthite in most of the examined samples. K-feldspar is rich in orthoclase component, with a broad compositional range of An_{2-3} Ab_{15-22} Or_{77-85} (sample 06661). K-feldspar around sapphirine or spinel is more orthoclase-rich as An_{0-1} Ab_{11-13} Or_{87-89} (e.g., sample 06660). Plagioclase lamella (An_{35-36} Ab_{62-64} Or_{0-2}) in K-feldspar (sample 06661) has a composition

similar to that associated with sapphirine and spinel (An_{34-40} Ab_{59-65} Or_{0-1} ; sample 06660).

6.7. Biotite

Biotite is present in all the examined samples as a retrograde product associated with the rim of ferromagnesian minerals such as orthopyroxene and garnet, although in rare cases they are included in poikiloblastic minerals such as garnet. The retrograde biotite has almost consistent X_{Mg} of 0.78 to 0.88 and TiO_2 content of 3.4–4.7 wt.% throughout all the examined samples. Retrograde biotite around garnet in sample 06661 show the highest Mg content ($X_{\text{Mg}}=0.84$ – 0.88) and the lowest Ti content ($\text{TiO}_2=3.4$ – 3.7 wt.%). Rare biotite inclusion in garnet in the same sample is slightly Mg-poor ($X_{\text{Mg}}=0.83$ – 0.84) and Ti-rich ($\text{TiO}_2=4.5$ – 4.7 wt.%) compared to the retrograde phase. Biotite inclusion in orthopyroxene in sample 06662 is also less magnesian ($X_{\text{Mg}}=0.79$ – 0.81) and Ti-rich ($\text{TiO}_2=4.1$ – 4.3 wt.%).

6.8. Other minerals

Composition of sillimanite is close to the ideal chemistry (Al_2SiO_5), although it contains small amounts of Fe_2O_3 up to

Table 3
Representative electron microprobe analyses of garnet in sapphirine-bearing granulites from North China Craton

Sample no.	06660		06661		06662		06663	
Remarks	Core	Rim	Core	Rim	Core	Rim	Core	Rim
O*	12	12	12	12	12	12	12	12
SiO ₂	40.17	40.45	40.01	40.05	39.74	40.27	39.65	40.10
Al ₂ O ₃	22.95	23.14	22.69	23.05	22.51	22.66	22.32	22.50
TiO ₂	0.00	0.06	0.02	0.02	0.08	0.06	0.06	0.02
Cr ₂ O ₃	0.00	0.04	0.00	0.01	0.01	0.02	0.07	0.00
Fe ₂ O ₃	0.00	0.00	0.25	0.62	0.00	0.00	0.00	0.00
FeO	23.29	24.22	24.25	24.28	24.05	25.22	25.27	26.40
MnO	0.28	0.30	0.37	0.38	0.37	0.36	0.60	0.69
MgO	12.11	12.08	12.18	12.27	11.55	11.35	10.53	10.31
ZnO	0.00	0.00	0.01	0.04	0.04	0.01	0.01	0.00
CaO	1.05	1.18	1.20	0.99	1.09	1.11	1.12	1.10
Na ₂ O	0.02	0.00	0.00	0.02	0.00	0.00	0.01	0.07
K ₂ O	0.00	0.00	0.00	0.00	0.01	0.01	0.00	0.01
Total	99.86	101.46	100.98	101.73	99.43	101.07	99.64	101.20
Si	3.017	3.002	2.992	2.974	3.014	3.016	3.021	3.022
Al	2.031	2.024	2.000	2.017	2.012	2.000	2.005	1.998
Ti	0.000	0.003	0.001	0.001	0.005	0.003	0.004	0.001
Cr	0.000	0.002	0.000	0.001	0.000	0.001	0.004	0.000
Fe ³⁺	0.000	0.000	0.014	0.035	0.000	0.000	0.000	0.000
Fe ²⁺	1.463	1.503	1.516	1.507	1.525	1.580	1.610	1.663
Mn	0.018	0.019	0.023	0.024	0.024	0.023	0.039	0.044
Mg	1.355	1.335	1.357	1.357	1.305	1.266	1.195	1.157
Zn	0.000	0.000	0.001	0.002	0.002	0.001	0.000	0.000
Ca	0.085	0.094	0.096	0.079	0.088	0.089	0.091	0.089
Na	0.002	0.000	0.000	0.003	0.000	0.000	0.002	0.010
K	0.000	0.000	0.000	0.000	0.001	0.001	0.000	0.001
Total	7.969	7.981	8.000	8.000	7.975	7.980	7.972	7.984
Alm	50.1	50.9	50.7	50.8	51.8	53.4	54.9	56.3
Pyr	46.4	45.3	45.4	45.7	44.4	42.8	40.7	39.2
Grs	2.9	3.2	3.2	2.7	3.0	3.0	3.1	3.0
Sps	0.6	0.6	0.8	0.8	0.8	0.8	1.3	1.5

*Number of oxygens.

Table 4

Representative electron microprobe analyses of orthopyroxene in sapphirine-bearing granulites from North China Craton

Sample no.	06660		06662			06663	
Remarks	Symp.	Around Crd	Core	Rim	With Sil	With Grt	With Crd
O*	6	6	6	6	6	6	6
SiO ₂	50.16	51.02	49.01	49.44	50.37	50.45	51.20
Al ₂ O ₃	5.86	4.55	8.18	6.99	6.84	6.90	4.49
TiO ₂	0.10	0.07	0.07	0.12	0.03	0.09	0.06
Cr ₂ O ₃	0.00	0.00	0.00	0.02	0.00	0.04	0.00
Fe ₂ O ₃	1.55	1.53	0.85	1.60	0.77	0.75	0.90
FeO	18.35	18.80	19.23	18.39	19.30	17.28	20.38
MnO	0.12	0.15	0.05	0.13	0.05	0.09	0.27
MgO	23.28	23.48	22.06	22.78	22.90	24.11	22.72
ZnO	0.02	0.04	0.01	0.07	0.00	0.00	0.03
CaO	0.09	0.11	0.06	0.06	0.06	0.08	0.07
Na ₂ O	0.00	0.03	0.00	0.00	0.01	0.00	0.01
K ₂ O	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Total	99.53	99.77	99.52	99.58	100.32	99.78	100.13
Si	1.848	1.879	1.809	1.823	1.842	1.839	1.889
Al	0.254	0.198	0.356	0.303	0.294	0.296	0.195
Ti	0.003	0.002	0.002	0.003	0.001	0.002	0.002
Cr	0.000	0.000	0.000	0.001	0.000	0.001	0.000
Fe ³⁺	0.043	0.042	0.024	0.044	0.021	0.021	0.025
Fe ²⁺	0.565	0.579	0.593	0.567	0.590	0.526	0.628
Mn	0.004	0.005	0.002	0.004	0.001	0.003	0.008
Mg	1.278	1.288	1.213	1.251	1.247	1.309	1.248
Zn	0.001	0.001	0.000	0.002	0.000	0.000	0.001
Ca	0.003	0.004	0.002	0.002	0.002	0.003	0.003
Na	0.000	0.002	0.000	0.000	0.001	0.000	0.001
K	0.000	0.000	0.000	0.000	0.000	0.000	0.000
Total	4.000	4.000	4.000	4.000	4.000	4.000	4.000
Mg/(Fe+Mg)	0.69	0.69	0.67	0.69	0.68	0.71	0.67
Fe ²⁺ /(Fe ²⁺ +Fe ³⁺)	0.07	0.07	0.04	0.07	0.03	0.04	0.04

*Number of oxygens.

1.53 wt.%. Cordierite shows a uniform magnesian composition in all samples with X_{Mg} in the range of 0.84–0.88. The analytical totals of cordierite are less than 100% by 1–3 wt.%, suggesting the presence of channel-filling volatiles such as CO₂ and/or H₂O.

7. Metamorphic reaction textures

We evaluate the mineral assemblages and microstructures described in the previous sections using the (K₂O–) FeO–MgO–Al₂O₃–SiO₂ (–H₂O) [(K)FMAS(H)] system. In order to understand the metamorphic evolution of the rocks, we group the microstructures in the examined samples into three major stages – prograde, peak, and retrograde – based on the host–inclusion relationships and the size and shape of the grains as distributed within the rocks. The compositions and assemblages of minerals in representative samples are shown in SiO₂–(FeO+MgO)–Al₂O₃ diagrams in Fig. 10. The equilibrium mineral assemblages are represented by tie lines in the diagrams.

7.1. Prograde assemblages

Although the prograde mineral assemblages in granulites are often destroyed at high temperature/pressure conditions of

the peak metamorphism, some relicts of the prograde history could be preserved as rare inclusions in coarse-grained poikiloblastic minerals. Garnet is a typical example of a robust mineral that commonly preserves inclusion minerals. A careful examination of the mineral inclusions in many garnet grains from the Mg–Al granulites show that most of these are monophase inclusions of sapphirine, quartz, spinel, and sillimanite. Sapphirine and quartz are occasionally trapped in a single garnet grain, although they do not show mutual grain contact. This is probably because of the low modal abundance of sapphirine included within garnet. As sapphirine+quartz equilibrium has been found in the quartzo-feldspathic layer of sample 06660, we infer that the sapphirine and quartz inclusions within garnet may also have formed during the peak UHT stage.

In all the examined samples, sillimanite is the only aluminosilicate commonly present as inclusion in garnet and also in the matrix. This suggests that the dominant phase of high grade metamorphism of the studied area took place within the stability field of sillimanite.

Table 5

Representative electron microprobe analyses of feldspars in sapphirine-bearing granulites from North China Craton

Sample no.	06660	06661	06662	06662	06660	06661	06662
Mineral name	Pl	Pl	Pl	Pl	Kfs	Kfs	Kfs
Remarks	Around Spl	Lamella	Around Spl	Matrix	Around Spr	Host	Around Pl
O*	8	8	8	8	8	8	8
SiO ₂	59.58	59.83	59.27	60.63	63.71	65.45	65.09
Al ₂ O ₃	25.51	25.66	25.50	24.72	18.52	18.92	18.82
TiO ₂	0.01	0.00	0.00	0.04	0.08	0.05	0.09
Cr ₂ O ₃	0.00	0.02	0.00	0.00	0.00	0.02	0.00
FeO**	0.07	0.07	0.39	0.04	0.04	0.00	0.03
MnO	0.00	0.00	0.00	0.00	0.00	0.03	0.00
MgO	0.00	0.00	0.01	0.00	0.00	0.00	0.00
ZnO	0.00	0.00	0.00	0.00	0.03	0.03	0.00
CaO	7.14	7.44	7.13	6.58	0.08	0.53	0.25
Na ₂ O	7.50	7.36	7.56	7.36	1.29	2.37	2.36
K ₂ O	0.22	0.15	0.13	0.21	14.20	12.64	12.56
Total	100.02	100.52	99.98	99.58	97.96	100.05	99.21
Si	2.657	2.655	2.649	2.704	2.985	2.985	2.990
Al	1.341	1.342	1.343	1.299	1.023	1.017	1.019
Ti	0.000	0.000	0.000	0.001	0.003	0.002	0.003
Cr	0.000	0.001	0.000	0.000	0.000	0.001	0.000
Fe ²⁺	0.002	0.003	0.015	0.002	0.002	0.000	0.001
Mn	0.000	0.000	0.000	0.000	0.000	0.001	0.000
Mg	0.000	0.000	0.000	0.000	0.000	0.000	0.000
Zn	0.000	0.000	0.000	0.000	0.001	0.001	0.000
Ca	0.341	0.353	0.341	0.314	0.004	0.026	0.012
Na	0.648	0.632	0.655	0.636	0.117	0.210	0.210
K	0.013	0.009	0.007	0.012	0.848	0.735	0.736
Total	5.002	4.994	5.011	4.969	4.983	4.977	4.971
An	34.0	35.5	34.0	32.7	0.4	2.7	1.3
Ab	64.7	63.6	65.2	66.1	12.0	21.6	21.9
Or	1.2	0.9	0.7	1.2	87.5	75.7	76.8

*Number of oxygens.

**Total Fe as FeO.

Table 6

Representative electron microprobe analyses of biotite in sapphirine-bearing granulites from North China Craton

Sample no.	06660	06661	06662	06663
Remarks	Around Grt rim	In Grt Matrix	Around Grt rim In Opx	With Crd
O*	22	22	22	22
SiO ₂	39.38	39.20	39.32	39.26
Al ₂ O ₃	13.74	13.99	14.23	14.36
TiO ₂	4.70	4.68	4.65	4.32
Cr ₂ O ₃	0.05	0.06	0.04	0.03
FeO**	8.48	6.83	8.09	6.08
MnO	0.00	0.00	0.00	0.01
MgO	18.83	20.43	18.68	21.60
ZnO	0.00	0.00	0.00	0.04
CaO	0.00	0.01	0.00	0.00
Na ₂ O	0.13	0.16	0.14	0.15
K ₂ O	9.89	9.86	9.96	9.87
Total	95.20	95.21	95.14	94.92
Si	5.709	5.644	5.698	5.654
Al	2.348	2.373	2.427	2.432
Ti	0.512	0.507	0.506	0.373
Cr	0.006	0.006	0.004	0.003
Fe ²⁺	1.028	0.821	0.979	0.731
Mn	0.000	0.000	0.000	0.000
Mg	4.066	4.381	4.027	4.627
Zn	0.000	0.000	0.000	0.007
Ca	0.000	0.001	0.000	0.001
Na	0.037	0.045	0.040	0.042
K	1.829	1.810	1.839	1.809
Total	15.535	15.588	15.520	15.680
Mg/(Fe+Mg)	0.80	0.84	0.80	0.86

*Number of oxygens.

**Total Fe as FeO.

Table 7

Representative electron microprobe analyses of cordierite in sapphirine-bearing granulites from North China Craton

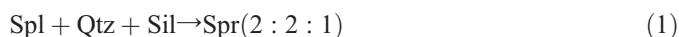
Sample no.	06660	06662	06663
Remarks	Between Opx–Sil	Between Opx–Sil	Around Grt Between Grt–Spl
O*	18	18	18
SiO ₂	49.06	48.76	50.32
Al ₂ O ₃	33.73	33.31	33.66
TiO ₂	0.00	0.00	0.02
Cr ₂ O ₃	0.01	0.02	0.01
FeO**	3.27	3.41	3.77
MnO	0.01	0.00	0.00
MgO	11.84	11.54	12.06
ZnO	0.03	0.01	0.01
CaO	0.01	0.01	0.01
Na ₂ O	0.04	0.03	0.04
K ₂ O	0.03	0.02	0.01
Total	98.02	97.09	99.90
Si	4.956	4.973	4.994
Al	4.014	4.002	3.937
Ti	0.000	0.000	0.002
Cr	0.001	0.002	0.001
Fe ²⁺	0.276	0.290	0.313
Mn	0.001	0.000	0.000
Mg	1.781	1.754	1.783
Zn	0.002	0.000	0.000
Ca	0.001	0.001	0.001
Na	0.008	0.006	0.008
K	0.003	0.002	0.002
Total	11.043	11.029	11.040
Mg/(Fe+Mg)	0.87	0.86	0.85

*Number of oxygens.

**Total Fe as FeO.

7.2. Peak assemblages

The mineral assemblages that equilibrated during the peak stage of metamorphism are inferred from the coarse-grained matrix minerals: quartz, garnet, orthopyroxene, sillimanite, sapphirine, and spinel. Quartzo-feldspathic layer in sample 06660 contains equilibrium sapphirine+quartz assemblage, suggesting that the two minerals formed in the stability field of sapphirine and quartz ($T > 1000$ °C). Close association of spinel and quartz, although the minerals are now separated by thin films of plagioclase and K-feldspar, would suggest that the rocks underwent metamorphism within the spinel+quartz stability field. This inference is further corroborated by the occurrence of equilibrium spinel+quartz in grain contact within khondalites in localities adjacent to that of present study as reported by Santosh et al. (2006). The reaction that led to the formation of the sapphirine+quartz assemblage in the present case cannot be fully resolved. However, the close association of sapphirine+quartz and spinel suggests the following FMAS reaction to explain the formation of sapphirine from spinel, sillimanite, and quartz:



The progress of the reaction is supported by the composition of sapphirine in equilibrium with quartz close to the ideal formula of 2:2:1 (Fig. 8).

Coarse-grained orthopyroxene and sillimanite occur together in the matrix within the melanocratic portion of sample 06662 (Fig. 6a). Small quartz grains also occur together with orthopyroxene and sillimanite, although garnet is not closely associated. This microstructure clearly suggests the stability of orthopyroxene+sillimanite+quartz assemblage. According to the FMAS petrogenetic grid of Harley (1998), the P – T condition

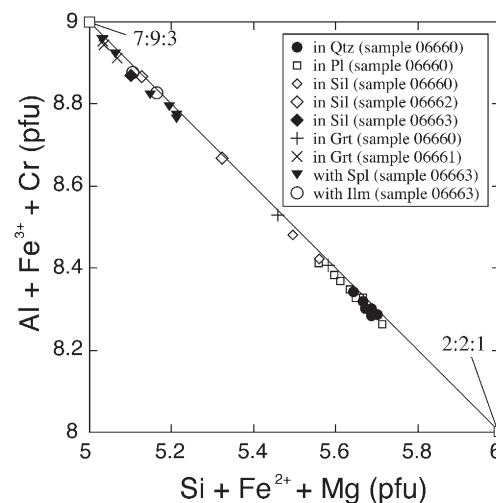


Fig. 8. Compositional diagram showing sapphirine chemistry.

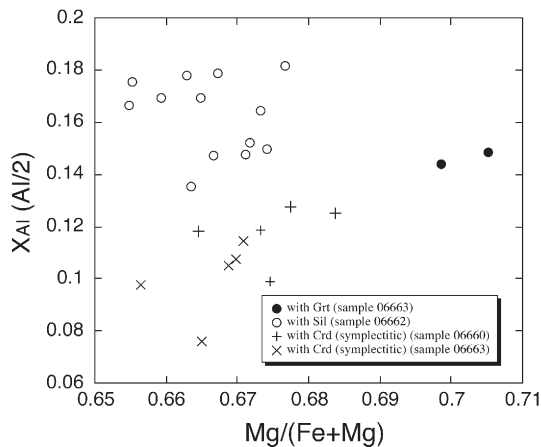


Fig. 9. X_{Mg} versus X_{Al} diagram showing orthopyroxene chemistry.

of this assemblage corresponds to $P > 10$ kbar and $T \sim 1030$ °C. This association is therefore a robust indicator of UHT metamorphic conditions in the North China granulites.

Occurrence of sapphirine and quartz inclusions trapped in sillimanite (Fig. 6a) suggest that the following FMAS univariant reaction took place.



Garnet is not in direct contact with orthopyroxene and sillimanite in this domain as it was probably totally consumed during the progress of the reaction. Such orthopyroxene + sillimanite + quartz assemblage probably formed at the peak UHT stage, but at slightly lower temperature conditions than that inferred from the stability of sapphirine + quartz ($T > 1000$ °C).

7.3. Retrograde assemblages

Retrograde reactions are observed as corona and symplectite textures around early coarse-grained minerals (e.g., garnet). One example is the occurrence of thin film of cordierite between orthopyroxene and sillimanite in the presence of quartz, suggesting the progress of following FMAS continuous reaction;



This reaction implies pressure decrease from the stability field of orthopyroxene + sillimanite + quartz to cordierite during exhumation along the retrograde path.

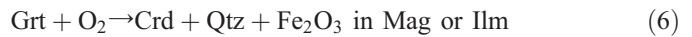
Another common example is the corona of cordierite ± orthopyroxene which mantle garnet grains (Fig. 6f). This microstructure suggests the following FMAS continuous reaction:



The progress of reactions (4) and (5) is supported by the decreasing Mg content in garnet from the core to the rim. Such cordierite ± orthopyroxene corona around garnet has been reported from many other granulite terrains and is often taken

as evidence of near isothermal decompression after peak metamorphism (e.g., Harley, 1989).

Symplectitic fine-grained aggregate of cordierite and quartz occurs around subhedral to anhedral garnet together with cordierite + orthopyroxene symplectite (Fig. 7f). In places, cordierite + quartz aggregate is closely associated with magnetite and ilmenite. Such a texture suggests the reaction,



The reaction texture has been reported from some UHT terrains (e.g., Ishii et al., 2006) as an evidence of retrograde

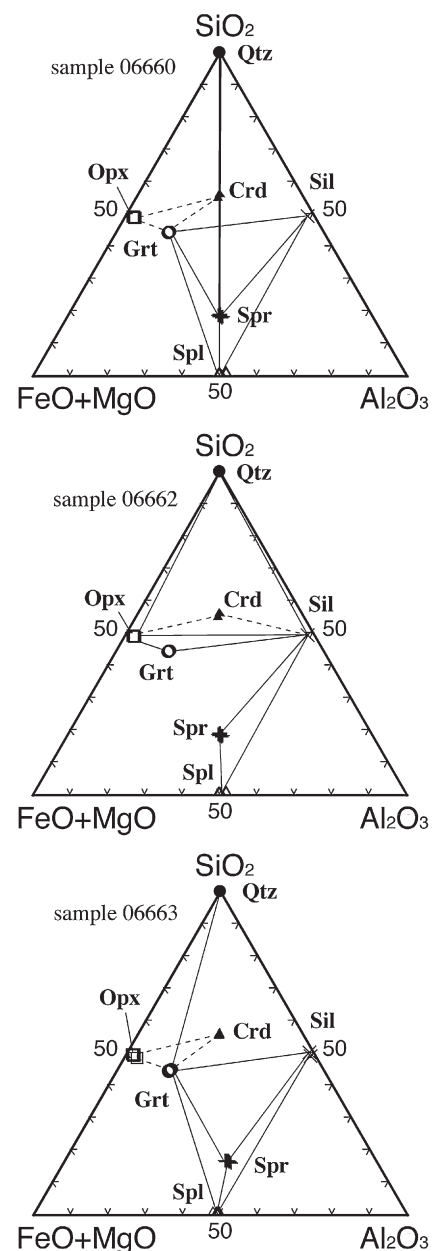


Fig. 10. SiO_2 –($\text{FeO} + \text{MgO}$)– Al_2O_3 diagrams showing compositions and assemblages of minerals in sapphirine-bearing granulites. The equilibrium mineral assemblages are presented by tie lines in the diagrams. Thick lines indicate assemblages in quartzofeldspathic part. Dashed lines show assemblages at retrograde stage.

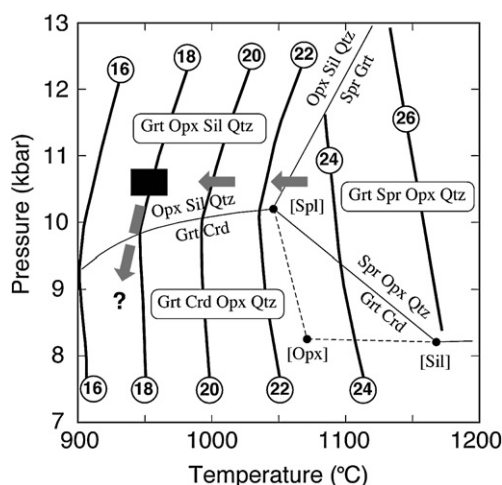


Fig. 11. P – T diagram showing X_{Al} isopleths of orthopyroxene after Harley (2004). The numerical values inside the circles indicate $X_{Al} \times 100$ following Harley (2004). The black square implies a range of X_{Al} values estimated from core of coarse-grained orthopyroxene in sample 06662. Arrows indicate inferred P – T path discussed in the text.

metamorphism, probably related to the activity of oxidized aqueous fluids.

Biotite is closely associated with ferromagnesian minerals such as orthopyroxene and garnet in the quartz-feldspathic matrix (quartz and K-feldspar) suggesting the progress of the following retrograde reaction (7) probably at the final stage of high-grade metamorphism.



8. Metamorphic pressure–temperature conditions

The assemblages and microstructures in the North China UHT rocks allow us to apply several conventional geothermobarometers to infer the peak and retrograde metamorphic conditions. We briefly summarize below the computed pressure–temperature (P – T) conditions from geothermobarometers that are appropriate for the assemblages in the studied samples. We could not apply the GASP geobarometer to these rocks because plagioclase is not in equilibrium with garnet and sillimanite. Garnet-biotite geothermometer was also not applied in this study because most of the biotite appears to have formed towards the final phase of retrograde metamorphism. Later in this section, we will also address the P – T conditions of the study area based on available mineral phase equilibria constraints.

8.1. Al solubility in orthopyroxene

The Al content and X_{Mg} of orthopyroxene coexisting with garnet, sillimanite (or cordierite), and quartz are potential indicators of P – T conditions (Hensen and Harley, 1990). For the estimates, we adopted core compositions of coarse-grained orthopyroxene because these are likely to preserve the near-peak conditions. Isopleths of Harley (2004) based on new experimental data (e.g., Hollis and Harley, 2003) indicate that the core

of orthopyroxene in sample 06662 ($X_{Al} = 0.17$ – 0.18) equilibrated under a temperature range of 930–970 °C at $P > 10$ kbar in the stability field of garnet + orthopyroxene + sillimanite + quartz (Fig. 11). The $T \sim 950$ °C temperature estimate obtained through this method is slightly lower than the stability field of sapphirine + quartz ($T > 1000$ °C, see below) probably due to retrograde re-equilibration of Al in orthopyroxene.

8.2. Ternary feldspar geothermometer

Recent petrological studies on UHT metamorphic rocks indicate that feldspars showing exsolution texture are potential tools for estimating the peak temperature conditions, with examples for temperature estimates around 1000 °C (e.g., Harley, 2004). In this study we applied ternary feldspar geothermometry using perthite in samples 06661 and 06662 to obtain near-peak metamorphic temperatures. The calculation technique involves the determination of early single-phase feldspar and the application of the thermometer following the method described by Hokada (2001). Compositions of perthite grains applied for the temperature estimates are shown in Fig. 12 as a ternary An–Ab–Or diagram with plots of the compositions of host K-feldspar and lamella plagioclase, together with early single-phase feldspar inferred from host–lamella volume proportions estimated from back-scattered images (e.g. Fig. 7d). Using the ternary feldspar geothermometer of Elkins and Grove (1990), we obtained a temperature range of ca. 900–1000 °C for the perthite formation. The results are almost consistent with those obtained from Al-in-orthopyroxene geothermometers. Calculated temperatures using models of Fuhrman and Lindsley (1988) and Lindsley and Nekvasil (1989) are about 100 °C lower than the results of Elkins and Grove (1990). The calculated temperature conditions of $T > 900$ °C further confirm that these rocks were subjected to UHT metamorphism.

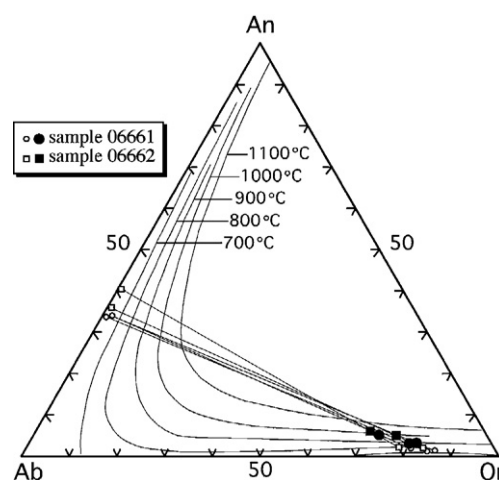


Fig. 12. An–Ab–Or diagram showing compositions of host plagioclase and lamella K-feldspar (open circles and squares) together with those of early single-phase feldspar (solid circles and squares) inferred from host–lamella volume proportions (see text for details). Ternary miscibility gaps are estimated for 8 kbar using the method of Elkins and Grove (1990).

8.3. Garnet–orthopyroxene thermometry

This thermometer was applied to orthopyroxene in equilibrium with sillimanite and adjacent garnet in sample 06662. The estimated temperature range for the garnet–orthopyroxene pairs is 870–920 °C at 9 kbar using the method of Lee and Ganguly (1988), which is based on experimental calibration of Fe–Mg fractionation between garnet and orthopyroxene in FMAS system at 20–45 kbar and 975–1400 °C. The ranges from Aranovich and Berman’s (1997) method (880–970 °C) at 9 kbar, a formulation based on the solubility of Al₂O₃ in ferrosilite (+almandine) at 12–20 kbar and 850–1100 °C, are almost consistent with that of Lee and Ganguly (1988). Application of the method of Pattison et al. (2003), which takes into account the effect of retrograde Fe–Mg exchange, also gave consistent temperature range of 920–960 °C. Calculated temperature ranges from the method of Harley and Green (1982), which is based on experimentally calibrated solubility of alumina in orthopyroxene coexisting with garnet at 5–20 kbar and 800–1200 °C, indicate slightly higher values of 930–990 °C. These temperature ranges obtained in this study (ca. 900–950 °C) are similar to the results from Al-in-orthopyroxene and ternary feldspar geothermometers. In contrast, markedly lower-temperature condition was obtained from garnet rim and adjacent fine-grained orthopyroxene in sample 06663. For example, application of the method of Lee and Ganguly (1988) yielded a temperature range of 670–740 °C, which is about 200 °C lower than the result from garnet–orthopyroxene–sillimanite–quartz assemblage in sam-

Table 8
Electron microprobe analytical data on monazites in sapphirine-bearing granulites from North China Craton

UO ₂ (wt.%)	ThO ₂ (wt.%)	PbO (wt.%)	Count error PbO (%)	Age Ma	Error age (Ma)	ThO ₂ [*] (wt.%)
0.037	7.690	0.646	0.603	1884	12.94	7.829
0.041	4.970	0.440	0.886	1956	19.42	5.125
0.058	8.250	0.705	0.553	1898	12.07	8.467
0.032	3.650	0.318	1.226	1924	25.94	3.770
0.044	3.810	0.335	1.162	1922	24.67	3.978
0.042	3.710	0.309	1.262	1826	25.17	3.867
0.055	7.110	0.601	0.649	1875	13.80	7.315
0.041	7.120	0.607	0.643	1902	13.89	7.276
0.045	7.720	0.647	0.603	1870	12.84	7.889
0.054	6.970	0.590	0.661	1877	14.05	7.175
0.043	7.570	0.640	0.610	1888	13.10	7.732
0.043	3.110	0.270	1.446	1880	29.68	3.273
0.071	6.090	0.530	0.736	1901	15.80	6.359
0.023	10.120	0.855	0.456	1911	10.14	10.206
0.042	3.710	0.307	1.269	1817	25.16	3.867
0.042	3.610	0.303	1.289	1835	25.84	3.766
0.047	4.020	0.347	1.125	1884	23.39	4.196
0.044	3.720	0.306	1.273	1804	25.04	3.883
0.031	5.090	0.446	0.874	1951	19.10	5.206
0.039	5.120	0.436	0.895	1888	18.81	5.266
0.028	5.000	0.430	0.907	1919	19.40	5.106
0.075	4.730	0.430	0.907	1952	19.89	5.016
0.047	3.520	0.295	1.323	1823	26.30	3.697

ple 06662. The temperature range probably indicates post-peak retrograde Fe–Mg exchange between garnet and orthopyroxene during exhumation. Garnet–orthopyroxene–plagioclase–quartz

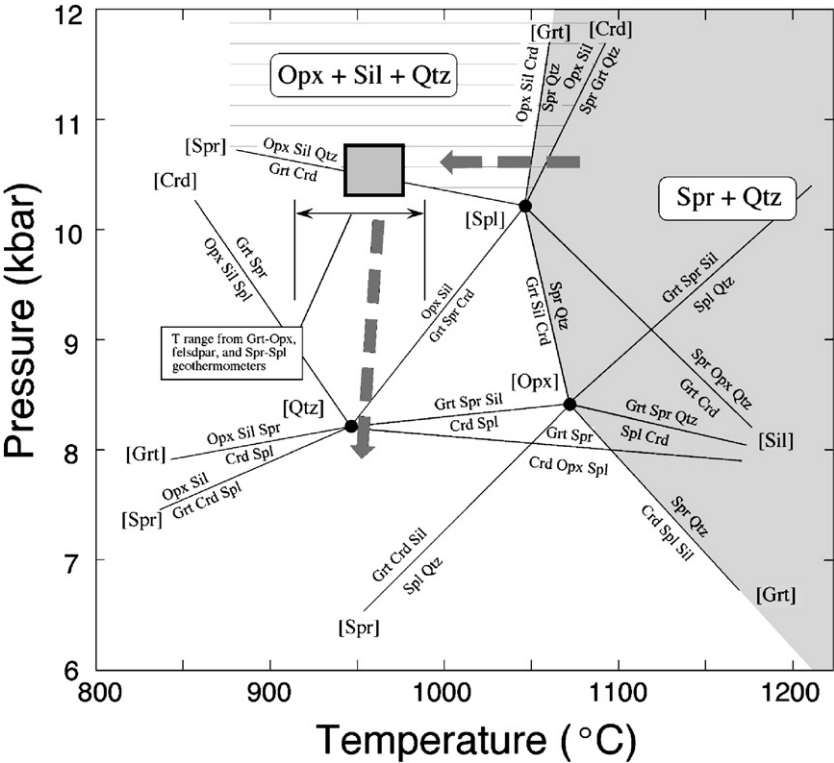


Fig. 13. Petrogenetic grid in the FMAS system (modified after Hensen and Harley, 1990; Harley, 1998). The stability fields of sapphirine + quartz and orthopyroxene + sillimanite + quartz are also shown. Arrows indicate the approximate P–T trajectories inferred from sapphirine-bearing granulites reported in this study. The square indicates the P–T condition calculated from Al-rich orthopyroxene. See text for further discussion.

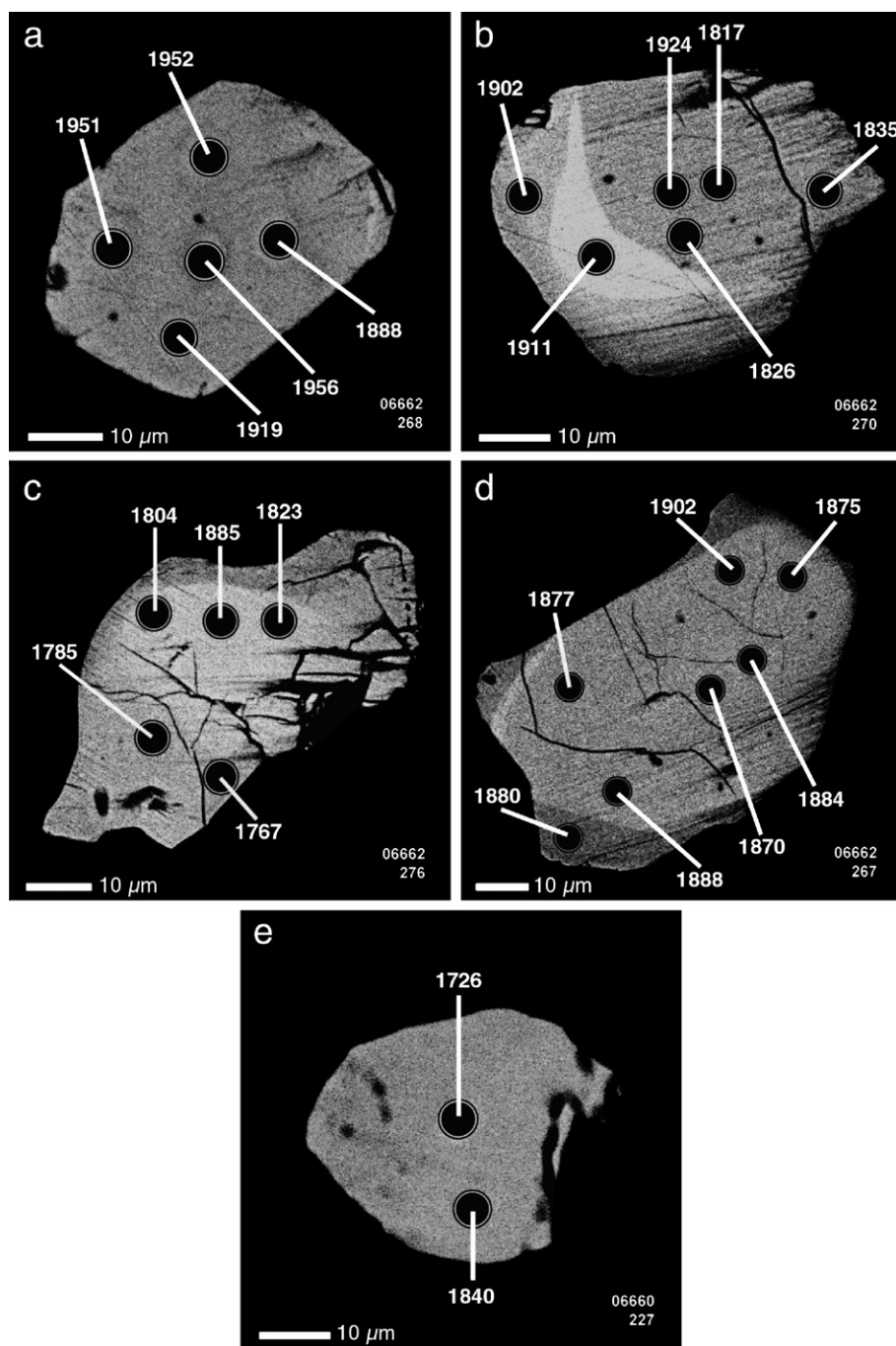


Fig. 14. Backscattered electron images of representative monazite grains in the matrix of the ultrahigh-temperature granulites from North China Craton. The spot ages obtained from U–Pb–Th electron microprobe method are also shown. The area covered by each analytical spot in the figure represents the cross-sectional area of the excitation volume by the electron beam. The numbers on the lower right indicate sample number (top) and grain analysis number (bottom).

geobarometry could not be applied to the samples because plagioclase is not in equilibrium with garnet and orthopyroxene judging from the textures.

8.4. Sapphirine–spinel geothermometer

Sapphirine and spinel in sample 06662 are included in sillimanite without other ferromagnesian minerals. The mineral pair can therefore be applied for temperature calculations. Das et al. (2003) formulated a new sapphirine–spinel geothermometer

using their experimental data at 9–12 kbar and 850–1100 °C. Application of the method yielded almost consistent results in the range of 830–820 °C. Recent experimental calibration of Sato et al. (2006) at 9–13 kbar and 950–1150 °C using natural UHT rocks yielded a temperature range of 910–940 °C, which is consistent with the results of Al-in-orthopyroxene geothermometer. Although these estimates are lower than the peak UHT range from other methods, the consistently high temperature obtained from various computations confirm the high temperature metamorphic history of these rocks.

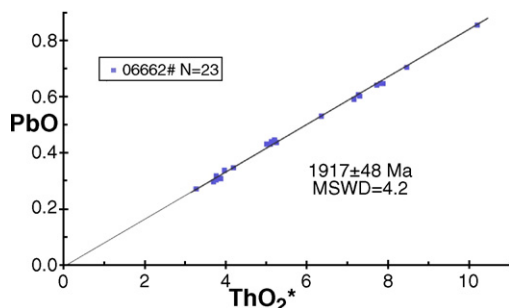


Fig. 15. PbO vs. ThO_2^* plots of monazite data from Mg–Al granulites of North China showing linear data trend and isochron with an age 1917 ± 48 Ma.

8.5. Garnet–cordierite geothermobarometer

As cordierite is obviously a product of retrograde metamorphism in the present rocks, this geothermometer mostly provides the post-peak conditions. Compositional data of rim garnet and corona cordierite were applied for the calculation. Several calibrations are available for this thermometer, although most of them yielded unusually low estimates as compared with the temperature estimates from other assemblages discussed in earlier sections. For example, the experimental geothermometer of Perchuk and Lavrent'eva (1983) gave a temperature range of 560–570 °C at 4 kbar when applied to a pair of rim garnet and adjacent cordierite in sample 06663. The experimentally calibrated geothermometer of Nichols et al. (1992) in FMASHZn system also gave a similar temperature range of 610–620 °C at the same pressure. The method of Bhattacharya et al. (1988), which is based on thermodynamic calculation of entropy and enthalpy, gave a temperature range of 640–650 °C at 4 kbar. These temperature estimates probably record the final closure temperature of Fe–Mg exchange between the two minerals. We also attempted pressure estimations using the garnet–cordierite–sillimanite–quartz geobarometer. The methods of Nichols et al. (1992), which were experimentally calibrated at 11 and 30 kbar and 950–1200 °C for both Fe and Mg end members, gave a pressure range of 4.0–4.6 kbar at 600 °C. It is obvious that the cordierite formation in these rocks occurred at a late stage of decompression, following the peak UHT metamorphism. As cordierite probably contains a few wt. % CO_2 or H_2O (cf. Table 7), the calculated pressures may shift toward higher pressure due to the effect of volatile components (e.g. Harley et al., 2002).

9. P – T grid for the North China UHT rocks

We evaluate below the possible P – T path of the sapphirine-bearing UHT granulites from North China using available petrogenetic grid in the FMAS system (Fig. 13). We adopted this system rather than KFMASH grid because biotite is mostly retrograde in origin and is not related to metamorphic reactions at UHT conditions. Low Zn and Cr contents in spinel and sapphirine further allowed us to use this grid. The pressure–temperature window in the figure was inferred from the Al content in the core of the orthopyroxene with sillimanite in sample 06662. The presence of equilibrium sapphirine–quartz

assemblage in the quartzo-feldspathic layer of sample 06660, formed under the stability field of sapphirine+quartz, is estimated as >1030 °C at 9.5 kbar (Hensen and Green, 1973) or >1050 °C at 11 kbar (Bertrand et al., 1991) and provides robust evidence for $T > 1000$ °C UHT peak metamorphism in the North China Craton. The formation of orthopyroxene–sillimanite–quartz assemblage in sample 06662 through the progress of reaction (2) suggests an isobaric cooling history from the stability field of sapphirine+quartz to orthopyroxene + sillimanite+quartz (Fig. 13). As the invariant point of [Spl] in the FMAS system corresponds to the P – T condition of 1050 °C at 10 kbar (Harley, 1998), we infer that the isobaric cooling took place at pressures higher than 10 kbar.

The sapphirine-bearing granulites discussed in this study exhibit various retrograde textures as indicated by reactions (3)–(5). One of the most spectacular reaction textures is the corona of orthopyroxene+cordierite around garnet (reaction (5); Fig. 6f). This is one of the most common decompression reaction textures documented from various granulite terrains in other parts of the world and the texture is regarded as a fingerprint of near isothermal decompression after peak metamorphism (Harley, 1989). Other reaction textures such as cordierite around garnet and cordierite between sillimanite and orthopyroxene also suggest near-isothermal decompression after the isobaric cooling event. Such isobaric cooling followed by isothermal decompression after the peak UHT metamorphism has been reported from other UHT terrains (e.g., Sajeew et al., 2004; Tateishi et al., 2004).

10. Monazite geochronology

Representative monazite grains in textural association with the sapphirine-bearing UHT mineral assemblage from Tuguiwula area of the North China Craton were dated using electron microprobe techniques for reconnaissance geochronology. More detailed geochronology work on both monazites and zircons is in progress. The analyses were performed *in situ* in thin sections of the same samples used for petrographic studies using an electron microprobe (JEOL JXA-8800) at the National Science Museum, Tokyo, Japan. The analyzed monazite grains occur within the matrix of the sample, in textural association with sapphirine, orthopyroxene and sillimanite. The theoretical concepts, analytical procedures and parameters for age computations followed those described in Santosh et al. (2003, 2005, 2006a, and references therein). The analytical data are summarized in Table 8.

The backscattered electron images (BEI) of representative monazite grains from the Mg–Al granulites are shown in Fig. 14, along with the computed spot ages. The monazite grains in the rocks range in size from 30 to 120 μm , and show euhedral to subhedral shapes. Many of the grains show a nearly uniform grain core (Fig. 14a,e) without complex growth patterns or zoning effects, suggesting crystallization from a single thermal event. However, some of them display a dark mantle of variable thickness surrounding the core (Fig. 14c,d).

The spot ages from monazites clearly indicate a Paleoproterozoic thermal history, ranging from 1726 to 1956 Ma. Majority

of the age values fall in the range of 1800–1950 Ma (cf. Table 8), with the notable absence of ages older and younger than the Paleoproterozoic. In PbO vs. ThO₂* diagram, the data define a linear trend (Fig. 15) with an isochron age of 1917 ± 48 Ma. In a previous study on monazites occurring as inclusions in garnet well as from the matrix of khondalites in North China, Santosh et al. (2006) recorded a similar age range (1766–2039 Ma), with the oldest apparent ages shown by cores of monazite grains that occur as inclusions within garnet. The outermost rims which isostructurally mantle the pre-existing monazites show slightly younger ages similar to those obtained from discrete monazite grains in the matrix assemblage. The monazite population from the khondalites studied by Santosh et al. (2006) defines two broadly sharp age peaks at 1927 ± 11 Ma and 1819 ± 11 Ma.

11. Discussion

11.1. Metamorphic *P–T* conditions

This is the first report on the occurrence of Mg–Al granulites within the khondalite belt in the northern margin of the North China Craton. The discovery of the stable coexistence of sapphirine and quartz and the occurrence of high alumina orthopyroxene + sillimanite + quartz assemblage in these rocks in our study provide robust evidence for extreme crustal metamorphism in this terrain. The highest alumina content in orthopyroxenes from these rocks (8.2 wt.%) is within the range for orthopyroxenes with 8–9.5 wt.% Al₂O₃ occurring in the orthopyroxene + sillimanite + garnet + quartz rocks from the Forefinger Point in Enderby Land, Antarctica, where Harley et al. (1990) estimated metamorphic *P–T* conditions of 10 ± 1.5 kbar and 950 ± 50 °C. The high Al₂O₃ content of orthopyroxenes from North China Craton also compare with similar, although slightly higher values in orthopyroxene porphyroblast from the UHT granulites of Ganguvarpatti (8.5–9.4 wt.%; Tamashiro et al., 2004) and Rajapalaiyam (8.5–9.7 wt.%; Tateishi et al., 2004) in southern India, where in all cases the metamorphic temperatures exceeded 1000 °C.

The temperatures computed from conventional geothermometers indicate that the peak metamorphism took place at $T \sim 970$ °C within the realm of UHT. However, the stability relations of some of the assemblages discovered in this study indicate that the peak metamorphic temperatures were much higher than this estimate. For example, the available phase equilibrium data suggest that sapphirine + quartz is stable at $T > 1030$ °C at 9.5 kbar (Hensen and Green, 1973) or $T > 1050$ °C at 11 kbar (Bertrand et al., 1991). Recent thermodynamic calculations in the KFMASH system also support the high-temperature nature of this assemblage ($T > 1005$ °C, Kelsey et al., 2004). Sapphirine + quartz assemblage similar to that reported in this study has also been documented from some granulite terrains and is taken as a key evidence for peak UHT metamorphism at $T > 1000$ °C. For example, sapphirine in equilibrium with quartz was reported from quartzo-feldspathic and pelitic granulites from the southern part of the Madurai Block around Rajapalaiyam in

southern India by Tateishi et al. (2004). In Rajapalaiyam, however, the assemblage is preserved only as rare inclusions in the core of coarse grained subhedral garnet. Similar sapphirine + quartz occurrences were also reported from some UHT terrains such as In Ouzzal (Ouzegane and Boumaza, 1996), Highland Complex (Sajeev and Osanai, 2004), and Lützow Holm Complex (Yoshimura et al., 2003). As compared to the rare nature of sapphirine + quartz assemblages in these reports, the common occurrence of this assemblage in the matrix of Mg–Al granulites from North China clearly prove their stability during peak metamorphic conditions. We therefore infer that the peak UHT metamorphism of the Mg–Al granulites in the khondalite belt in North China Craton occurred at $T > 1000$ °C and $P > 10$ kbar.

The oxygen fugacity conditions at the peak metamorphic stage exert a significant control on the stability of some of the UHT assemblages. *P–T* estimates should therefore take into consideration the *f*O₂ conditions. Hensen (1986) and Powell and Sandiford (1988) pointed out that at high oxygen fugacity conditions around the magnetite + ilmenite buffer, sapphirine and quartz probably coexist at temperatures as low as 850 °C. Although our samples contain magnetite lamella in ilmenite, detailed textural observations indicate that such exsolution probably took place during retrograde metamorphism and cooling. This observation is consistent with calculated temperature range from ilmenite–magnetite pairs in sample 06663 using the method of Spencer and Lindsley (1981), which gave a very low temperature value of ca. 500 °C. We therefore infer that magnetite was not present during the peak metamorphism of these rocks within the stability field of sapphirine + quartz, suggesting that oxygen fugacity was low. This inference is also consistent with the low Fe³⁺/(Fe²⁺+Fe³⁺) ratio of sapphirine coexisting with quartz in sample 06660 (0.12–0.21). Sapphirine + quartz association in our samples therefore clearly suggest UHT metamorphism in this area at $T > 1000$ °C. Grew et al. (2006) reported beryllium in sapphirine, which may lower the stability field of sapphirine toward lower temperatures. Although beryllium content in sapphirine has not been analyzed in this study, we regard that the beryllium content in our sapphirine is low as Be-bearing sapphirine has been reported in anatectic veins in UHT terrains (e.g., Napier Complex, Antarctica), but not from granulites.

In a previous study, Santosh et al. (2006) reported coexisting spinel + quartz from a locality in the Jining area adjacent to that of the present study within the khondalite belt of North China Craton. The study reported two types of spinel + quartz microstructural association with contrasting Zn contents in spinel. The spinel in equilibrium with quartz occurring as inclusions within garnet has high ZnO (up to 14.47 wt.%), indicating their stability prior to peak metamorphism. The spinel inclusions in garnet have X_{Mg} up to 0.498, and the X_{Zn} isopleths at X_{Mg} 0.5 define a high pressure pre-peak thermal history for the rocks. From the lack of biotite in most of the samples studied, Santosh et al. (2006) interpreted that the biotite was consumed through dehydration melting in the presence of sillimanite along a prograde *P–T* path to produce spinel + quartz + garnet + melt. This reaction took place at a pressure

above 12 kbar and temperature of ca. 900 °C, and the Zn in spinel was probably sourced from biotite during melting. The second category of spinel is characterized by low ZnO (down to 1.17 wt.%) and occurs in grain contact with quartz in the matrix assemblage formed during peak UHT conditions estimated as ca. 975 °C and 9 kbar, towards the low temperature side of spinel+orthopyroxene+quartz stability field, based on the petrogenetic grid of Das et al. (2001). Santosh et al. (2006) suggested this to be a clear but unique case of decompression through the high-(high Zn) and low-(low Zn) pressure spinel+quartz fields. The study showed that the prograde and peak metamorphic conditions of the North China khondalites traversed mostly through the spinel+quartz (extended) stability field. Further retrograde cooling and decompression are indicated by the late stage formation of cordierite, sillimanite and biotite bearing assemblage in the khondalites.

11.2. Timing of UHT metamorphism

U–Pb–Th electron microprobe data on monazites in textural association with UHT minerals in the matrix of the Mg–Al rock from North China define an age of 1917 ± 48 Ma, indicating Paleoproterozoic metamorphism. In a previous study of monazites associated with spinel+quartz microstructure within the khondalites of the Jining area in North West China Craton, Santosh et al. (2006) reported two age peaks at 1927 ± 11 and 1819 ± 11 Ma. The data were interpreted to indicate monazite growth during the prograde metamorphic event, as well as during the peak ultrahigh-temperature metamorphic event. The monazite grains with older ages formed during the prograde stage were taken over by garnet with new monazite crystallization in the matrix during the UHT event. Some of monazite grains within garnet developed thin overgrowths rimming the older cores during this thermal event. Experimental data on the closure temperature and Pb diffusivity in monazite (e.g., Cherniak et al., 2004) indicate that a 10 μm -sized monazite grain would have a Pb-closure temperature in excess of 900 °C/Ma. The monazite grains analyzed in the present study and those reported in Santosh et al. (2006) are larger than this size, and hence it is evident that their closure temperature exceeded the above temperature. The khondalites as well as the Mg–Al granulites experienced extreme metamorphism at $T > 950$ – 1000 °C, and therefore the monazite ages conclusively fix the timing of the UHT event in North China Craton to be Paleoproterozoic. It is possible that the 1819 ± 11 Ma obtained by Santosh et al. (2006) marks the retrograde event, since the 1927 ± 11 Ma obtained in that study and the 1917 ± 48 Ma in the present study are in close agreement, and therefore we interpret that the ca. 1.9 Ga event marks the timing of UHT metamorphism in this belt. The collisional orogeny which led to extreme crustal metamorphism in this belt correlates with the assembly of the Paleoproterozoic supercontinent Columbia (Santosh et al., 2006). The ca. 1.9–1.8 Ma ages obtained for these rocks coincide with the timing of accretion of the Columbia supercontinent as proposed by Rogers and Santosh (2002).

In a recent study, Wan et al. (2006) reported U–Pb zircon ages from various Palaeoproterozoic metasedimentary units in

the North China Craton. These rocks contain detrital igneous zircon grains with ages ranging from 3.0 to 2.1 Ga, with the majority displaying an age between 2.3 and 2.0 Ga. According to these authors, the latter age group represents a series of magmatic events that occurred between the formation of the Archean basement and deposition of the sediments. The metasedimentary rocks analyzed by Wan et al. (2006) occur as two associations; older volcano-sedimentary units that formed between 2.37 and 2.0 Ma, and younger sedimentary units that formed mainly between 2.0 and 1.88 Ga. All these rock associations were then subjected to metamorphism during the Late Palaeoproterozoic (1.88 and 1.85 Ga) tectonothermal event. Wan et al. (2006) correlate this age with a continent–continent collision event that led to the amalgamation of the North China Craton. Wan et al. (2006) did not analyze any UHT rocks from the North China Craton. However, the zircon SHRIMP ages for the Late Paleoproterozoic tectonothermal event (1.88–1.85 Ga) broadly correlate with the ca. 1819 Ma obtained from monazites by Santosh et al. (2006) from the khondalite belt in North China Craton.

In another recent study, Liu et al. (2006) analyzed U–Th–Pb in monazites from the low- to medium-grade Luliang and Wutai Complexes in the Trans-North China Orogen. The data revealed five age populations: (1) 1940–1938 Ma, (2) 1880–1847 Ma, (3) 1795–1755 Ma, (4) 1720–1703 Ma, and (5) ~1648 Ma. Of these, the first four have also been recorded in the Wutai Complex. The oldest age range of 1940–1930 Ma is interpreted to record the widespread emplacement of mafic dikes in rocks of the Luliang and Wutai Complexes. The age range of 1882–1822 Ma is interpreted as the age of the major metamorphic event caused by amalgamation between the Eastern and the Western Blocks. The age range of 1795–1755 Ma is consistent with emplacement of large scale unmetamorphosed mafic swarms at 1800–1765 Ma, interpreted as the time of post-orogenic extension. The 1720–1703 Ma and ca. 1648 Ma ages are considered to date later multiple stages of hydrothermal alteration, since monazites with such young ages occur only along fractures in older monazite grains and at the rims. Liu et al. (2006) considered their data to support a tectonic model for the evolution of the North China Craton involving the collision of discrete Eastern and Western Blocks along the Trans-North China Orogen at ~1.85 Ga and post-collisional extension in the period 1795–1755 Ma. Although Liu et al. (2006) did not study any UHT granulite assemblages, it is interesting to note that some of their monazite ages overlap with the age ranges from monazites in the UHT granulites of North China Craton obtained in the present study as well as those reported in Santosh et al. (2006).

11.3. Nature of P–T trajectory

Santosh et al. (2006) used the two types of spinel+quartz microstructures, together with other retrograde assemblages, in the khondalites to define the P–T trajectory of the metapelite belt in North China Craton. The P–T path constructed in their study traverses from the high-Zn spinel field (12 kbar and 900 °C) through the low-Zn spinel field (ca. 975 °C and 9 kbar)

revealing a clockwise trajectory, with retrograde cooling and decompression indicated by the late stage formation of cordierite, sillimanite and biotite bearing assemblages.

Zhao et al. (2006) proposed that the granulites of the North China Craton belong to different age groups and that they were generated during multiple geological events. They believe that the granulites along the northern margin of the Western and Eastern Blocks formed at 2.5 Ga, with anticlockwise P – T paths, whereas those along the northern margin of the Trans-North China Orogen formed during the collision between the Eastern and Western Blocks at 1.85 Ga with clockwise P – T paths. The P – T trajectories of both Santosh et al. (2006) and Zhao et al. (2006) for the Paleoproterozoic rocks in the North China Craton are essentially clockwise paths.

The P – T trajectory obtained in the present study for the Mg–Al granulites is also broadly similar, emerging from the stability field of sapphirine+quartz. As the melanocratic domain in sample 06662 contains orthopyroxene+sillimanite+quartz assemblage probably formed by reaction (2), the cooling path should have passed above the spinel invariant point in the FMAS grid (Fig. 13, >9 kbar), that defines an additional initial isobaric cooling component of about 100 °C from the stability field of sapphirine+quartz. This P – T condition is consistent with obtained temperature ranges from garnet–orthopyroxene, ternary feldspar, and sapphirine–spinel geothermometers. The rocks subsequently underwent near-isothermal decompression that formed cordierite+orthopyroxene corona around garnet (Figs. 6f and 7f) and cordierite between sillimanite and orthopyroxene (Figs. 6a and 7e).

The exhumation of most granulite terrains has followed either isothermal decompression (ITD) or isobaric cooling (IBC) path following peak metamorphism (Harley, 1989) with few exceptions. Some of the UHT granulites are characterized by several stages of ITD and IBC segments. Most of the IBC paths are considered to be retrograde segments of an anticlockwise P – T path while the ITD has been correlated to fast uplift along a clockwise path (Harley, 1989). Recently, Santosh and Sajeve (in press) reported anti-clockwise P – T paths in UHT granulites from the Palghat–Cauvery Shear Zone area, at the leading edge of the Proterozoic continental margin in southern India. Several UHT localities within the Madurai Block, south of the Palghat–Cauvery Shear Zone record clockwise P – T paths. Clockwise and anticlockwise P – T paths are generally determined on the basis of the nature of the prograde path. Santosh and Sajeve (in press) suggested that it is possible to have isobaric cooling and isothermal decompression retrograde trajectories associated with both clockwise and anticlockwise loops. It is often true that most granulites characterized by anticlockwise P – T path first show isobaric cooling as a result of thermal relaxation of the overthickened crust, with or without other tectonic factors. The Mg–Al granulites of present study are located along a continental collision zone, where the tectonic scenario was characterized by deep subduction and subsequent rapid exhumation. Most collision-type metamorphic belts have had a high pressure history, with an anticlockwise kink-point at around 10 kbar (Maruyama et al., 2004). The high pressure signature is

subsequently erased due to extensive retrogression, sometimes involving hydration. Although the high pressure signature and anticlockwise component are not evident in the present case, it is possible that the P – T trajectory defined in this study with an initial isobaric cooling component followed by rapid isothermal decompression along a clockwise path is the partial trace of a larger loop which has an anticlockwise kink-point since the sedimentary pile comprising the khondalite belt and associated Mg–Al lithologies might have undergone initial heating and subsequent deep burial prior to UHT metamorphism and retrograde isothermal decompression. Further studies, particularly careful examination of inclusion minerals, would help unravel the prograde metamorphic history of these rocks in more detail.

The extreme temperatures attained during metamorphism of the Mg–Al granulites would suggest convective thinning or detachment of the lithospheric thermal boundary layer during or after crustal thickening (cf. Harley, 1998). The study area within the northern margin of the North China Craton defines a collisional orogenic belt where continental collision and deep subduction were possibly followed by extensional collapse of the orogen. Evidence for the extensional phase is provided by the several mafic magmatic intrusives of Paleoproterozoic age that are seen emplaced within the granulite facies supracrustals. The thermal anomaly displayed by the UHT granulites may thus correlate with asthenospheric upwelling and emplacement of mafic magmas which provided the heat input to trigger extreme crustal metamorphism.

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