Metamorphic $P$–$T$ path and tectonic implications of medium-pressure pelitic granulites from the Jiaobei massif in the Jiao-Liao-Ji Belt, North China Craton

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A R T I C L E   I N F O

Article history:
Received 20 June 2012
Received in revised form 1 August 2012
Accepted 14 August 2012
Available online 23 August 2012

Keywords:
Medium-pressure pelitic granulites
$P$–$T$ path
Paleoproterozoic
Jiao-Liao-Ji Belt
North China Craton

A B S T R A C T

Both medium- and high-pressure pelitic granulites were found in the Jiaobei massif in the southern segment of the Jiao-Liao-Ji Belt, which is a Paleoproterozoic tectonic belt in the Eastern Block of the North China Craton. Studies on mineral assemblages and metamorphic reaction textures reveal that the medium-pressure pelitic granulites from the Jiaobei massif underwent four metamorphic stages: (1) the prograde stage (M1) represented by mineral inclusions within the core of garnet grains; (2) the peak stage (M2) indicated by the formation of sillimanite + K-feldspar and the growth of inclusion-free garnet mantle; (3) the decompression stage (M3) characterized by the cordierite + sillimanite symplectite, and (4) the decompression-cooling stage (M4) forming cordierite coronas. An integrated study of mineral reaction textures, mineral chemistry and pseudosection modelling in the NCFKMASHTO system with the THERMOCALC technique constrains the $P$–$T$ conditions of 9.4–10 kbar and 870–900 °C for the M2 stage, 6.3–6.6 kbar and 840–900 °C for the M3 stage, and 4–5.2 kbar and 815–830 °C for the M4 stage. The $P$–$T$ conditions of the early prograde (M1) assemblage cannot be estimated due to reworking at later metamorphic stages (M2–M4). The mineral assemblages of the M2–M4 stages and their $P$–$T$ conditions define a clockwise $P$–$T$ path involving isothermal decompression (ITD) and subsequent decompression-cooling for the medium-pressure pelitic granulites, which are consistent with a continent–continent collision environment. This suggests that the Jiao-Liao-Ji Belt was not formed simply by the closure of a Paleoproterozoic rift basin as conventionally considered, but the tectonic evolution of its southern segment must have been involved in subduction and/or continent–continent collisional processes. Based on this study and previous data, we propose that the Jiao-Liao-Ji Belt represented a Paleoproterozoic rifting- and collision-belt along which the Langgang and Langrim Blocks amalgamated to form the Eastern Block of the North China Craton.

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1. Introduction

Granulite facies rocks are representative lithologic associations in many orogenic belts, and their pressure–temperature–time paths allow interpretation of the tectonic settings and processes operative in the formation and evolution of orogenic belts (Thompson and England, 1984; Harley, 1989, 1992; Brown, 1993; Clarke et al., 2000; O’Brien and Rötzler, 2003; Touret and Huizenga, 2012). Mafic and pelitic granulites are of particular interest in this regard, because they have mineral assemblages that are suitable for estimating $P$–$T$ conditions of metamorphism and in some cases, they preserve textural evidence that can be used to infer metamorphic reaction relations and their relative timing. Therefore, an integrated petrological, structural, metamorphic and geochronological study on mafic and pelitic granulites will not only assist in the inference of the tectonic mechanisms and processes responsible for the granulite formation and the role of granulite terranes in the larger context of growth and development of the continental crust, but will also provide insights into understanding the tectonic environments and processes of orogenic belts.

Mafic and pelitic granulites are widespread in the North China Craton (NCC) and the reconstruction and interpretation of their $P$–$T$ paths are critical to understanding the tectonic evolution of the craton. In the past decade, extensive investigations on metamorphic $P$–$T$ paths of mafic and pelitic granulites from the craton have been carried out. Combined with petrological, structural, geochemical and geochronological data, these investigations have led to recognition of three Paleoproterozoic linear structural belts,
named the Trans-North China Orogen, Khondalite Belt (also called the Inner Mongolia Suture Zone; Santosh, 2010; Santosh et al., 2010, 2012a,b) and Jiao-Liao-Ji Belt (Fig. 1; Zhao et al., 1998, 2001, 2005, 2006a,b, 2007; Zhao, 2009; Lu et al., 1996, 2004, 2006, 2008; Li et al., 2001a,b, 2005, 2006, 2011, 2012; Liu et al., 2011a,b; Guo et al., 2002, 2005; Faure et al., 2004, 2007; Kröner et al., 2005, 2006; Trap et al., 2007, 2008, 2009, 2011; Wilde et al., 2002, 2004; Wilde and Zhao, 2005; Kusky, 2011; Zhai and Santosh, 2011a,b; Zhou et al., 2004, 2008a; Santosh et al., 2010, 2012a,b). The Khondalite Belt and Trans-North China Orogen are considered to be two Paleoproterozoic collisional orogenic belts, with the former formed by the collision of the Yinshan and Ordos Blocks to form the coherent Western Block at \( \sim 1.95 \) Ga (Santosh et al., 2006, 2007a,b, 2008, 2009a,b, 2010, 2012a,b; Yin et al., 2009, 2011; Zhao, 2009; Tsunogae et al., 2011), and the latter formed by amalgamation of the Western and Eastern Blocks to form the NCC at \( \sim 1.85 \) Ga (Wilde et al., 2002, 2004; Zhao et al., 2002, 2006a,b, 2007, 2008a,b; Kröner et al., 2005, 2006; Guo et al., 2005; Liu et al., 2006; Zang et al., 2007, 2009). However, the tectonic nature of the Paleoproterozoic Jiao-Liao-Ji Belt still remains controversial, with some people arguing that its formation and evolution invoked the opening and closing of a Paleoproterozoic intra-continental rift in the Eastern Block of the NCC, whereas others believe that it involved the development of an island arc and its collision with continental blocks in the Paleoproterozoic. These controversies reflect the lack of geological data for the Jiao-Liao-Ji Belt.

Recently, medium- and high-pressure pelitic granulites have been discovered in the Jiaobei massif in the southern segment of the Jiao-Liao-Ji Belt (Zhou et al., 2004, 2007, 2008a; Tam et al., 2011, 2012). The metamorphic evolution of these pelitic granulites can provide important insights into understanding the tectonic setting and processes of the Paleoproterozoic Jiao-Liao-Ji Belt. For this reason, Tam et al. (2011, 2012) and Zhou et al. (2008a) carried out detailed geochronological and metamorphic studies on the high-pressure pelitic granulites from the Jiaobei massif, and the results show that these rocks underwent high-pressure granulite facies metamorphism at \( \sim 1.90 \) Ga, characterized by a clockwise \( P-T \) path involving isothermal decompression, which are consistent with subduction and/or continent–continent collision. This led Tam et al. (2012) to propose that the Jiao-Liao-Ji Belt represented a Paleoproterozoic continent–continent collisional belt along which two Archean blocks (called the Longgang Block in the north and the Langrim Block in the south) amalgamated to form the Eastern Block of the NCC. However, high-pressure pelitic granulites have only been found in one location in the Jiaobei massif, whereas most pelitic granulites are medium-pressure pelitic granulites. Therefore, the metamorphic evolution of the medium-pressure pelitic granulites is critical to test the above model. This forms the justification for this study in which we carry out an integrated study on metamorphic reaction textures, mineral chemistry and pseudosection modelling in order to reconstruct the metamorphic \( P-T \) path of the medium-pressure pelitic granulites, which is important for inferring the tectonic setting and processes of the Jiao-Liao-Ji Belt.

2. Geological background

2.1. Regional geology

As mentioned above, the Precambrian basement of the NCC can be divided into the Eastern and Western Blocks, which collided along the Trans-North China Orogen at \( \sim 1.85 \) Ga (Fig. 1; Wilde et al., 2002, 2004; Zhao et al., 2001, 2002, 2005, 2006a,b, 2007, 2008a,b;
et al., 2005, 2006; Guo et al., 2005; Zhang et al., 2007, 2009; Liu et al., 2006; Li et al., 2010). The Western Block is subdivided into the Yinshan Block in the north and the Ordos Block in the south, which amalgamated along the E-W trending Khondalite Belt, at ~1.95 Ga (Fig. 1; Zhao et al., 2005; Guo et al., 2002; Santosh et al., 2006, 2007a,b, 2009a,b, 2010, 2012a,b; Xie et al., 2006a,b; Yin et al., 2009, 2011; Santosh, 2010; Santosh and Kusky, 2010), whereas the Eastern Block is subdivided into the Longgang Block in the northwest and the Langrim Block in the southeast, separated by the Paleoproterozoic Jiao-Liao-Ji Belt (Fig. 1; Zhang and Yang, 1988; Liu et al., 1992; Jahn et al., 1988; Li et al., 2001a,b, 2005, 2011; Luo et al., 2006; Wu et al., 2005; Wan et al., 2005; Li and Zhao, 2007; Zhou et al., 2008a; Tam et al., 2011, 2012).

The Paleoproterozoic Jiao-Liao-Ji Belt consists of a northern segment extending from southern Jilin, through northern Liaoning, into the Liaodong Peninsula, and a southern segment that extends across the Bohai Sea into the Jiaobei massif (Figs. 1 and 2). The belt is composed predominantly of sedimentary and volcanic successions that were deformed and metamorphosed under greenschist to lower amphibolite facies conditions (Fig. 2). The sedimentary and volcanic successions include the Macheonayeong Group in North Korea, the Ji’an and Laoling Groups in southern Jilin, the North and South Liaobe Groups in the Liaodong Peninsula, the Fenzishan and Jingshang Groups in the Jiaobei massif, and the Wuhe Group in Anhui Province (Fig. 2). These groups display similar stratigraphic successions that are transitional from a basal clastic-rich sequence and a lower bimodal-volcanic sequence, through a middle carbonate-rich sequence, to an upper pelite-rich sequence (Li et al., 2001a,b, 2004a, 2005, 2012; Luo et al., 2004, 2008; Zhao et al., 2005; Lu et al., 2006; Li and Zhao, 2007; Zhou et al., 2007, 2008a,b; Tam et al., 2011, 2012). Stratigraphically, the North Liaobe Group in Liaoning is well correlated with the Laoling Group in southern

**Fig. 2.** Map of the Paleoproterozoic Jiao-Liao-Ji Belt in the Eastern Block of the North China Craton showing the distribution of the Wuhe Group in Anhui, Fenzishan and Jingshan Groups in Eastern Shandong, South and North Liaobe Groups in Liaoning, Laoling and Ji’an Groups in Southern Jilin, and Macheonayeong Group in North Korea (Zhao et al., 2005). The dotted box indicates the location of the Jiaobei massif in the southern segment of the Jiao-Liao-Ji Belt.
Jilin and the Fenzishan Group in eastern Shandong, whereas the South Liaobe Group is well correlated with the Ji'an Group and the Jingshan Group (Fig. 2; Zhao et al., 2005). Therefore, the Jiao-Liao-Ji Belt itself can be further subdivided into a northern belt, which comprises the Fenzishan, North Liaobe and Laoling Groups, and a southern belt that consists of the Jingshan, South Liaobe and Ji'an Group (Fig. 2). Separating the two belts are ductile shear zones and faults (Lu et al., 1996; He and Ye, 1998; Li et al., 2005; Zhao, 2009).

Associated with sedimentary and volcanic successions are mafic intrusions and granitoids (Fig. 2). The mafic intrusions are dominated by dolerites and gabbrons metamorphosed in sub-greenschist-amphibolite facies (Zhao et al., 2005; Li and Zhao, 2007), while the granitoids include pre-tectonic monogranitic gneisses and post-tectonic alkaline syenites, rapakivi granites and porphyritic monzogranites, mainly exposed in southern Jilin and eastern Liaoning (Lu et al., 2004; Zhao et al., 2005). Available SHRIMP U–Pb zircon data reveal that monograni tic gneisses were emplaced at 2.17–2.14 Ga and metamorphosed at ~1.91 Ga, while the post-tectonic porphyritic monzogranites and alkaline syenites were emplaced at c. 1.88–1.84 Ga (Li and Zhao, 2007).

2.2. Jiabei massif

The Jiabeici massif is a Precambrian terrain located in the northern part of eastern Shandong Province in the southern segment of the Jiao-Liao-Ji Belt (Fig. 3). It is bounded by the Tanlu Fault to the west and the Wulian–Yantai Fault to the east (Fig. 2), and consists mainly of the Precambrian basement rocks, overlaid by the Mesoproterozoic Penlai Group and the Mesozoic sedimentary rocks and igneous intrusions (Fig. 3). The Precambrian basement rocks include the late Archean rocks and the Paleoproterozoic Jingshan and Fenzishan Groups (Fig. 3). The late Archean rocks are dominated by tonalitic–trondhjemitic–granodioritic (TTG) gneisses distributed mainly in the eastern and the central parts of the Jiabeici massif and are tectonically in contact with lenses of mafic granulites and amphibolites (Fig. 3). Recent geochronological data indicate that the late Archean TTG gneisses were emplaced in the period between 2.69 Ga and 2.54 Ga (Tang et al., 2007; Zhou et al., 2008a), but the protolithic ages of the associated mafic granulites and amphibolites have not been well constrained.

Overlying the late Archean basement rocks are the Paleoproterozoic Jingshan and Fenzishan Groups that have been
metamorphosed from amphibolite to granulite facies (Fig. 3; Li et al., 2005, 2012). Their stratigraphic sequences are transitional from a lower arkose- and volcanic-rich sequence, through a middle carbonate-rich succession, to an upper argillaceous-rich sequence (Fig. 3; Li et al., 2005, 2012). The Jingshan and Fenzishan Groups are considered to have similar protoliths except that the former contain more volcanic materials. Both the Jingshan and Fenzishan Groups comprise pelitic schist, pelitic gneiss, medium- and high-pressure pelitic granulite, calc-silicate rock, marble and minor amphibolite and mafic granulite (Fig. 3; Li et al., 2005, 2012). Available isotopic data indicate that the Jingshan and Fenzishan Groups underwent a peak metamorphic event at 1.93–1.90 Ga and a post-peak metamorphic event at 1.86–1.8 Ga (Zhou et al., 2008a; Tam et al., 2011).

The Meso–Neo-protrozoic Penlai Group, predominated by slates, quartzite and meta-limestone, is unconformably overlying the Paleoproterozoic Jinshan and Fenzishan Groups (Fig. 3; Faure et al., 2004; Zhou et al., 2008b) and the key sources of the Penlai Group came from the late Archean rocks and the Paleoproterozoic Jinshan and Fenzishan Groups (Zhou et al., 2008b). Also unconformably overlying the Paleoproterozoic Jinshan and Fenzishan Groups are Mesozoic sedimentary–volcanic successions which are associated with the Mesozoic granitoids that were emplaced at 160–150 Ma and 132–120 Ma (Zhang et al., 2003; Wang et al., 1998).

Both medium- and high-pressure pelitic granulites have been found in the Fenzishan Group of the Jiaobei massif, but as mentioned earlier, the high-pressure pelitic granulites are very limited and so far have only been found on one outcrop close to Taipingzhuang Village (Zhou et al., 2004, 2008a; Tam et al., 2012). In contrast, the medium-pressure pelitic gneisses and granulites are widespread in the Jiaobei massif. In this study, we collected a number of fresh medium-pressure pelitic granulite samples in the Laixi area (Figs. 3 and 4a and b). The studied medium-pressure pelitic granulites are medium- to coarse-grained and show a gneissosity, and in places they are intruded by S-type granites.

3. Petrography and metamorphic stages

Six samples of medium-pressure pelitic granulites (JB04-1-6) were collected from the Laixi area for detailed petrographic studies, and they are composed mainly of garnet (10–15%), biotite (15–20%), cordierite (15–25%), sillimanite (~2%), plagioclase (5–15%), K-feldspar (15–20%) and quartz (10–20%), with minor accessory minerals of ilmenite, zircon, monazite and apatite. Based on mineral reaction relationships, four metamorphic stages are recognized: the prograde (M1), peak (M2), post-peak decomposition (M3), and decomposition-cooling (M4) stages.

3.1. Prograde (M1) stage

The prograde (M1) stage is represented by the enclosed minerals within the cores of porphyroblastic garnet (grt), which is interpreted to grow at equilibrium with the inclusion-type minerals at the early metamorphic stage (Fig. 4a, e and f). The inclusion-type biotite (bi) is usually a fine-grained elongated lath in various sizes, while the inclusion-type quartz (q) and plagioclase (pl) appear as irregular grains within garnet grains (Fig. 4a, e and f). Generally, muscovite and staurolite are two of the common pelitic minerals present under medium-grade metamorphic conditions, but they are major reactants in most reactions and might have been entirely consumed during peak metamorphism (Yardley, 1989). This is the reason why neither inclusion-type muscovite nor staurolite is observed within the cores of garnet in the studied rocks, though they may have been mineral phases of the M1 assemblage. Therefore, the representative mineral assemblage of the prograde (M1) stage in the medium-pressure pelitic granulites should be “garnet core (grt) + biotite (bi) + plagioclase (pl) + quartz (q) + ilmenite (ilm) + muscovite (mu) + staurolite (st).”

3.2. Peak (M2) stage

The peak (M2) stage is featured by the formation of sillimanite + K-feldspar and the growth of garnet; the latter is represented by the inclusion-free garnet grain or inclusion-free garnet mantled domains (Fig. 4a and f), and it is interpreted to have grown coeval with the matrix minerals of plagioclase (pl), K-feldspar (Kfdm), biotite (bi) and quartz (qm) (Fig. 4a–c and f). Sillimanite (sillm) appears as fibrous crystals coexisting with K-feldspar, biotite, plagioclase and quartz in the matrix (Fig. 4b and c). Thus, the representative mineral assemblage of the peak (M2) stage is “garnet (grt)+ sillimanite (sillm)+ K-feldspar (Kfdm)+ biotite (bi)+ plagioclase (pl)+ quartz (qm)+ ilmenite (ilm).”

The generation of the inclusion-free garnet and sillimanite + K-feldspar was probably produced by the following reaction (Schumacher et al., 1990):

\[
\text{Muscovite} + \text{staurolite} + \text{quartz} \rightarrow \text{sillimanite} + \text{K-feldspar} + \text{garnet} + H_2O
\]

3.3. Post-peak decomposition (M3) stage

The post-peak decomposition (M3) stage is represented by the formation of the cordierite + sillimanite symplectite, which usually occur in the matrix with tiny biotite (bi) and K-feldspar (Kfdm) inclusions that formed at the peak (M2) stage (Fig. 4b–f). In most cases, the cordierite + sillimanite symplectites appear in contact with garnet (grt), plagioclase (pl), K-feldspar (Kfdm), quartz (qm) and ilmenite (ilm) (Fig. 4b–f). Therefore, the representative mineral assemblage locally equilibrated at the M3 stage should be cordierite (cd) + sillimanite (sill) + garnet (grt) + plagioclase (pl) + K-feldspar (Kfdm) + quartz + ilmenite. The cordierite + sillimanite symplectite and K-feldspar (Kfdm) might have formed by a reaction between biotite and quartz as follows (Lu et al., 1996; Yardley, 1989):

\[
\text{Biotite} + \text{quartz} \rightarrow \text{sillimanite} + \text{cordierite} + \text{K-feldspar} + H_2O
\]

3.4. Decompression-cooling (M4) stage

The decompression-cooling (M4) stage is characterized by the development of the inclusion-free cordierite corona (cd) around garnet grains (Fig. 4e and f). The inclusion-free cordierite (cd) corona is interpreted to have been generated after the formation of the cordierite (cd) + sillimanite symplectite because it is found to separate garnet rims (grt) from the cordierite + sillimanite symplectite (Fig. 4e and f). The following reaction between garnet and symplectite sillimanite (sill) may explain the development of the inclusion-free cordierite corona (Currie, 1971; Santosh, 1987; Yardley, 1989; Harley and Hensen, 1990; Deer et al., 1992; Spear et al., 1999):

\[
\text{Garnet} + \text{symplectic} \rightarrow \text{sillimanite} + \text{quartz} \rightarrow \text{coronitic cordierite}
\]
Fig. 4. (a) A garnet grain enclosing plagioclase, biotite, quartz, ilmenite and zircon in the cores and surrounded by biotite, quartz, K-feldspar (BSE image). (b) Sillimanite (sill, sub) appearing as fibrous crystals in the matrix or relic crystals in the cordierite + sillimanite symplectites, in contact with small matrix-type biotite (bi, sub) and large retrograded biotite (bi, sub) (cross-polarized image). (c) Plagioclase, K-feldspar, quartz and the cordierite + sillimanite symplectite in the matrix (cross-polarized image). (d) The cordierite + sillimanite symplectites in contact with small relic biotite, quartz and ilmenite in the matrix (plane-polarized image). (e) A K-feldspar inclusion and small relic biotite flakes within the cordierite + sillimanite symplectites and enclosed plagioclase, biotite, quartz within a garnet grain surrounded by cordierite corona and plagioclase (BSE image). (f) Cordierite corona separating a garnet grain from the cordierite + sillimanite symplectite (plane-polarized image). [grt: garnet; sill: sillimanite; bi: biotite; pl: plagioclase; cd: cordierite; Kfd: K-feldspar; q: quartz; ilm: ilmenite; Zr: zircon].
biorite flakes were mostly likely to have formed from the following
decompression-cooling reaction (Deer et al., 1992):

\[ \text{Garnet} + K\text{-feldspar} + H_2O \rightarrow \text{biotite} + \text{quartz} \]  

Therefore, the mineral assemblage locally equilibrated at the
(M4) stage is corionic cordierite (cd) + garnet (gt) + biotite
(bm) + plagioclase rim (pl) + K-feldspar (Kf) + quartz
(qm) + ilmenite (ilmn) (4).

### 4. Mineral chemistry

Quantitative mineral analysis on the selected medium-pressure
pelitic granulites were carried out at the Institute of Geology and
Geophysics, Chinese Academy of Sciences in Beijing, the Guangzhou
Institute of Geochemistry, Chinese Academy of Sciences and the
Institute of Earth Sciences, Academia Sinica in Taipei. Electron
Probe Micro-Analysis (EPMA) with JOEL JXA-8100 electron micro-
probe and a JEOL JXA-8500F field emission electron probe were
applied. The former electron probe operated at an accelerating
voltage of ~15 kV, using a probe current of ~10 nA, with
counting times of 20 s for peaks and 10 s for each background,
while the latter one operated at an accelerating voltage of ~12 kV,
using a beam current of ~6 nA, with counting times of 10 s and 5 s
for each element and upper/lower baselines respectively. The
collected data were then calibrated with natural and synthetic phases
as standards from an online ZAF-type correction and representa-
tive compositions of key minerals garnet, plagioclase, biotite and
cordierite are listed in Tables 1–4.

#### 4.1. Garnet

Four distinct textural types of garnet are analysed: (I) garnet core
(gt) with inclusion-type minerals, (II) clean garnet core or mantle
(gtm) without any inclusions or garnet rim in contact with plagi-
oclass, K-feldspar, sillimanite and quartz, (III) garnet rims (gtt)
in contact with “cordierite + sillimanite” symplectites, and (IV) garnet
rim (gtr) surrounded by corionic cordierite or biotite (Table 1).
It is believed that most of garnet cores have been overprinted by
the later metamorphism, and thus no reliable composition of gar-
net cores forming at the M1 stage could be preserved, and thus
Table 1 only lists the compositions of garnet mantles and garnet
rims. As shown in Table 1, both garnet mantles and garnet rims are
dominated by almandine (0.667–0.756) and pyrope (0.214–0.271),
with minor grossular (0.036–0.050), spessartine (0.011–0.018) and
andradite (0–0.06). Most garnet cores are relatively enriched in
Mg in the mantle and relatively enriched in Fe in the rim, with a
rimward decrease in pyrope content (from 0.258–0.271, through
0.214–0.220, to 0.184–0.215) and a rim-ward increase in almandine
content (from 0.667–0.680, through 0.724–0.727, to 0.729–0.770)
(Table 1). From the mantle to rim, the spessartine content varies
from 0.011 to 0.018, but the grossular content gradually decreases
(from 0.047 to 0.050, 0.044 to 0.046 and 0.036 to 0.041) (Table 1),
which is interpreted to be consistent with isothermal decompres-
sion (ITD) after the peak metamorphism.

#### Fig. 5 is a core–rim analysis over a profile of a selected gar-
net grain (900 µm) that has a core with mineral inclusions, an
inclusion-free mantle and a rim surrounded by cordierite corona.
The garnet exhibits a weak compositional zoning, with a flat profile
from the core to mantle and slightly higher XAlm and lower Xsp at...
Table 2
Representative compositions of plagioclase (cations are calculated based on 8 oxygens).

<table>
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<td>99.19</td>
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<td>98.81</td>
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</table>

Si: inclusion-type plagioclase enclosed in garnet core; Pl: core of matrix-type plagioclase in contact with sillimanite, K-feldspar and quartz; Pl: rim of matrix-type plagioclase in contact with cordierite, garnet and large biotite.

Xₕᵢ = Ca/(Ca + Na + K); Xₜᵢ = Na/(Ca + Na + K); X₅ᵢ = K/(Ca + Na + K).

Mineral symbols are after Kretz (1983).

Table 3
Representative compositions of biotite (cations are calculated based on 11 oxygens).

<table>
<thead>
<tr>
<th>Sample</th>
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<th>Bi₃</th>
<th>Bi₄</th>
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<tr>
<td></td>
<td>JB04-4</td>
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<td>JB04-1</td>
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<tr>
<td>SiO₂</td>
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<td>36.90</td>
<td>36.34</td>
<td>36.32</td>
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<tr>
<td>Al₂O₃</td>
<td>18.15</td>
<td>18.51</td>
<td>17.57</td>
<td>17.46</td>
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<tr>
<td>Cr₂O₃</td>
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<td>0.107</td>
<td>0.116</td>
<td>0.060</td>
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<tr>
<td>FeO</td>
<td>11.87</td>
<td>12.72</td>
<td>15.12</td>
<td>15.47</td>
</tr>
<tr>
<td>MnO</td>
<td>0.015</td>
<td>0.015</td>
<td>0.014</td>
<td>0.033</td>
</tr>
<tr>
<td>MgO</td>
<td>14.59</td>
<td>14.47</td>
<td>11.71</td>
<td>11.50</td>
</tr>
<tr>
<td>CaO</td>
<td>0.014</td>
<td>0.003</td>
<td>0.011</td>
<td>0.002</td>
</tr>
<tr>
<td>Na₂O</td>
<td>0.176</td>
<td>0.144</td>
<td>0.136</td>
<td>0.151</td>
</tr>
<tr>
<td>Total</td>
<td>95.25</td>
<td>96.89</td>
<td>94.62</td>
<td>94.75</td>
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</table>

Si: fine-grained inclusion-type biotite lath within porphyroblastic garnet; Bi₂: small matrix-type biotite in contact with sillimanite and "cordierite + sillimanite" symplectite; Bi₃: coarse-grained biotite grown at the rim of garnet and in contact with K-feldspar, quartz and/or cordierite.

Fe³⁺ is the K-feldspar core playing a role in the decomposition of the rim (Fig. 5). The non-pronounced compositional variation from the core to the rim might have resulted from the re-equilibration during the peak (M2) metamorphism, while the higher Xₐₘₐ and lower Xₚₑᵣ at the rim are interpreted as the results of ion diffusion with the surrounding ferromagnesium minerals like biotite after the peak metamorphism (Fig. 5).

4.2. Plagioclase

Three major textural types of plagioclase were analysed: (I) inclusion-type plagioclase (p₁) enclosed in the garnet core, (II) core of matrix-type plagioclase (p₂) in contact with sillimanite, K-feldspar and quartz, and (III) rim of matrix-type plagioclase (p₃) in contact with cordierite, garnet and biotite (Table 2). All the three types of plagioclase are anorthite-rich (An₇₁–₈₀Ab₂₀–₂₉), with the inclusion-type plagioclase (p₁) in the garnet core displaying the greatest variation in anorthite (0.740–0.801) and albite (0.199–0.258) contents (Table 2). There is a pronounced variation in anorthite (An) contents between the core and rim of the matrix-type plagioclase, with the former’s Xₐₘ values ranging between 0.707 and 0.721, whereas the Xₐₘ range of the latter is between 0.743 and 0.778 (Table 2). A core-to-rim increase in Ca is considered as a result of garnet–plagioclase ion exchange during the decompression stages (M3–M4).

4.3. Biotite

Biotite was analysed for three major textural varieties: (I) fine-grained inclusion-type biotite (b₁) within porphyroblastic garnet, (II) matrix-type biotite (bₘ₁) in contact with the cordierite+sillimanite symplectite, and (III) large retrograded biotite (bₘ₂). The three varieties of biotite display differences in Al₂O₃, Xₐₑᵣ and TiO₂. The inclusion-type biotites (b₁) show higher Al₂O₃ contents (18.15–18.51) and lower Xₐₑᵣ values (0.313–0.330) than other two types of biotites, whose Al₂O₃ and Xₐₑᵣ range between 17.46–17.77 and 4.20–4.78, respectively (Table 3). The TiO₂ contents vary relatively much in the inclusion-type biotite (b₁), with a range of 2.789–3.508 (Table 3), whereas the matrix-type biotite (bₘ₁) in contact with the cordierite+sillimanite symplectite show a narrow TiO₂ range between 4.211 and 4.348, much higher than the TiO₂ contents (3.073–3.675) of the retrograded biotite (bₘ₂).

4.4. Cordierite

Cordierite was analysed for two textural types: (I) symplectic cordierite (c₁) from the cordierite+sillimanite symplectite, and (II) coronic cordierite (c₂) surrounding garnet crystal (Table 4). They are both Mg-rich and have a typical Si:Al ratio of ~5:4, with Si ranging from 5.035 to 5.123 and Al varying from 3.867 to 3.955 (Deer et al., 1992). There is a slight variation in Xₐₑᵣ value between symplectic and coronic cordierites, with the former having higher Xₐₑᵣ (0.29–0.331) than the latter (0.231–0.267), probably reflecting an ion exchange between biotite and garnet (Table 4).

<table>
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<tr>
<th>Sample</th>
<th>Cd₁</th>
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<tbody>
<tr>
<td>SiO₂</td>
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</tr>
<tr>
<td>TiO₂</td>
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<tr>
<td>Al₂O₃</td>
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<td>5.07</td>
</tr>
<tr>
<td>Cr₂O₃</td>
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</tr>
<tr>
<td>MnO</td>
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<td>0.00</td>
</tr>
<tr>
<td>FeO</td>
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</tr>
<tr>
<td>MgO</td>
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<td>0.00</td>
</tr>
<tr>
<td>Fe₃+</td>
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<td>0.00</td>
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<tr>
<td>Fe²⁺</td>
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<td>0.00</td>
</tr>
<tr>
<td>Total</td>
<td>11.00</td>
<td>11.00</td>
</tr>
</tbody>
</table>

Xₐₑᵣ = Fe²⁺/(Fe²⁺ + Mg²⁺). Fe³⁺ is derived from scheme of Droop (1987). Mineral symbols are after Kretz (1983).

**Table 4** Representative compositions of cordierite (cations are calculated based on 18 oxygens).
5. **Pseudosection modelling**

5.1. **P–T pseudosection modelling**

Based on the mineral assemblages and chemical compositions of selected sample (JB04–5), biotite–cordierite–garnet pelitic granulite, a P–T pseudosection in the \( \text{Na}_2\text{O}-\text{CaO}-\text{K}_2\text{O}-\text{FeO}-\text{MgO}-\text{Al}_2\text{O}_3-\text{SiO}_2-\text{H}_2\text{O}-\text{TiO}_2-\text{Fe}_2\text{O}_3 \) (NCKFMASHTO) system was constructed. This P–T pseudosection is a mineral equilibrium modelling in which the chosen bulk-rock composition was used to predict various metamorphic mineral assemblages at equilibrium within a certain pressure–temperature (P–T) range of 3–11 kbar and 650–950 °C (Holland and Powell, 1998; White et al., 2001, 2003, 2007; Clark and Hand, 2010). The powder of the chosen sample was analysed with X-ray fluorescence spectrometry (XRF) with standard wet chemical methods to get the bulk-rock composition at the Northwest University (Xi’an). The bulk-rock composition in wt%, contains \( \text{SiO}_2 = 55.82, \text{Al}_2\text{O}_3 = 22.65, \text{CaO} = 0.96, \text{MgO} = 4.68, \text{Fe}_2\text{O}_3 = 10.72, \text{Na}_2\text{O} = 0.39, \text{TiO}_2 = 0.67, \text{MnO} = 0.12, \text{K}_2\text{O} = 3.01 \) and \( \text{P}_2\text{O}_5 = 0.05 \). Being the insignificant amounts in the bulk-rock composition, MnO and \( \text{P}_2\text{O}_5 \) were ignored in constructing the pseudosection. It is a common phenomenon that dehydration melting of muscovite and biotite and \( \text{H}_2\text{O} \)-saturated melting result in partial melting, leading to melt loss from the metapelite protolith during metamorphism (White et al., 2001; Saha et al., 2008; Korhonen et al., 2011; Cutts et al., 2012), which caused the change of the bulk-rock composition after subsequent melt extraction (White et al., 2001; Saha et al., 2008; Korhonen et al., 2011; Cutts et al., 2012). Although it seems that the melt-integration during pseudosection calculation is more appropriate, it is hard to tell the exact amount of melt lost, and melt re-integration might not be a perfect solution to reveal the reality of the metamorphic processes even when leucosomes are found (Saha et al., 2008). In spite of the limitation, melt re-integration is often applied in constructing P–T pseudosections. In this modelling, \( \text{H}_2\text{O} \) content was adjusted to be 5.6 mol% which is just enough to saturate the mineral assemblages with consistent phase relations at supra- and subsolidus. As there is a great variety in the \( \text{FeO}/\text{Fe}_2\text{O}_3 \) ratio among metapelites, 12% of total \( \text{Fe}_2\text{O}_3 (\text{FeO}/\text{Fe}_2\text{O}_3) \) content was assumed to be \( \text{Fe}_2\text{O}_3 \), which is comparable to \( \text{FeO}/\text{Fe}_2\text{O}_3 \) and \( \text{FeO} \) contents of metapelites obtained by Wei et al. (2009). Quartz and ilmenite are set to be in excess when calculating the P–T pseudosection. With these assumptions, the bulk-rock composition was then corrected to the effective bulk-rock composition, in mol%, of \( \text{Na}_2\text{O} = 0.43, \text{CaO} = 1.17, \text{K}_2\text{O} = 2.19, \text{FeO} = 8.11, \text{MgO} = 7.97, \text{Al}_2\text{O}_3 = 15.24, \text{SiO}_2 = 63.75, \text{H}_2\text{O} = 5.6, \text{TiO}_2 = 0.58, \text{and} \ (\text{FeO}/\text{Fe}_2\text{O}_3) = 0.55 \) for pseudosection construction.

THERMOCALC 3.33 (Powell and Holland, 1998) and thermodynamic data set "tc-ds55.txt" updated in November 2003 (Holland and Powell, 1998) were applied in constructing the pseudosection. The minerals (with abbreviations) involved in the pseudosection construction include: garnet (g), sillimanite (sill), kyanite (ky), cordierite (cd), K-feldspar (ksp), plagioclase (pl), biotite (bi), muscovite (mu), paragonite (pa), quartz (q), rutile (ru), ilmenite (ilm), orthopyroxene (opx), silicate melt (liq) and fluid phase which is regarded as pure water phase (\( \text{H}_2\text{O} \)). The datafile used in the calculations was based on those from White et al. (2007), with updated activity–composition \( (a-x) \) relationships for garnet, biotite and silicate melt from White et al. (2007), cordierite from Holland and Powell (1998), plagioclase and K-feldspar from Holland and Powell (2003), muscovite and paragonite from Coggon and Holland (2002), ilmenite from White et al. (2002) and orthopyroxene from White et al. (2002).

5.2. **P–T pseudosections**

A specific P–T pseudosection in the NCKFMASHTO system for the chosen sample (JB04–5) is shown in Fig. 6. Within the P–T range of 3–11 kbar and 650–950 °C, key mineral assemblages involving melt are stable above 670 °C, with ilmenite appearing in all pseudosection fields (Fig. 6). Garnet is present in higher pressure fields with garnet-in line starting from 5.4 kbar/650 °C to 4 kbar/820 °C and to 5.4 kbar/950 °C (Fig. 6). In contrast, the cordierite–present region occurs in the lower pressure fields with pressures not exceeding 8.3 kbar (Fig. 6). Plagioclase is dominant in most of the fields except those with pressures above 7–8 kbar and temperature above ~900 °C, while biotite is also a common phase existing in fields below 920 °C at pressures over 8.2 kbar and below 850 °C at pressures under 6.4 kbar (Fig. 6). From 3 kbar to 6.4 kbar, sillimanite becomes less stable when temperature rises from 680 °C to 835 °C along the sillimanite–out line but it is stable above 6.4 kbar when temperature reaching 840 °C (Fig. 6). The kyanite–sillimanite conversion line starts from 6.6 kbar/650 °C to 11 kbar/850 °C, indicating that kyanite is more stable and replacing sillimanite at higher temperatures (Fig. 6). There are two feldspar–present zones, with one above 9.4 kbar between 840 and 930 °C and another below 7.6 kbar between 660 and 900 °C (Fig. 6). The presence of two feldspar–present zones within the P–T range of 3–11 kbar and 650–950 °C is not common but is thought to be due to the effect of melt-loss. In the transition zone from amphibolite-facies to granulite facies, the dehydration melting of hydrated minerals, such as muscovite and biotite, occurred during metamorphism, and potassium ions and \( \text{H}_2\text{O} \) from those minerals dissolved into the melt. Once the melt started losing from the rocks, the dissolved potassium was being extracted and its amount left in the rocks dropped. Continuous melt lost from the rocks led to insufficient potassium for generation of K-feldspar, resulting in the absence of K-feldspar in some pseudosection fields, especially between the pressure conditions of 7.6 kbar and 9.4 kbar. Quartz is no longer stable at low-pressure and high-temperature conditions starting from 820 °C but orthopyroxene exists at similar metamorphic conditions beginning at 810 °C (Fig. 6). Muscovite and paragonite are minor minerals in the pseudosection, only present in low-pressure–high-temperature areas below 10.7 kbar and 670 °C (Fig. 6).

5.2.1. **Pseudosection for the prograde (M1) assemblage**

As discussed above, the representative mineral assemblage of the prograde stage (M1) is garnet core (\( \text{grt}_c \)) + biotite (\( \text{bi}_c \)) + plagioclase (\( \text{pl}_c \)) + quartz (\( \text{q}_c \)) + ilmenite (\( \text{ilm}_c \)) ± muscovite (\( \text{mu}_c \)) ± stauriolite (\( \text{st}_c \)). However, this M1 mineral assemblage does not match with any pseudosection field in the pressure–temperature (P–T) range of 3–11 kbar and 650–950 °C; it is considered to have formed under much lower P–T conditions outside the defined P–T range. Due to strong reworking at the later metamorphic stages (M2–M4), it is difficult to acquire reliable chemical compositions of the minerals forming at the M1 stage, and thus the P–T conditions of the M1 stage cannot be quantitatively estimated.

5.2.2. **Pseudosection for the peak (M2) assemblage**

The peak (M2) mineral assemblage of garnet (\( \text{grt}_p \)) + sillimanite (\( \text{sill}_p \)) + K-feldspar (\( \text{Kfs}_p \)) + biotite (\( \text{bi}_p \)) + plagioclase (\( \text{pl}_p \)) + quartz (\( \text{q}_p \)) + ilmenite (\( \text{ilm}_p \)) matches with the pseudosection field of bi-g-ksp-pl-sill-liq-ilm-q, with pressure-dependent \( \text{X}_{\text{Kfs}_p} \) isopleths ranging from 0.046 to 0.076 (Fig. 6). By projecting \( \text{X}_{\text{Kfs}_p} \) of 0.48–0.50 from garnet mantle, P–T conditions for the peak (M2) metamorphism can be well constrained at 9.4–10 kbar and 870–900 °C (Fig. 6). From M1 to M2, muscovite (\( \text{mu}_c \)) and stauriolite (\( \text{st}_c \)) might have been completely consumed through
5.2.3. Pseudosection for the decompression (M3) assemblage

The decompression (M3) assemblage of symplectic cordierite (cd) + symplectic sillimanite (sill) + garnet rims (grt1) + matrix-type plagioclase rim (plr) + K-feldspar (Kfdm2) + quartz (q) + ilmenite (ilm) suits the cd-g-ksp-pl-sill-liq-ilm-q pseudosection field (Fig. 6). The $X_{Fe(g)}$ isopleth of cordierite in the field decreases from 0.32 to 0.19 with increasing pressure (Fig. 6). The $P$-$T$ conditions of the M3 assemblage can be constrained by $X_{Fe(g)}$ of the symplectic cordierite (0.290–0.331) at 6.3–6.6 kbar and 840–900 °C (Fig. 6). The metamorphic $P$-$T$ path from M2 to M3 passes through the K-feldspar-out, cordierite-in, K-feldspar-in and biotite-out lines (Fig. 6), of which the cordierite-in, K-feldspar-in and biotite-out are consistent with dehydration reaction (2) through which most biotite ($bi_{m}$) has been consumed to produce cordierite and K-feldspar (Kfdm2) during the decompression process.

5.2.4. Pseudosection for the decompression-cooling (M4) assemblage

The decompression-cooling (M4) assemblage of coronoic cordierite (cd) + garnet rim (grt2) + retrograded biotite ($bi_{m}$) + plagioclase rim (plr) + K-feldspar (Kfdm2) + quartz (q) + ilmenite (ilm) fits the bi-cd-g-ksp-pl-liq-ilm-q field (Fig. 6). Both the temperature-dependent $X_{Fe(g)}$ isopleths of garnet and $X_{Ca(pl)}$ isopleths of plagioclase are shown in the field, with $X_{Fe(g)}=0.71–0.8$ and $X_{Ca(pl)}=0.71–0.79$ (Fig. 6). The intersecting area of $X_{Fe(g)}$ values from the garnet rim (0.773–0.804) and $X_{Ca(pl)}$ values from the plagioclase rim (0.743–0.778) constrains the $P$-$T$ conditions of the M4 assemblage at 4–5.2 kbar and 815–830 °C (Fig. 6). During the decompression and cooling process from M3 to M4, the path left the sillimanite-free field and re-entered the biotite-present field (Fig. 6). This is in accordance with the instability of sillimanite that had been reacted during the
post-peak and retrograde metamorphism, and the formation of large retrograded biotite (bi_{m2}) via reaction (4).

5.3. P–T path in pseudosection

By combining the petrographic study and pseudosection modelling, we recognized four metamorphic stages in the medium-pressure pelitic granulites from the Jiaobei massif: the prograde (M1), peak (M2), decompression (M3) and decompression-cooling (M4) stages, of which the P–T conditions of the latter three metamorphic stages (M2–M4) have been well constrained. The peak assemblage (M2) is garnet (gr_{m2})+ sillimanite (sill_{m2})+ K-feldspar (K_{fddm2})+ biotite (bi_{m2})+ plagioclase (pl_{2})+ quartz (q_{m2})+ ilmenite (ilmm_{2}), with P–T conditions of 9.4–10 kbar and 870–900 °C. The decompression assemblage (M3) is symplectic cordierite (cd_{2})+ symplectite sillimanite (sill_{2})+ garnet rims (gr_{1})+ matrix-type plagioclase rim (pl_{1})+ K-feldspar (K_{fddm2})+ quartz (q_{m2})+ ilmenite (ilmm_{2}), with P–T conditions of 6.3–6.6 kbar and 840–900 °C. The decompression-cooling assemblage (M4) is corronic cordierite (cd_{4})+ biotite (bi_{m2})+ garnet rim (gr_{2})+ plagioclase rim (pl_{2})+ K-feldspar (K_{fddm2})+ quartz (q_{m2})+ ilmenite (ilmm_{2}), with P–T conditions of 4.5–5.2 kbar and 815–830 °C. Taken together, these mineral assemblages and their P–T conditions define a clockwise P–T path involving isothermal decompression (ITD) and decompression-cooling following the peak (M2) metamorphism (Fig. 6), suggesting that the Jiaobei massif underwent initial crustal thickening followed by rapid exhumation/uplift or rapid erosion, which is consistent with a continent–continent collision environment.

6. Discussion

As mentioned earlier, tectonic nature of the Jiao-Liao-Ji Belt has been controversial for almost two decades, which are focused on two contrasting models: the continent–arc–continent collision model (Bai, 1993; Bai and Dai, 1998; Faure et al., 2004) and the rifting- and closure-model (Zhang and Yang, 1988; Li et al., 2004a, 2004b, 2005, 2006; Li and Zhao, 2007).

The continent–arc–continent collision model was firstly proposed by Bai (1993) who suggested that the Longgang Block (Northern Liaoning–Southern Jilin complexes) and the Langrim Block (Southern Liaoning–Langrim complexes) belonged to two exotic Archean continental blocks, and the Jiao-Liao-Ji Belt represented an island arc and back basin in-between. Between Paleoproterozoic time, the two blocks were separated by an old ocean, and the oceanic lithosphere underwent a NW-directed subduction beneath the southern margin of the Longgang Block, leading to the development of an active-type continental margin on the southern margin of the Longgang Block, while the northern margin of the Langrim Block was a passive continental margin. In the late Paleoproterozoic, the subduction was completed and all the oceanic lithosphere was consumed, leading to the continent–arc–continent collision, forming the Paleoproterozoic Jiao-Liao-Ji intervening between the two Archean blocks (Bai, 1993; Bai and Dai, 1998). Later, Faure et al. (2004) modified the continent-arc collision model by interpreting the mafic–ultramafic rocks and marine sedimentary rocks of the North Liaoh Group as an active continental arc (magmatic belt) developing on a south-directed subduction zone lying between a northern Archean block (Longgang Bock) and a southern block that is largely composed of the South Liaoh Group. This magmatic belt was considered to have been overthrust upon the north Archean basement during the continent–arc–continent collision. Lu et al. (2006) also advocated the continent–arc–continent collision model but stressed differences between the northern and southern zones of the Jiao-Liao-Ji Belt. They argued that the northern zone, whose metamorphic evolution is characterized by clockwise P–T paths, represented a continent–arc–continent collisional belt, whereas the southern zone resulted from a post-orogenic thermal event, probably related to the rising of the asthenospheric mantle following the orogenic collapse. However, all these continent–arc–continent collision models are inconsistent with the absence of calc–alkaline igneous associations in the Jiao-Liao-Ji Belt, which commonly occur in modern magmatic arcs. In addition, Hf isotope analysis and U–Pb zircon age data suggest that the protoliths of the North and South Liaoh Groups should have been derived from the same sources, implying that they developed on the same Archean basement. Moreover, the metamorphism of the North and South Liaoh Groups occurred coevally at ~1.9 Ga. All these suggest that the above continent–arc–continent collision models cannot reasonably explain the tectonic nature and processes of the Paleoproterozoic Jiao-Liao-Ji Belt (Luo et al., 2004, 2008; Li et al., 2005).

The rifting- and closure-model was initially proposed by Zhang and Yang (1988) who suggested that the Longgang and Nangrim blocks originally belonged to a single continental block that underwent early Paleoproterozoic rifting, associated with the formation of the sedimentary–volcanic rocks and granitoid and mafic intrusions, and closed upon itself in the late Paleoproterozoic to form the Jiao-Liao-Ji Belt (Zhang and Yang, 1988; Li et al., 2004a, 2005, 2006; Luo et al., 2004, 2008; Li and Zhao, 2007). This model can reasonably explain (1) the presence of large volumes of A-type granites in the Jiao-Liao-Ji Belt (Lu et al., 2006); (2) the presence of bimodal volcanic assemblages in the belt (Zhang and Yang, 1988; Peng and Palmer, 1995); (3) geochemically and geochronologically similar late Archean TTG basement gneisses and mafic dyke swarms on the opposite sides of the Jiao-Liao-Ji Belt (Zhang and Yang, 1988); and (4) anticyclonic P–T paths of the Jiao-Liao-Ji Belt (Luo et al., 1996; He and Ye, 1998), which are not consistent with a continent–continent collision model; and (5) the presence of borate deposits of a non-marine origin in the Jiao-Liao-Ji Belt, which have many similarities to those borate-bearing successions in other Proterozoic rifting environments, e.g., the Upper Proterozoic Damaran Orogen of South Africa (Peng and Palmer, 1995). However, this model cannot reasonably explain the polyphase compressive deformation and clockwise P–T paths of the Laolong, North Liaoh and Fenzishan Groups. In particular, the rifting- and closure-model cannot well explain the presence of high- and medium-pressure pelitic granulites in the southern segment of the Jiao-Liao-Ji Belt (Zhou et al., 2004, 2008a; Li et al., 2012; Tam et al., 2011, 2012).

More importantly, the rift- and closure-model cannot well explain the presence of high-pressure pelitic granulites in the southern segment of the Jiao-Liao-Ji Belt reported by Tam et al. (2012), because if subduction and/or continent–continent collision processes were not involved, it is hard to imagine how the sedimentary protoliths of the high-pressure pelitic granulites were brought down to a lower crustal level where they experienced high-pressure granulite-facies metamorphism. Tam et al. (2012) recognized three metamorphic mineral assemblages from the high-pressure pelitic granulites from the Jiaobei massif: the early prograde assemblage (M1) represented by inclusion-type minerals (biotite + kyanite + muscovite + plagioclase + quartz + ilmenite) within the cores of garnet; the peak high-pressure granulite-facies assemblage (M2) of garnet (mantle) + K-feldspar + kyanite + plagioclase + biotite + rutile + ilmenite + quartz; and the decompression assemblage (M3) represented by garnet (rim) + sillimanite + plagioclase + biotite + ilmenite + quartz. Applying the pseudosection modelling in the NCKFMASHITO system, Tam et al. (2012) constrained the P–T conditions of the M1, M2 and M3 assemblages at 9.3–10.7 kbar/645–670 °C, 14.8–16.2 kbar/860–890 °C, and 6.3–8.5 kbar/710–740 °C, respectively, which define a clockwise P–T path involving decompression...
and cooling following the peak high-pressure granulite facies metamorphism. This suggests that the high-pressure pelitic granulites from the Jiaohe massif experienced the initial crustal thickening (M1 and M2), followed by exhumation and cooling (M3), which is consistent with a continent–continent collision environment. Thus, the southern segment of the JLJB must have been involved in subduction- or collision-related tectonic processes.

In this study, we show that the medium-pressure pelitic granulites from the Jiaohe massif underwent tectonothermal processes similar to those of the high-pressure pelitic granulites, characterized by near-isothermal decompression and decompression-cooling following the peak granulite facies metamorphism. The peak medium-pressure granulite-facies metamorphism is represented by the mineral assemblage of garnet + sillimanite + K-feldspar + biotite + plagioclase + quartz + ilmenite. The near-isothermal decompression metamorphism is marked by the formation of the cordierite + sillimanite symplectites replacing the matrix-type biotite and quartz, and the decompression-cooling metamorphism is indicated by the cordierite coronas surrounding the garnet grains and retrograded biotite replacing the garnet. Mineral reaction relations, mineral chemistry and pseudosection modelling in the NCKFMASHO system constrain the P–T conditions of the peak (M2), decompression (M3) and decompression-cooling (M4) assemblages at 9.4–10 kbar and 870–900 °C, 6.3–6.6 kbar and 840–900 °C, and 4–5.2 kbar and 815–830 °C, respectively, which define a clockwise P–T path involving near-isothermal decompression and cooling for the medium-pressure pelitic granulites from the Jiaohe massif. Such a P–T path suggests that the high-pressure pelitic granulites, the widely exposed medium-pressure pelitic granulites in the Jiaohe massif also underwent initial crustal thickening followed by rapid exhumation/uplift or rapid erosion. This indicates that the development of the whole Jiaohe massif in the southern segment of the Jiao-Liao-Ji Belt must have been involved in subduction- or collision-related tectonic processes, not formed by a simple closure of a rift system. Therefore, like the Paleoproterozoic Khondalite Belt in the Western Block and the Trans-North China Orogen in the central part of the NCC, the Jiao-Liao-Ji Belt may represent another Paleoproterozoic collisional belt along which the Longgang and Langrim Blocks amalgamated to form the Eastern Block. It deserves mentioning here that though our data presented in this study are consistent with a continent–continent collisional model, we cannot preclude the possibility that the Jiao-Liao-Ji Belt was initialized from an intracratonic rift basin as proposed by many researchers (Zhang and Yang, 1988; Peng and Palmer, 1995; Li et al., 2004a, 2005, 2006; Luo et al., 2004, 2008; Li and Zhao, 2007; Li et al., 2012). What we can conclude from this study and previous data is that even though the Jiao-Liao-Ji Belt initially developed from a rift basin, this basin must have developed into an ocean basin at least in its southern segment in the period 2.2–1.9 Ga (e.g. Luo et al., 2004, 2008; Lu et al., 2006; Li and Zhao, 2007), where the oceanic lithosphere was subducted, leading to the final closure of the ocean basin with the formation of the high- and medium-pressure pelitic granulites at ~1.9 Ga. Further petalogical, structural, geochemical and geochronological investigations are needed to establish the detailed tectonic architectures and evolution of the Jiao-Liao-Ji Belt.

7. Conclusions

1. The medium-pressure pelitic granulites from the Jiaohe massif in the southern segment of the JLJB experienced four distinct metamorphic stages (M1–M4). The prograde stage (M1) is represented by mineral inclusions within the core of garnet grains, and at the peak stage (M2), the garnet mantle and matrix-type sillimanite + K-feldspar + biotite + plagioclase + quartz + ilmenite.

2. A combination of chemical compositions (isopleths) of key minerals (e.g. garnet, plagioclase and cordierite) and the pseudosection modelling in the NCKFMASHO system defines P–T conditions of 9.4–10 kbar and 870–900 °C for the peak (M2) metamorphic stage, 6.3–6.6 kbar and 840–900 °C for the post-peak (M3) metamorphic stage and 4–5.2 kbar and 815–830 °C for the retrograde (M4) metamorphic stage.

3. A clockwise P–T path involving isothermal decompression (ITD) and subsequent decompression and cooling is reconstructed for the medium-pressure pelitic granulites. This implies that the southern segment of the JLJB must have undergone subduction and/or collision-related tectonic processes. Therefore, the JLJB may represent another Paleoproterozoic collisional belt along which the Longgang and Langrim Blocks amalgamated to form the Eastern Block.

Acknowledgements

This study was financially supported by the Hong Kong RGC GRF project (HKU7069/12P and HKU7057/08P) and the Chinese NSFC Grants (40730315 and 40872123). We thank Prof. M. Santosh and an anonymous reviewer for their thoughtful comments that are helpful in improving the quality of the paper. Special thanks are given to Prof. Chunjing Wei for great comments on constructing pseudosections, Dr. Qian Mao and Dr. Yuguang Ma for helping the EPMA probing, and Dr. Jean Wong and Mr. Jianfeng Gao for constructive and helpful discussions.

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