Metamorphic evolution and Zircon ages of pelitic granulites in eastern Hebei, North China Craton: Insights into the regional Archean P–T–t history

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ABSTRACT

Archean supracrustal granulites occur as localized slivers or rafts in granitoids, and this particular occurrence of supracrustal granulites has not been much concerned from the perspective of metamorphic evolution. The pelitic granulites in eastern Hebei occur as slivers in TTG gneisses and charnockite. Three pelitic granulite samples

involving orthopyroxene–garnet gneiss (JD1334), biotite–garnet gneiss (JD1453) and

sillimanite–garnet gneiss (JD15126) were collected from Taipingzhai and Laolijia

areas. Based on textural observations and mineral compositions, four generations of

assemblages involving final, peak, inclusion and overprinting stages are identified.

The observed final assemblage yield P–T conditions of 8.0–9.5 kbar/870–890 °C. The

observed peak assemblage characteristic of the absence of the matrix biotite record

condition of T > 950 °C with fixed P = 9 kbar. An isobaric cooling path from peak to

final assemblage is constrained by mineral isopleths. The inclusion assemblages

characterized by plagioclase inclusions with high X_{An} contents are predicted to be

stable at low pressure and/or high temperature field. They form an anticlockwise P–T

path. The overprinting assemblage characteristic of coronary garnet (g_2) coexisting

with fine-grained mineral aggregates defines P–T condition of 10–12

kbar/830–880 °C. The deposition age of the sedimentary protolith of pelitic granulites

was ~2.5 Ga, coeval with the crystallization age of TTG gneisses. Both of them were

later subjected to granulite facies metamorphism at ~2.48 Ga. The Archean

anticlockwise P–T paths suggest a tectonic scenario to explain the particular

occurrence of the Archean supracrustal granulites. Firstly, supracrustal sediments

experienced thermal event under low pressures attributed to the intrusion of

high-temperature TTGs, resulting in dehydrations of hydrous phases. Then, the dried

metasediments interlayered with iron-bearing formations potentially denser than the

surrounding lithologies tend to drop into the TTG magma ocean, causing the

pressure-increase in the metasediments. Thirdly, the dropped metasediments

experienced cooling process under lower crust in coupled with the crystallization of the surrounding TTG magma to form the observed final assemblages. Finally, they uplifted as a consequence of the doming processes in Archean terranes.

1. Introduction

The Archean terranes are dominantly composed of granitoids of tonalite–trondhjemite–granodiorite (TTG) compositions and supracrustal rocks, and are divided into low- and high-grade domains according to metamorphic grades (Condie, 1981). The low-grade or granitoid-greenstone terrains are prevailed with the large-scale development of ''dome-and-keel'' structures where the greenstone belts occur as synclinal keels among granitoid domes (e.g. Kröner, 1984; Campbell and Hill, 1988; Collins et al., 1998; Bleeker, 2002; Lin and Beakhouse, 2013). The metamorphic grade of Archean greenstone rocks ranges from greenschist- to amphibolite-facies, and higher grades typically occur around the margins of greenstone belts, probably caused by intrusion of plutons (Boniface and Mruma, 2012). In contrast, the high-grade terrains are characteristic of the development of granulites. Brown (2007) summarized the peak P–T conditions of granulites from twenty Archean terranes, where nine of them fall into ultrahigh-temperature (UHT) type with P–T conditions of 9–14 kbar/950–1100 °C, eight of them into normal

granulite type with P–T conditions of 6.5–9.9 kbar/880–950 °C and three of them into

medium-temperature eclogite–high-pressure granulite type with P–T conditions of

11–17.5 kbar/700–880 °C. It seems that the Archean granulites are dominated by

UHT type. In some Archean high-grade terrains, supracrustal granulites occur as

slivers or rafts of tens to thousands square meters in granitoids, for instance, the

supracrustal rocks from the Akia and the Tasiusarsuaq terranes in southern West

Greenland (Riciputi et al., 1990), the Vedlozero–Segozero high-grade greenstone belts

in Karelian Province (e.g. Käyaho et al., 2007), the Sargur Group in Western Dharwar

Craton (e.g. Peucat et al., 1993; Ghosh et al., 2004) and the supracrustal granulites of

North China Craton (NCC) exposed in eastern Hebei (Wu et al., 1998). This particular

occurrence of the Archean supracrustal granulites has not been too much concerned

especially from a point of view of metamorphic evolution.

The temperature conditions of granulites are commonly underestimated using

conventional Fe–Mg exchange thermometers due to the fast cation diffusion during

cooling (Frost and Chacko, 1989; Harley, 1989; Powell and Holland, 2008). An

alternative approach using the Al-solubility in orthopyroxene in equilibrium with

garnet proposed by Pattison et al. (2003) can correct the effect of late Fe–Mg

exchange to some extent. However, it cannot be widely applied to rocks without the

specific mineral assemblages. Recently, pseudosection thermobarometry is

increasingly used to determine the P–T conditions of granulites, which typically

yields higher peak temperature estimates because it uses the assemblage as the basis

for estimating P–T conditions and the assemblage is less prone to post-peak

modification than mineral compositions (e.g. White et al., 2002; Kelsey et al., 2003; Korhonen et al., 2013). However, as granulite metamorphism is commonly concomitant with melt, the post-peak metamorphic (or retrograde) reactions with melt may modify the mineral modes and compositions and even cause the scenario that the preserved 'peak' assemblages in granulites commonly represent the retrograde solidus conditions, rather than the real peak conditions that the rock has reached (White and Powell, 2002; Korhonen et al., 2013; Li and Wei, 2016). The assemblages on the retrograde fluid-absent solidus in granulites are nominated as final assemblages, which can be differentiated from the peak assemblages through detailed petrographic observations (Korhonen et al., 2013). Generally, some components, such as, anorthite of plagioclase, grossular of garnet and Ti in biotite, that are less affected by diffusion are significant for constraining the peak P–T conditions of granulites (Li and Wei, 2016).

The Eastern Hebei terrane is located at the northwestern part of the NCC Eastern Block and comprises predominantly Neoarchean metamorphic basement, which are overlain at their periphery by Mesoproterozoic to Mesozoic cover sequences, and are intruded by several Mesozoic granitic plutons (e.g. Yang et al., 2008; Li et al., 2010; Nutman et al., 2011; Zhao and Zhai, 2013). The Neoarchean basement is composed of TTG gneisses, granitoid plutons and supracrustal rocks (Fig. 1a), and the intrusive rocks of 2.6–2.5 Ga cover 80–70% of the surface (e.g. Geng et al., 2006, 2012; Yang et al., 2008; Li et al., 2010; Nutman et al., 2011; Zhao and Zhai, 2013; Bai et al., 2014, 2015). In Taipingzhai region, the TTG gneisses and supracrustal rocks were

considered to have been subjected to granulite-facies metamorphism in Neoarchean

time with an anticlockwise cooling P–T path (Zhao et al., 1998). Using conventional

thermobarometers, the peak P–T conditions for mafic granulites in the supracrustal

sequences were estimated to be 9–11 kbar/810–940 °C (He and Ye, 1992), 9.5

kbar/780–816 °C (Chen and Li, 1996) and 8–8.5 kbar/850–900 °C (Zhao et al., 1998).

Recently, the mafic granulites and charnockite at Taipingzhai were estimated using

pseudosection modelling with THERMOCALC, giving 9.6–10.3 kbar/860–900 °C

(Kwan et al., 2016) and 7–14 kbar/800–920 °C (Yang et al., 2015) for the peak

conditions, similar to the results from the conventional thermometers. However, there

have been no data available for pelitic granulites in eastern Hebei.

In this paper, we present systematic data of petrography, major element

compositions, LA-ICP-MS zircon U–Pb ages and phase modelling results for pelitic

granulites from eastern Hebei to elucidate their metamorphic evolution and geological

significance.

2. Geological setting

The North China Craton is one of the oldest cratonic blocks on Earth, containing

crustal rocks up to ~3.85 Ga (Liu et al., 1992; Song et al., 1996; Liu et al., 2007,

2008). The Precambrian basement of the NCC can be divided into high-grade gneiss

complexes and low- to medium-grade granite-greenstone belts that are widespread

over the whole NCC (Zhao and Zhai, 2013). For the past decade, it has been widely

accepted that the NCC is subdivided into Western Block and Eastern Block that are

welded along the Trans-North China Orogen (TNCO), although the timing and

tectonic processes involved in the amalgamation of the two blocks remains

controversial (e.g. Zhao et al., 1998, 2001, 2005, 2012; Zhao and Zhai, 2013; Santosh

et al., 2012; Liu et al., 2012; Wilde et al., 2002, 2005; Wei et al., 2014).

   The early Precambrian Eastern Hebei terrane can be divided into five litho-tectonic

units after Wu et al. (1998): (I) the Zunhua high-grade meta-greenstone belt, (II)

Taipingzhai granulite facies unit, (III) Qian'an gneiss dome region, (IV) Luanxian

amphibolite-facies and (V) Qinglong amphibolite/greenschist-facies metamorphic

belts (Fig. 1b). The Zunhua high-grade meta-greenstone belt is composed mainly of

supracrustal rocks and orthogneiss (Wu et al., 1998; Geng, 1998; Geng et al., 2006;

Zhai and Liu, 2001; Nuttman et al., 2011, 2014; Guo et al., 2013, 2015; Liu et al.,

2014; Bai et al., 2015). These supracrustal rocks were metamorphosed under

granulite-facies conditions and comprised metasedimentary rocks, Banded Iron

Formations (BIF) and metavolcanic rocks that were dated at 2.61–2.52 Ga (Guo et al.,

2013). Wall rocks from the Shirengou BIF deposit, comprising hornblende plagioclase

gneisses and plagioclase amphibolites, have magmatic crystallization ages of

2.55–2.54 Ga (Zhang et al., 2012). Recent investigations based on progressive zircon

U–Pb dating methods reveal that most of these orthogneisses series have Neoarchean

magmatic precursors that were dated at 2.55–2.50 Ga (e.g. Geng et al., 2006; Nuttman

et al., 2011; Guo et al., 2013; Bai et al., 2014, 2015). The Taipingzhai granulite-facies

unit consists of TTG gneisses, charnockite and supracrustal rocks (Wu et al., 1998).

The magmatic precursors of the TTG gneisses and charnockite were emplaced

between 2.53–2.49 Ga and 2.55–2.52 Ga, respectively, which were followed by

regional granulite-facies metamorphism at 2.45–2.42 Ga (Geng et al., 2006; Bai et al.,

2015). The supracrustal rocks including mafic, pelitic granulites and magnetite

quartzites occur as slivers of hundreds to thousands of meters in the extensive TTG

gneisses (Figs. 2 and 3a). The Qian'an gneiss dome, which is surrounded by

supracrustal residues of Paleo-Mesoarchean in its southwest border, is composed

mainly of tonalitic-granodioritic gneisses of amphibole to granulite facies

metamorphism. Detrital zircon grains obtained from fuchsite-bearing quartzite in

these supracrustal residues show ages > 3.85 Ga (Liu et al., 1992; Wilde et al., 2008;

Nuttman et al., 2011, 2014). Most orthogneisses that were metamorphosed in

amphibole/granulite-facies were dated 2.55–2.51 Ga, but a few emplaced at 3.28–2.94

Ga (Nuttman et al., 2011, 2014). The Luanxian amphibolite-facies belt is composed of

a volcano-sedimentary series that was grouped as the Luanxian Group (Qian et al.,

1985), and a number of tonalitic-granodioritic gneisses that were dated with an

emplacement age of 2.55–2.53 Ga (Geng et al., 2006). The Qinglong

amphibolite/greenschist-facies belt consists of turbiditic sediments interlayered with

basic and acidic volcanics grouped as the Shuangshanzi Group (amphibolite-facies)

and Zhuzhangzi Group (greenschist-facies). Some researchers interpreted the

Shuangshanzi and Zhuzhangzi Groups as Paleoproterozoic products in the previous

studies (e.g. Sun, 1984; Qian et al., 1985). In addition, Liu et al. (2014) reported that

the sedimentation of the Shuangshanzi Group began at ~2.35 Ga which were

evidenced by the crystallization age of 2352 ± 4 Ma of a meta-porphyritic dacite and

the youngest detrital zircon age peaks of 2337–2329 Ma from the Shuangshanzi

Group, and the youngest age peaks of 2331–2329 Ma have also been identified from

the overlying Zhuzhangzi Group. Alternatively, both Shuangshanzi and Zhuzhangzi

Groups were proven to be actually deposited at the end of the Neoarchean, based on

zircon ages of 2.56–2.50 Ga from the low-grade metasedimentary rocks (Sun et al.,

2010; Lü et al., 2012; Guo et al., 2015), and represent sedimentary covers in an

intra-continental rift basin after the ~2.5 Ga cratonization (e.g. Zhai, 2011; Lü et al.,

2012). A few mafic dykes were described cross-cutting the Archean TTG gneisses and

supracrustal rocks in the Zunhua meta-greenstone belt and the Taipingzhai complex

(Chen, 1990; Song, 1990), which were determined to have experienced high pressure

granulite-facies metamorphism in ~1.82 Ga (Duan et al., 2015).

The studied pelitic granulites from the Taipingzhai granulite-facies unit are

interlayered with magnetite quartzite and two-pyroxene granulites (Fig. 3a). They are

intensively partial melted with leucosomes of several millimeters wide (Fig. 3b).

3. Petrological analyses

3.1 Analysis method

The bulk-rock compositions of the pelitic granulites were determined by

ICP-OES analysis at China University of Geoscience (Beijing). Microprobe analyses

of minerals were performed using a JXA-8100 microprobe at the Laboratory of

Orogenic Belt and Crustal Evolution of Peking University. The analyses were

conducted under conditions of 15 kV accelerating voltage, a 10-nA current with a

beam diameter of 2 μm. Natural and synthetic minerals of SPI Company were used

for standardization. Compositional X-ray maps of garnet were obtained with an

energy dispersive X-ray spectrometry of Oxford INCA X-MAX50 250 +, installed on

a scanning electron microscope (SEM) of FEI Quanta 650 FEG. The running

condition was set to a 20 kV acceleration voltage and spot 5 beams current.

3.2 Bulk-rock compositions

Three samples (JD1334, JD1453 and JD15126) were selected from the

Taipingzhai and Louzishan areas (Fig. 2). Their locations, main mineral components

and modal proportions are listed in Table 1, and bulk-rock compositions are presented

in Table 2. They show SiO_2 = 58.32–62.74 (wt.%), Al_2O_3 = 16.60–17.41, CaO =

2.20–3.46, Na_2O= 2.23–3.68, K_2O= 1.65–2.19, A/CNK = 1.23–1.70, and $Mg^{\#}$= 30–52,

indicating CaO-rich and K_2O-poor pelitic compositions.

3.3 Petrography and mineral chemistry

Pelitic granulites are porphyroblastic with garnet or occasionally with plagioclase

porphyroblasts in a foliated matrix consisting of plagioclase, K-feldspar, quartz and

minor biotite and ilmenite with or without orthopyroxene and sillimanite.

Representative mineral analyses are listed in Tables 3 and 4. Field and micrographs of

representative pelitic granulites are shown in Figs. 3–4 and representative mineral

chemistry is presented in Fig. 5.

Sample JD1334 is an orthopyroxene-bearing pelitic granulite in which garnet can

be divided into two types. Type-1 garnet (g_1) occurs as subhedral to anhedral

porphyroblasts of 0.5–2 mm across and contains inclusions of quartz and ilmenite,

whereas type-2 garnet (g_2) occurs as coronae around porphyroblastic garnet (Fig. 3c).

Type-1 garnet contains 67–68 mol.% almandine, 25–27 mol.% pyrope, 5–7 mol.%

grossular and ~1 mol.% spessartine, and shows weak zoning with core-rim increasing

in grossular from 5 to 7 mol.% (Fig. 5a). Type-2 garnet shows higher grossular of

10–13 mol. % than type-1 (Fig. 5a and Table 3). Plagioclase is divided into two types

on the basis of grain sizes. The coarse-grained plagioclase is anhedral and 0.2–0.5 mm

across in the matrix (Fig. 3d), showing core-rim decreasing in X_{An} [=Ca/(Ca + Na +

K)] from 0.31 to 0.26 (Fig. 5f). The fine-grained plagioclase is of 0.02–0.06 mm

across with little lower X_{An} of 0.24–0.29, and forms aggregates with quartz, biotite

and K-feldspar (Fig. 3e). Biotite occurs as medium-grained flakes of 0.1–0.3 mm long

in the matrix with X_{Mg} [= Mg/(Mg + Fe^{2+})] of 0.56–0.59 and Ti = 0.29–0.32 p.f.u.

These biotite flakes show irregular shapes formed along grain boundaries, or as

pressure shadows of garnet (Fig. 3e), indicating of later generation. Orthopyroxene

occurs as anhedral grains of 0.1–0.5 mm across and is fairly homogeneous in

chemical composition with X_{Mg} [= Mg/(Mg + Fe^{2+})] of 0.52–0.53. K-feldspar occurs

as medium-grained tabular crystals in the matrix, with Ab = 0.09–0.11 and Or =

0.89–0.91. Based on the textural observation and mineral compositions presented

above, three generations of assemblages involving final, peak and overprinting of

metamorphism can be identified. The final assemblage mainly consists of type-1

garnet, orthopyroxene, K-feldspar, plagioclase (coarse-grained), biotite, quartz and

ilmenite, and the peak assemblage is characteristic of the absence of the matrix biotite.

The overprinting assemblage consists of type-2 garnet, aggregates of quartz, biotite,

K-feldspar and plagioclase (fine-grained).

Sample JD1453 is a pelitic granulite with higher amounts of plagioclase and biotite than in the other samples. Garnet occurs as porphyroblasts of 1–5mm across. It contains X_{alm} = 0.52–0.57, X_{py} = 0.37–0.41, X_{gr}= 0.04–0.08 and X_{sps}= 0.01. A few large garnet grains of 4–5mm across contain numerous inclusions of plagioclase and a few grains of quartz (Fig. 4a) and show higher X_{gr} of 0.07–0.08 in the core and mantle than in the rim (Fig. 5b). However, most medium-coarse garnet grains of 1–3 mm across contain a few inclusions of quartz (Fig. 4b) and show lower X_{gr} of 0.04 without any change from core to rim (Fig. 5c). Garnet commonly develops an outer rim that is higher in X_{gr} and lower in X_{py} than in the inner rim (Fig. 5c), similar to the overprinting garnet (g_2) in sample JD1334. Biotite occurs as small flakes of irregular shapes in the matrix, and commonly forms in the pressure shadows of garnet, showing X_{Mg} = 0.62–0.64 and Ti = 0.31–0.33. A few biotite flakes are formed along the cracks in garnet (Fig. 4b) and shows higher X_{Mg} of 0.73 and lower Ti = 0.28 p.f.u. than in the matrix grains, interpreted to be formed under low temperature with fluid infiltration. Plagioclase occurs as anhedral grains of 0.3–3 mm across in the matrix (Fig. 4e) or as rounded inclusions of 0.1–0.4 mm across in garnet (Fig. 4a). Plagioclase in the matrix has X_{An} of 0.29–0.31 without clear zoning (Fig. 5g1), whereas plagioclase inclusions in garnet have higher X_{An} of 0.43–0.45 (Fig. 5g2). K-feldspar occurs as medium-grained anhedral grains in the matrix, with Ab = 0.05–0.07 and Or = 0.93–0.95. Based on the textural relations and mineral compositions presented above, four generations of assemblages involving final, peak, inclusion and overprinting of

metamorphism can be identified. The peak assemblage is observed to include garnet

(g_1), K-feldspar, plagioclase, quartz and ilmenite, the final assemblage is

characteristic of the formation of the matrix biotite together with other minerals

observed in the peak assemblage. The inclusion and overprinting assemblages are

characterized by plagioclase with high X_{An} (0.43–0.45) and garnet with high grossular

content (g_2), respectively.

   Sample JD15126 is a sillimanite-bearing pelitic granulite with good foliation

defined by oriented sillimanite, biotite and plagioclase. Garnet occurs as anhedral or

subhedral porphyroblasts of 0.5–2 mm across and contains X_{alm} = 0.56–0.58, X_{py} =

0.37–0.41, X_{gr}= 0.03–0.05 and X_{sps}= 0.00–0.01. Garnet has inclusions of plagioclase

and shows little higher X_{gr} in the core than garnet without plagioclase inclusions (Figs.

5d, e). Both of them show flat profiles of X_{gr} in the core but decreasing in the narrow

rim section (g_1) and are locally surrounded by Ca-rich garnet corona (g_2, Figs. 4d, 5d,

e). Biotite occurs as small fibers of irregular shapes in the matrix, and commonly

forms in the pressure shadows of garnet, showing X_{Mg} = 0.75 and Ti = 0.27–0.30.

Plagioclase occurs as anhedral grains of 0.3–0.8 mm across in the matrix (Fig. 4f) or

as rounded inclusions of 0.1–0.3 mm across in garnet (Fig. 4d). Most plagioclase

grains in the matrix are not significantly zoned relative to X_{An}, and a few grains show

core-rim zoning with X_{An} decreasing from 0.34 to 0.30 in the matrix (Fig. 5h1). The

plagioclase inclusions in garnet show core-rim zoning with X_{An} decreasing 0.49–0.47

(Fig. 5h2). Sillimanite exhibits oriented columnar and granular crystals in the matrix.

K-feldspar occurs as tabular crystals in the matrix, with Ab = 0.09–0.10 and Or =

0.89–0.91. Based on the textural observation and mineral compositions presented above, the observed peak assemblage contains garnet (g_1), K-feldspar, sillimanite, plagioclase, quartz and rutile, the final assemblage is characteristic of the formation of the matrix biotite together with the other minerals observed in the peak assemblage. The inclusion and overprinting assemblages are characterized by plagioclase with high X_{An} (0.47–0.49) values and Ca-rich garnet (g_2) than inner rim, respectively, same as sample JD1453.

In addition, the overprinting of garnet in some samples presents fantastic coronae around the early-stage garnet porphyroblasts (Fig. 3f).

4. Phase equilibria modelling using pseudosections

Based on the mineral assemblages and compositions for samples JD1334, JD1453 and JD15126 mentioned above, the model system (Mn)NCKFMASHTO ($MnO–Na_2O–CaO–K_2O–FeO–MgO–Al_2O_3–SiO_2–H_2O–TiO_2–Fe_2O_3$) was chosen to calculate P–T and T–M(H_2O) pseudosections. This system provides the most realistic compositional approximation to the studied rocks within which suprasolidus conditions can be calculated (White et al., 2007). Pseudosection calculations were performed using THERMOCALC 3.33, using the October 2009 updated version of the Holland and Powell (1998) dataset (files tcds55.txt). Mineral activity–composition relationships are those defined for silicate melt (White et al., 2007), biotite (White et al., 2007), cordierite (Holland and Powell, 1998), garnet (White et al., 2005, 2007), orthopyroxene (White et al., 2002), plagioclase and K-feldspar (Holland and Powell,

2003) and ilmenite (White et al., 2000). Kyanite/sillimanite, rutile and quartz are pure end-member phases.

The whole-rock chemical compositions obtained by ICP-OES analysis for a piece of rock generally cannot be used to model well the phase equilibria observed in a thin-section because of (1) fractionation consequent to the growth of porphyroblasts, commonly evident as zoned minerals, (2) the heterogeneous accumulation of some minerals (e.g. mica-rich folia), and (3) the effect of retrograde metamorphism. The effective bulk-rock compositions would be better when modelling phase diagrams than the whole-rock chemical compositions. The effective bulk-rock compositions were generated according to mass balance constraints by integrating mineral compositions and modal abundance data for the phases present (Carson et al., 1999). The modal abundance for each mineral was obtained by point-counting. In addition, the H_2O content used in the modelling is adjusted using T–M(H_2O) (referring to the H_2O mol.% in a rock) diagrams so that the final phase assemblage is stable just above the solidus (Korhonen et al., 2013). The effective bulk-rock compositions for the calculated pseudosections are listed in Table 2.

Sample JD1334

A P–T pseudosection is calculated using an effective bulk-rock composition on the basis of the final assemblage for K-feldspar, plagioclase, quartz and ilmenite in excess with the P–T window of 2–13 kbar and 700–1100 °C (Fig. 6). The solidus is predicted at temperatures between 760 and 870 °C. Biotite-out occurs at temperatures

just above the solidus. Cordierite is stable at pressures less than ~5 kbar.

The observed final assemblage g–bi–opx–ksp–pl–q–ilm(–liq) is quadrivariant in the model system and defines a narrow temperature range of 865–870 °C under a fixed pressure of 9 kbar. The measured X_{py} = 0.26–0.27 in the core and mantle of garnet (g_1) and the X_{Mg-bi} = 0.59 in the matrix biotite are plotted in the stability field for the final assemblage, but the measured X_{py} of 0.25 in the rims of g_1 is plotted in the subsolidus fields. The measured X_{Mg-opx} = 0.52–0.53 is predicted to mostly reflect cooling conditions as the X_{Mg-opx} increases with cooling in the subsolidus fields. The observed biotite-absent peak assemblage g–opx–ksp–pl–q–ilm(–liq) is quinivariant in the model system and defines a wide P–T range of 5.0–12 kbar and 825–1090 °C. The measured core-rim zoning of increasing X_{gr} (0.05→0.07) and decreasing X_{An} (0.31→0.26) are modelled to be consistent with a cooling path, where the isopleths of the maximum X_{An} and minimum X_{gr} from grains cores define a P–T condition of ~1050 °C and 9.8 kbar. The P–T estimate has uncertainties of ~0.4 kbar and ~24 °C (two-sigma level) calculated with THERMOCALC for the X_{An} and X_{gr} isopleths, respectively. However, such uncertainties should be considered a minimum as they are propagated from the uncertainties on the enthalpy alone and do not include other sources of uncertainty. The predicted Ti content in biotite is only 0.18 p.f.u. at temperatures on the biotite-out, significantly lower than the measured values (Ti = 0.26–0.32 p.f.u.). From the above-mentioned, an isobaric cooling path can be inferred as shown in Fig. 6. On the basis of the changes of phase modes and compositions calculated (Table 5), metamorphic evolution is dominated by melt crystallization via

liq = g + ksp + pl + opx + q + ilm in the P–T field of the peak assemblage, followed

by ksp + opx + ilm + liq = g + q + pl + bi in the narrow field of the final assemblage.

It is noted that the observed mineral modes are roughly consistent with the predicted

values at point B on the solidus. Further cooling under the subsolidus is predicted by

exchange reactions causing Mg-rich in biotite and orthopyroxene and Fe-rich in

garnet.

For modelling the P–T evolution of the overprinting assemblage, an effective

bulk rock composition was generated according to the equilibria in local domains by

assuming type-2 garnet (g_2) to be ~10 vol. % and the modal abundance for other

minerals to be their original proportions in the final assemblage. Its effective

bulk-rock composition is listed in Table 2. A P–T pseudosection is calculated for

K-feldspar, plagioclase, quartz and ilmenite in excess (Fig. 7). The observed

overprinting assemblage g–bi–ksp–pl–q–ilm(–liq) defines a P–T range of 10–12 kbar

and 830–880 °C, consistent with the predictions from the measured X_{An} = 0.24–0.29

in the fine-grained plagioclase and the type-2 garnet compositions with higher X_{py} =

0.22–0.23 and X_{gr} = 0.12–0.13

Sample JD1453

The P–T pseudosection calculated for sample JD1453 is drawn for plagioclase

and quartz in excess with the P–T window of 2–12 kbar and 750–1150 °C (Fig. 8).

The fluid-absent solidus occurs at temperatures about 850–870 °C under pressure of

5–10 kbar. The observed final assemblage g–bi–ksp–pl–q–ilm(–liq) is quadrivariant in

the model system and defines a P–T range of 7.0–9.5 kbar and 860–900 ℃. The measured X_{py} = 0.38–0.41 and X_{gr} = 0.04–0.08 in garnet and X_{An} = 0.29–0.31 in the matrix plagioclase are predicted to yield P–T conditions roughly consistent with the stability of the final assemblage, but the plagioclase compositions suggest higher temperature in orthopyroxene-stability fields. The measured X_{Mg-bi} of 0.62–0.64 from the matrix biotite could reflect lower pressure and temperature cooling conditions, while the predicted Ti content in biotite is only 0.22 p.f.u. at the biotite-out temperature, significantly lower than the measured Ti = 0.28–0.33 p.f.u..

The observed biotite-absent peak assemblage g–ksp–pl–q–ilm(–liq) cannot be well predicted in Fig. 8 because orthopyroxene, a common phase in the higher temperatures, is absent in the final assemblage. The measured matrix plagioclase compositions can also suggest temperatures above 1000 ℃ at 9 kbar, and a subsequent cooling path would not yield regular zoning as observed in sample JD1334. However, the measured X_{An} of 0.43–0.45 in the inclusion plagioclase could yield much higher temperatures of 990–1130 ℃ from 2–9 kbar (Fig. 8). If taking this into account, there would be two possible P–T paths: path-I dominant of cooling and path-II dominant of increasing pressure at first followed by cooling. The calculated changes in phase modes and compositions (Table 5) suggest that metamorphic evolutions along the both P–T paths would cause growth of garnet in the assemblages above the stability of bitoite, and garnet and orthopyroxene will be consumed in the biotite-bearing assemblages. Metamorphic evolution is dominated by melt crystallization via opx + ilm + pl + liq = g + ru + q + ksp in the modelled peak

assemblage g–ksp–opx–ru–ilm–pl–q–liq, and g + ilm + ksp + liq = bi + q + pl in the

final assemblage g–bi–ksp–ilm–pl–q–liq. These metamorphic reactions may have

been responsible for the disappearance of orthopyroxene and rutile in the peak

assemblage and appearance of biotite in the final assemblage.

The overprinting assemblage in this sample is neglected for its weak

development.

Sample JD15126

The P–T pseudosection calculated for sample JD15126 is drawn for plagioclase,

quartz and rutile in excess with the P–T window of 2–12 kbar and 700–1200 °C (Fig.

9). The fluid-absent solidus occurs at ~880 °C under 6–12 kbar. The observed final

assemblage g–bi–ksp–pl–q–ru–sill(–liq) defines a P–T range of 7–12 kbar and

880–890 °C. The measured X_{gr} of 0.03–0.05 and X_{An} of 0.34–0.30 define a pressure

range of 7.5–10 kbar in the final assemblage, while the measured X_{py} = 0.37–0.40 in

garnet and X_{Mg-bi} in the matrix biotite mostly suggest cooling conditions in the

subsolidus fields. The predicted Ti content in biotite is only 0.19 p.f.u. at the

biotite-out temperature, significantly lower than the measured values (Ti = 0.23–0.29

p.f.u.).

The observed peak assemblage g–ksp–pl–q–ru–sill(–liq) without biotite is

predicted to define a temperature range of 890–1080 °C under 9.5 kbar. The flat

profile with X_{gr} = 0.04–0.05 in the core of type-1 garnet (Figs. 5d, e) seems to suggest

a P–T path with deceasing pressure and temperature in between the X_{gr} isopleths of

gr4 and gr5. However, metamorphic evolution along this path will not cause the

growth of garnet. Thus, the flat X_{gr} profile could be better attributed to ion diffusion

and an isobaric cooling path is inferred under a pressure range of 9–10 kbar. The

decreasing of X_{gr} (0.04→0.03) from the narrow rim of type-1 garnet may suggest a

P–T vector dominant of pressure decrease (Fig. 9). The core-rim decreasing in X_{An} in

the coarse-grained matrix plagioclase may suggest that the growth of plagioclase may

have occurred by cooling or pressure increasing in the higher-T assemblages beyond

the stability of K-feldspar and sillimanite. The measured core-rim zoning with

decreasing X_{An} (0.49→0.47) from plagioclase inclusions in type-1 garnet could yield

a much higher temperature or lower pressure assemblages without K-feldspar and

sillimanite. Like that for sample JD1453, two possible P–T paths can be inferred as

shown in Fig. 9: path-I dominant of cooling and path-II dominant of increasing

pressure at first followed by cooling. The calculated changes in phase modes and

compositions (Table 5) indicate that metamorphic evolution may result in growth of

garnet and palgioclase together with melt crystallization during the early cooling stage

of path-I and the pressure-increasing stage of path-II, and metamorhic reactions are

modelled to be pl + liq = g + q + sill +ksp in the field of peak assemblage, and g + ksp

+ ru + liq = pl + q + sill + bi in the field of the final assemblage.

5. Zircon U–Pb dating

Samples JD1334，JD1453 and JD15126 were selected for zircon U–Th–Pb

isotopic analyses. The detailed analytical results are listed in Supplementary Table S1.

Thirty five analyses were performed on thirty-four zircon grains for *sample JD1334*. Zircon from this sample exhibits oval to round shapes with lengths ranging from 50–100 μm and length/width ratios of 1:1 to 2:1 (Fig. 10a). Cathodoluminescence images suggest two types of zircon grains: type I showing core-rim structure with blurred oscillatory zoning (e.g. spot #28, 33) or dark structureless characters in the core (e.g. spot #1, 20, 35), and with relatively high luminescence metamorphic growth rims (e.g. spot #21, 29), and type II showing sector zoning (e.g. spot #9). Due to the abnormal signal for analyses #6, #14, #15 and #30, they were excluded from age calculations. The remaining thirty one analyses were divided into three age groups (Fig. 11a). The first group comprises eleven analyses of oscillatory-zoning and/or structureless cores, with apparent $^{207}Pb/^{206}Pb$ ages ranging from 2525 ± 8 Ma to 3340 ± 12 Ma and Th/U ratios of 0.23–0.67 (av. 0.49), representing the ages of detrital zircon of magmatic origin. The second group consists of seventeen analyses of metamorphic growth rims and/or sector zoning, with apparent $^{207}Pb/^{206}Pb$ ages of 2427 ± 9 Ma to 2507 ± 9 Ma and Th/U ratios of 0.02–0.85 (av. 0.31). These analyses yield a weighted mean $^{207}Pb/^{206}Pb$ age of 2478 ± 13 Ma (MSWD = 7.8) in the concordia diagram (Fig. 11a). Spots #20, #02, #01 constitute the third age group and give apparent $^{207}Pb/^{206}Pb$ ages of 2388 ± 8 Ma, 2378 ± 8 Ma and 2078 ± 9 Ma with Th/U ratios of 0.04, 0.14 and 0.16, respectively. They plot on the concordia curve and spots #20, #02 define a concordia age of 2382 ± 6 Ma (MSWD = 0.022).

Thirty three U–Pb analyses were performed on thirty-one zircon grains for

Sample JD1453. Zircon in this sample mostly exhibits short prismatic and spherical-oval shapes with a size of 70–150 μm and length/width ratios of 1:1 to 2:1 (Fig. 10b). Most zircon grains are characterized by oscillatory zoned cores (e.g. spot # 19, 20) and metamorphic growth rims (e.g. spot # 2, 23). A few zircon grains display dark or irregular cores (e.g. spot #14, 30), and some grains show sector zoning (e.g. spot # 21). They were divided into three age groups (Fig. 11b). The first group involving twenty-three analyses from oscillatory-zoned cores have apparent ^{207}Pb/^{206}Pb ages ranging from 2502 ± 13 Ma to 2689 ± 13 Ma and Th/U ratios of 0.12–1.64 (av. 0.61), and plot on or near the concordia curve, except for two analyses (spots #19 and #24) which plot below the curve, indicating effect of Pb-loss. We suggest these oscillatory cores are of magmatic origin. The second group, including six analyses from metamorphic growth rims and grains with sector zoning, have apparent ^{207}Pb/^{206}Pb ages from 2482 ± 15 Ma to 2509 ± 10 Ma with a weighted mean ^{207}Pb/^{206}Pb age of 2498 ± 9 Ma (MSWD = 0.69), and relatively low Th/U ratios of 0.04–0.34 (av. 0.11). They plot on the concordia curve, yielding a concordant age of 2497 ± 5 Ma (MSWD = 0.014). The third group involving the remaining four analyses from dark or irregular cores, give much younger apparent ^{207}Pb/^{206}Pb ages from 2301 ± 13 Ma to 2404 ± 15 Ma with Th/U ratios of 0.48–0.64 (av. 0.56), three of which yield a concordant age of 2401 ± 7 Ma (MSWD = 0.031) with weighted mean ^{207}Pb/^{206}Pb age of 2400 ± 14 Ma (MSWD = 0.20).

Thirty spots were analyzed on thirty zircon grains for *sample JD15126*, where zircon shows lath and spherical-oval shapes with a size of of 50–110 μm, with

length/width ratios of 1:1 to 2:1 (Fig. 10c). Most zircon grains exhibit clear or blurred

oscillatory zoning (e.g. spots #1, #10, #18, #6 and #23), and a few grains are

relatively homogeneous with low luminescence (e.g. spots #13). Five analyses (spots

#7, #12, #14, #26 and #28) are rejected for age calculations, owing to their abnormal

signal peaks. The remaining twenty five analyses were divided into two age groups

(Fig. 11c). The first group including twenty two analyses from oscillatory zoning

exhibits apparent $^{207}Pb/^{206}Pb$ ages of 2505 ± 10 Ma to 3585 ± 12 Ma. The second

group including three analyses (spots #19, #29 and #13) shows $^{207}Pb/^{206}Pb$ ages of

2473 ± 13 Ma, 2478 ± 14 Ma and 2496 ± 12 Ma, and Th/U ratios of 0.02, 0.01 and

0.13, yielding a weighted mean $^{207}Pb/^{206}Pb$ age of 2483 ± 15 Ma (MSWD = 0.95) in

the concordia diagram (Fig. 11c).

Zircon with the first age group from the three samples have significantly high

chondrite-normalized REE values and the heavy rare earth element (HREE) contents

are particularly high, showing left-facing slopes on the chondrite-normalized REE

plots (Fig. 12), which further prove their magmatic origin. In contrast, zircons with

the second group show relatively low total REE, and distinctly low HREE contents,

exhibiting flat HREE patterns that could be a result of zircon-garnet coexistence

(Rubatto, 2002; Rehman et al., 2016), further supporting their metamorphic origin.

Zircon grains with the third age group show variant REE patterns, similar to those of

magmatic zircon in sample JD1453, and similar to those of metamorphic zircon in

sample JD1334.

6. Discussion and conclusions

6.1 Metamorphic evolution of pelitic granulites

Based on the petrographic characteristics and phase equilibria modelling, four generations of assemblages involving final, peak, inclusion and overprinting stages were identified for pelitic granulites from the eastern Hebei terrane.

P–T conditions of the final assemblages

The observed final assemblages with presence of biotite for the three samples yield similar P–T conditions of 8.0–9.5 kbar/870–890 °C (Figs 6–9), giving an apparent thermal gradient of 25 °C/km (Fig. 13). Although the low–T limits are constrained by the fluid-absent solidi that depend on the H_2O contents (higher H_2O contents yield lower solidus), the right match between the measured and the predicted mineral chemical compositions and modal proportions (Table 5) can provide additional constraints on the P–T conditions of the final assemblages or the location of the solidi. In addition, the locations of mineral composition isopleths in the suprasolidus fields would not be significantly affected by the solidus position. It is noteworthy that the predicted Ti content in biotite is significantly lower than the measured values, suggesting that the temperature of the final assemblages would be somewhat underestimated. The obtained P–T conditions for the final assemblages are in agreement with those of the previous studies for mafic granulites (Chen and Li, 1996; He and Ye, 1992; Zhao et al., 1998).

Isobaric cooling of the peak assemblages

The observed peak assemblages without biotite in samples JD1334 and JD15126

are modelled to be stable in large P–T ranges (Figs. 6, 9). In sample JD1334, the

measured core-rim zoning with X_{An} decreasing from coarse-grained plagioclase in the

matrix and weakly garnet core-rim zoning with X_{gr} increasing are predicted to be

consistent with a cooling path, where the isopleths of maximum X_{An} and minimum

X_{gr} from grains cores define a peak P–T condition of ~1050 °C and 9.8 kbar in the

field of peak assemblage. In sample JD15126, the flat X_{gr} profiles in the garnet core

could be attributed to ion diffusion and is better interpreted to reflect isobaric cooling

under 9–10 kbar. The observed peak assemblage in sample JD1453 cannot be well

modelled in pseudosection for orthopyroxene is predicted to be common under higher

temperatures but consumed through metamorphic reactions during cooling under

suprasolidus conditions. However, the flat compositional zoning of garnet and

plagioclase in the matrix from sample JD1453 only record the final solidus condition

and cannot be well used to constrain a cooling path.

Metamorphic evolution recorded by the plagioclase inclusions

 Plagioclase inclusions with high X_{An} in garnet in samples JD1453 and JD15126

are predicted to be stable at much lower-P and/or higher-T fields than the stabilities of

the peak assemblages. This suggests two possible P–T paths: an isobaric cooling

from > 1100 °C and a path dominant of pressure increasing at first followed by

cooling in the P–T fields of the peak assemblages. These two paths are consistent with

the measured zoning with core-rim decreasing in X_{An} from both inclusion and the

coarse matrix grain of plagioclase in sample JD15126 (Fig. 9). It should be noted that

there are considerable uncertainties for the phase modelling results in the fields

beyond the stability of peak assemblages due to the uncertainty of water contents. The water contents used for calculating the pseudosections were determined on the basis of the stability of the final assemblages. If the rocks contain lower water contents at their peak stages, the predicted differences in phase equilibria from those shown in Figs 6–9 are that the stability fields of K-feldspar and sillimanite will be enlarged and the amount of melt will decrease (White et al., 2001; Wei, 2016), which will result in considerable changes in phase topology. The low water contents in pelitic granulites may have been attributed to melt loss during prograde metamorphic evolution (White et al. 2001, 2002). However, the bulk rock compositions of the concerned samples are distinct from the UHT granulites that have experienced a large amount of melt loss with high $Mg^{\#}$ = 70–80 and low content of Na_2O and CaO (e.g. Kelsey et al., 2004). That is to say, there would not be a lot of melt lost from the studied samples. In addition, phase modelling of zircon growth or dissolution in melt-bearing system implies that zircon in high-grade rocks tends to be dissolved in melt (Kelsey and Powell, 2011). The presence of numerous detrital zircons also suggests lack of abundant melt in the peak stage, or at least, much less than the modelled values in Table 5.

There is one possibility for a pelitic rock to have low melt (or water) contents under granulite facies is that the rock underwent heating under low pressures. For example, if a rock was heated under low pressures along a P–T path as shown in Fig. 13, the rock will be dehydrated through decomposition of hydrous phases, such as muscovite and biotite under subsolidus conditions, and there still be very small

amount of melt formed when the P–T path meets the dry solidus of granodiorite, i.e. at 1010 °C/3 kbar. However, the X_{An} isopleths in plagioclase inclusions are plotted above the dry solidus. Phase modelling suggests that their locations will move to higher temperatures with decreasing H_2O contents, and the resulting melt mode will be very high even under dry conditions. Thus, we cannot rule out a possibility that the Ca-rich plagioclase inclusions in garnet may not represent the equilibrium metamorphic evolution as modelled in Figs. 8, 9, but represent local equilibrium conditions because of the fast moving of rocks or lack of fluids. As a consequence, the X_{An} isopleths of the plagioclase inclusion are not relied too much for constructing the P–T paths of the granulites, and the P–T paths are inferred to proceed under temperatures below the dry solidus in Fig. 13.

Although there are considerable uncertainties in the locations of the plagioclase X_{An} isopleths in the fields beyond the stabilities of the peak assemblages, they do suggest a possibility that the concerned rocks may have experienced UHT metamorphic stages with temperature much higher than that recorded by their final assemblages. This can be supported by the following lines of evidence:

(i) The measured Ti in biotite ranging from 0.27–0.33 p.f.u. is much higher than the modelled values on the biotite-out, suggesting the Ti-rich phase could be stable to higher temperatures if the stability fields of biotite could be enlarged by incorporating components, i.e. fluorine (e.g. Hensen and Osanai, 1994). Biotite from the Zhaojiayao granulites in the Khondalite belt of North China Craton shows similar Ti contents to those in our samples, which were estimated to be stable > 1000 °C (Li and Wei, 2016).

Moreover, experimental work of Stevens et al. (1997) suggests that the upper limits of Ti-biotite in the pelites and greywackes are > 950 °C at 9 kbar (Fig. 13), where biotite contains Ti up to 0.25 p.f.u., lower than the Ti contents in the concerned samples. Thus, the final assemblages with Ti-rich biotite could be stable to higher temperatures than the modelled values, and the peak assemblages without biotite could represent UHT conditions greater than 950 °C.

(ii) The pelitic granulites coexist with two pyroxene granulites and charnockites. The two pyroxene granulites are dominated by anhydrous minerals with variant amounts of amphibole where amphibole was observed to be retrograde phase by Chen and Li (1996). Although these mafic granulites were estimated to show peak temperatures of 866–922 °C by Chen and Li (1996) using conventional thermometers, they may suggest much higher temperature conditions according to experimental results that the disappearance of amphibole in mafic rocks mostly occurs in 1000–1100 °C (e.g. Moyen and Stevens, 2006, Fig. 13).

As a summary, the final, peak and inclusion assemblages in the pelitic granulites suggest an anti-clockwise P–T path involving early-stage heating under low pressures, followed by pressure-increasing to the P–T fields of the peak assemblages, and subsequent cooling to the P–T fields of the final assemblages. It is worthy noting that a P–T vector dominant of pressure-decreasing can be inferred from the decreasing X_{gr} in the narrow rim section of garnet in sample JD15126, indicating decompression under subsolidus conditions (Figs 9, 13).

Metamorphic evolution recorded by the overprinting assemblages

The pelitic granulites were differently overprinted by the assemblage

characteristic of formation of garnet coronae or outer rims around garnet, together

with the formation of tiny grains of plagioclase, K-feldspar, biotite and quartz etc. in

local domains. A case study from sample JD1334 suggests that the P–T conditions of

the overprinting assemblage are 10–12 kbar and 830–880 °C (Fig. 7). Chen and Li

(1996) observed that these overprinting textures were developed in various granulites

and mafic dykes and inferred that they were Paleoproterozoic products on the basis of

field geological context. Duan et al. (2015) reported that the mafic dykes have

experienced high-pressure granulite-facies metamorphism characteristic of clockwise

P–T paths with peak conditions of 11–13 kbar/790–810 °C and post-peak cooling

ages of ~1.82 Ga. The peak P–T conditions from the metabasic dykes are similar to

that estimated for the overprinting assemblage in sample JD1334 (Fig. 13), suggesting

they were the products of a same metamorphic event of high-pressure granulite-facies

in the Paleoproterozoic era. The Paleoproterozoic granulites show lower peak

temperature and higher peak pressure than those of the granulites of early stages.

6.2 Zircon U–Pb data of pelitic granulites

Zircon age dating for three pelitic granulites yields wide range of apparent

$^{207}Pb/^{206}Pb$ ages from 3.5 Ga to 2.1 Ga, which are divided into three groups. The first

age group from 3.5 Ga to 2.5 Ga with a peak at 2.6–2.5 Ga is obtained from zircon

grains characterized by oscillatory zoned cores and high HREE contents with

left-facing slopes on the chondrite-normalized REE plots, indicative of their

magmatic origin. The peak ages of 2.6–2.5 Ga are consistent with the abundant

magmatic precursors of TTG gneisses in eastern Hebei (e.g. Geng et al., 2006, 2010;

Yang et al., 2008; Nutman et al., 2011; Wan et al., 2012; Zhang et al., 2012; Guo et al.,

2013, 2015; Bai et al., 2014, 2015). The youngest detrital zircon age peak of ~2.5 Ga

in this group is interpreted as the sedimentary deposition time of the pelitic granulites.

In addition, the older ^{207}Pb/^{206}Pb ages from 3.5 Ga to 2.7 Ga of detrital zircon in the

group are consistent with those reported for eastern Hebei (e.g. Jahn et al., 1987; Liu

et al., 1992; Nutman et al., 2011, 2014).

The second age group with a weighted mean ^{207}Pb/^{206}Pb age of ~2.48 Ga is

obtained from zircon grains with metamorphic growth rims, sector zoning and low

HREE contents with flat HREE patterns, suggesting their metamorphic origin.

Accordingly, the 2.48 Ga is thought to be the metamorphic age of the first phase of

pelitic granulites coeval with the time of regional granulite-facies metamorphism

revealed in the previous studies for mafic granulites, TTG gneisses and charnockites

(e.g. Zhao et al., 1998; Geng et al., 2006; Nutman et al., 2011; Yang et al., 2015).

Phase equilibria modelling in the ZrO_2-bearing system suggests that in melt-bearing

high-grade metamorphic rocks zircon tends to be dissolved in melt and newly grown

zircon should record the age of melt crystallization during cooling (Kelsey and Powell,

2011; Zhang et al., 2013). Accordingly, the age of 2.48 Ga could be interpreted to

represent uplifting and cooling ages of granulite terranes (Fig. 13).

The third age group of ^{207}Pb/^{206}Pb ages from 2.4–2.1 Ga is obtained from zircon

grains with irregular character, and/or blurred oscillatory zoning and shows both

magmatic and metamorphic zircon REE pattern character, which was considered to be

complex or unknown origin. As there are only a few data in this group, it is hard to

interpret their real geological meanings (e.g. Guo et al., 2013, 2015; Bai et al., 2014,

2015).

6.3 Tectonic implications

As mentioned above, the Archean supracrustal rocks of granulite-facies commonly

occur as localized slivers or rafts in vast oceans of granitoids, and there is a tendency

that the higher the metamorphic grades the smaller the scales of the supracrustal

slivers or rafts. The anticlockwise P–T paths that involve heating, pressure-increasing,

cooling and uplifting stages recovered from the pelitic granulites in this study suggest

a tectonic scenario to explain the particular occurrence of the Archean supracrustal

granulites. Firstly, supracrustal sediments experienced thermal event under low

pressures attributed to the intrusion of high-temperature TTGs, resulting in

dehydrations of hydrous phases like muscovite and biotite under subsolidus

conditions (Fig.13). Then, the dried metasediments, potentially denser than the

surrounding lithologies tend to drop into the TTG magma ocean, causing the

pressure-increase in the metasediments. As the metasediments dehydrated, they could

skip intensive melting due to their fast dropping thus could prevent metamorphic

reequilibration even their P–T path could meet the dry solidus. Actually, as almost

every supracrustal sliver is interlayered with iron formations in the Taipingzhai region,

it could be expected that the metasediments dropped following the dense iron-bearing

rocks. Thirdly, the dropped metasediments experienced cooling under crustal level of ~35 km for the case we concerned in coupled with the crystallization of the surrounding TTG magma to form the observed final assemblages. Finally, they uplifted as a consequence of the doming processes in Archean terranes (Campbell and Hill, 1988; Collins et al., 1998; Parmenter et al., 2006; Lin and Beakhouse, 2013). It could be postulated that the lower-metamorphic grade greenschist belts represent their shallow dropping into magma oceans, and some supracrustal blocks of amphibolite-facies represent their deeper positions. Zircon U–Pb data suggest that the deposition age of pelitic granulites from eastern Hebei was ~2.5 Ga, and the metamorphic age was ~2.48 Ga, indicating that the heating, dropping and cooling processes occurred in an time interval of ~20 My.

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Figure captions

Fig.1 (a) Tectonic subdivision of the North China Craton (NCC) (after Zhao et al.,

2005, 2012). Abbreviations for metamorphic complexes: JP, Jianping; MY, Miyun; EH, Eastern Hebei; CD, Chengde; NH, Northern Hebei; XH, Xuanhua; HA, Huai'an; FP, Fuping; HS, Hengshan; WT, Wutai; LL, Lüliang; ZH, Zanhuang; GY, Guyang; JN-Jining; WD, Wulashan-Daqingshan. (b) Geological map of the Eastern Hebei (modified after Wu et al., 1998; Geng et al., 2006). I. Zunhua high-grade meta-greenstone belt; II. Taipingzhai granulite facies unit; III. Qian'an gneiss dome; IV. Luanxian amphibolite-facies metamorphic belt; V. Qinglong amphibolite–greenschist-facies metamorphic belt.

Fig.2 A geological map of eastern Hebei showing distribution of pelitic granulites.

Fig.3 Field and micrographs of representative pelitic granulite samples from eastern Hebei. (a) A field photo showing the interbedded relationship of pelitic granulites and magnetite quartzite. (b) Partial melting of pelitic granulites. (c) Ca map showing low Ca in the core increasing to high Ca towards inner rim (g_1), overgrown by high Ca outer rim (g_2) from sample JD1334. The red end of the colour spectrum represents high concentrations and the blue-black end represents low concentrations. (d) Anhedral medium- to coarse-grained plagioclase in sample JD1334 (crossed polars). (e) Anhedral fine-grained plagioclase, biotite, quartz and K-feldspar aggregates in sample JD1334. (f) Subhedral coarse-grained garnet porphyroblast (g_1) surrounded by tiny crystal garnet (g_2). The mineral abbreviations are as follows: bi – biotite; g – garnet; ilm – ilmenite; mt – magnetite; ksp – K-feldspar; pl – plagioclase; q – quartz;

opx – orthopyroxene.

Fig. 4 (a) Ca map showing high Ca in the plagioclase inclusion-rich core decreasing to low Ca at rim from sample JD1453. (b) Garnet porphyroblast without plagioclase inclusions from JD1453. (c) Coarse-grained plagioclase in sample JD1453 (crossed polars). (d) Ca map showing high Ca in the plagioclase inclusion-rich core decreasing to low Ca at rim from sample JD15126. (e) Garnet porphyroblast without plagioclase inclusions from JD15126. (f) Anhedral medium- to coarse-grained plagioclase in sample JD15126 (crossed polars). Abbreviations: sill – sillimanite; others are the same as those in Fig. 3.

Fig. 5 Compositional zoning profiles of X_{alm} (=Fe^{2+}/(Fe^{2+} + Mn + Mg + Ca)), X_{sps}, X_{py} and X_{gr} defined accordingly across garnet grains from the representative samples JD1334 (a), JD1453 (b, c) and JD15126 (d, e). (f) Compositional zoning of plagioclase characterized by decreasing X_{An} [= Ca/ (Ca + Na + K)] from core to rim in sample JD1334. (g) Zoning profiles of X_{An} of plagioclase in the matrix (g1) and plagioclase inclusions in garnet (g2) from sample JD1453. (h) Decreasing X_{An} from core to rim of plagioclase in the matrix (h1) and plagioclase inclusions (h2) in garnet grain from sample JD15126. Labels of A, B, g_1 and g_2 correspond to those in Figs. 3–4.

Fig.6 Calculated P–T pseudosection for sample JD1334 in the system

NCKFMASHTO (+ ksp + pl + q + ilm). The pseudosection is presented with isopleths of X_{py}, X_{gr} in garnet, X_{Mg-bi} in biotite, X_{Mg-opx} in orthopyroxene and X_{An} in plagioclase for relevant assemblages. Mineral abbreviations: cd – cordierite; ky – kyanite; liq – silicate liquid/melt; sp – spinel; ru – rutile; other abbreviations are the same as those in Figs. 3–4. Compositions used for modelling are listed in Table 2. The circles with the labels P, A and B are corresponding to those in Table 5.

Fig.7 Calculated P–T pseudosection for sample JD1334 (overprinting) in the system NCKFMASHTO (+ ksp + pl + q + ilm). Others are the same as those in Fig. 6.

Fig.8 Calculated P–T pseudosection for sample JD1453 in the system NCMnKFMASHTO (+ pl + q). The circles with the labels M, N, P, A and B are corresponding to those in Table 5. Others are the same as those in Fig. 6.

Fig.9 Calculated P–T pseudosection for sample JD15126 in the system NCKFMASHTO. Mineral abbreviation: Sa – sapphirine; Others are the same as those in Fig. 6.

Fig10 Cathodoluminescence (CL) images of selected zircons from pelitic granulite samples in eastern Hebei. Solid circles shows the inner positions of LA–ICP–MS analytical sites with their identification numbers as in Supplementary Table S1，and ages in Ma, respectively.

Fig.11 U–Pb isotopic age distributions of analyzed zircons from pelitic granulites in eastern Hebei.

Fig.12 Chondrite-normalized REE patterns (Sun and McDonough, 1989) for zircons of magmatic (group 1), metamorphic (group 2) and complex (group 3) origin from pelitic granulites in eastern Hebei.

Fig. 13 The summarized P–T conditions and paths for granulites from eastern Hebei. Thick blue arrows represent the P–T paths for pelitic granulites in our study. The P–T conditions of the final assemblages for the three samples in this article are shown as the translucent boxes labelled with JD1334, JD1453 and JD15126, respectively. The P–T condition recovered from the overprinting assemblages from sample JD1334 is shown as the translucent quadrangle labelled with "Overprinting". The P–T path labeled with Duan-15 is cited from Duan et al. (2015) for metabasic dykes of high-pressure granulite-facies. The X_{An} = 43–45 isopleths are from sample JD1453. The abbreviations are as follows: NG – 'normal' granulite; UHTG – ultrahigh-temperature granulite; E-HPG – eclogite-high-pressure granulite. The areas of NG, UHTG and E-HPG are cited from Brown (2007). The granite solidus and the disappearance of amphibole (amp-out) in tholeiite composition are adopted from Moyen and Stevens (2006). Mu-out line is from Spear and Cheney (1989). The lines of Ti-biotite-out (Ti-bi-out in pelite and Ti-bi-out in greywacke) are from Stevens et al.

(1997). The dry solidus of granodiorite is cited from Robertson and Wyllie (1971).

The transition lines of Al_2SiO_5 are calculated by THERMOCALC.

(b)

N

0   10   20 km

40°30′

I

I

Qinglong

Shuangshanzi

V

Fig.2

I

II

Santunying

Taipingzhai

Qianxi

Shuichang

Jinchangyu

III

Qian'an

Lulong

IV

Funing

(a)

110°E   115°E   120°E

TRANS-NORTH CHINA OROGEN

Jianping

Bayan Obo

GY   Yinshan Block   NH   CD   JP

XH

WD   JN   MY

HA   EH

WESTERN BLOCK

HS

Ordos Block   WT   FP   Beijinig

LL   ZH   EASTERN BLOCK

Taiyuan

Zunhua

118°30′   119°00′

Mesozoic granitoids	Mesozoic/Proterozoic cover	TTG gneisses	TTG gneisses/Charnockite
Supracrustal rocks of amphibolite-granulite facies	Supracrustal rocks of amphibolite facies	Supracrustal rocks of amphibolite-greenschist facies	

(a) pelitic granulite / magnetite quartzite

(b) leucosome

(c) g₂ / A / Fig.5a / ilm / g₁ / q / ksp / B / g₂ / Ca / 0.2mm

(d) pl / A / Fig.5f / pl / q / B / opx / opx / ksp / 0.2mm / 100 µm

(e) pl / pl / pl / pl / q / ksp / q / ksp / bi / q / aggregates of pl-ksp-bi-q / q / pl / g / bi / 0.5mm

(f) g₂ / g₂ / mt / g₁ / q / 0.2mm

NCKFMASHTO (+q+ksp+pl+ilm)　　　JD1334(overprinting)

NCMnKFMASHTO (+pl+q)   JD1453

P(kbar)

T(℃)

solidus
mb70 — $X_{mg} = 0.70$ in biotite (X_{Mg-bi})
py38 — $X_{py} = 0.38$ in garnet
gr5 — $X_{gr} = 0.05$ in garnet
an30 — $X_{an} = 0.30$ in plagioclase

g bi ksp ru ky
g bi ksp ru
g bi ksp ru liq
g ksp opx ru liq
g ksp bi opx ru liq ilm
g opx ru liq
g bi ksp ru liq ilm
g ksp opx ru liq ilm
g bi ksp ru ilm
g bi ksp ilm liq
g bi ksp ilm
g bi ksp opx ilm liq
g ksp opx ilm liq
g bi ksp opx ilm
g bi ksp opx cd ilm
g bi ksp opx cd ilm liq
g ksp opx cd ilm liq
g opx cd ilm liq
g opx ilm liq
g opx ru liq ilm
g opx ru ilm liq (-q)
g opx ilm liq (-q)
g opx cd ilm liq (-q)

NCKFMASHTO (+pl+q+ru) **JD15126**

P(kbar) vs T(℃) phase diagram

Field labels:
g ksp ky bi
g ksp bi sill liq
g ksp bi sill
g ksp sill cd liq
g ksp cd liq
g ksp cd bi
g ksp cd opx liq
g ksp cd bi opx
ksp cd opx liq
g cd opx liq
cd opx liq
g ksp sill liq
g sill liq
g liq
g liq(-pl)
g opx liq(-pl)
g opx liq
g opx liq(-pl-q)
opx liq
g opx liq(-q)
opx liq(-q)
g opx liq(-q)
cd opx liq(-q)

Contour labels: py37, py38, py39, py40, py43, py42, mb75, mb77, gr5, gr4, gr3, an30, an31, an32, an33, an34, an47, an49, an55

Points: A, B, P, N, M, I, II

Legend:
— solidus
--mb75-- $X_{mg} = 0.75$ in biotite (X_{Mg-bi})
--py40-- $X_{py} = 0.40$ in garnet
--gr5-- $X_{gr} = 0.05$ in garnet
--an30-- $X_{an} = 0.30$ in plagioclase
(1). g ksp liq (2).g ksp opx liq
(3). g ksp sill opx liq (4). g sill opx liq
(5). g opx sa liq

(a)JD1334
50μm
33
3340±12Ma
2613±15Ma
35
2480±10Ma
29
28
2558±9Ma
2502±9Ma
21
2499±9Ma
09
20
2388±8Ma
2078±9Ma
01

(b)JD1453
2644±14Ma
20
2521±13Ma
19
2489±13Ma
23
24
2512±13Ma
02
2504±10Ma
2496±13Ma
21
14
2394±12Ma
2404±15Ma
30

(c)JD15126
10
3504±11Ma
2947±11Ma
06
18
2607±13Ma
2523±13Ma
23
01
2589±10Ma
2496±12Ma
13
29
2478±14Ma

(a) JD1334

group 1
group 2
group 3

(b) JD1453

(c) JD15126

Table 1. Main mineral components, modal proportion (vol. %) and their locations of pelitic granulites from eastern Hebei

sample	location	garnet	biotite	sillimanite	K-feldspar	plagioclase	orthopyroxene	quartz	rutile	ilmenite	pyrite
JD1334	40°14.75'N, 118°31.12'E	25	3	-	16	30	2	23	-	1	+
JD1453	40°15.49'N, 118°28.78'E	12	10	-	5	50	-	23	-	+	-
JD15126	40°10.72'N, 118°41.92'E	21.5	3	2	11	32	-	30	0.5	-	-

-, absent and +, present (< 0.5 vol. %).

Table 2. Bulk-rock compositions of pelitic granulites from eastern Hebei

Sample	ICP-OES whole rock compositions (wt. %)												
	SiO_2	TiO_2	Al_2O_3	Fe_2O_3	MnO	MgO	CaO	Na_2O	K_2O	P_2O_5	LOI	A/CNK	$Mg^{\#}$
JD1334	58.32	0.72	16.43	13.83	0.12	3.02	2.76	2.23	2.19	0.11	0.29	1.48	30
JD1453	62.74	0.79	17.41	5.97	0.06	3.17	3.46	3.68	1.65	0.14	0.41	1.23	52
JD15126	62.71	0.90	16.60	9.32	0.08	3.24	2.20	2.27	1.86	0.08	0.39	1.70	41

Sample	Normalized molar proportion used for phase equilibria modelling											
	Figs	H_2O	SiO_2	Al_2O_3	CaO	MgO	FeO	K_2O	Na_2O	TiO_2	O	MnO
JD1334	Fig. 6	0.38	65.06	10.53	2.59	4.89	11.55	1.69	2.40	0.72	0.20	0.00
JD1334 (overprinting)	Fig. 7	0.40	65.90	10.95	3.50	3.64	10.33	1.80	2.66	0.72	0.10	0.00
JD1453	Fig. 8	1.21	67.73	11.15	3.74	4.99	5.32	1.37	3.85	0.48	0.09	0.08
JD15126	Fig. 9	0.40	68.74	10.63	2.67	5.55	7.30	1.25	2.47	0.68	0.30	0.00

Table 3. Selected microprobe analyses for pelitic granulites JD1334 and JD1453 from eastern Hebei

Mi	JD1334								JD1453								
	g_1-	g_1-	g_2	bi	pl-	pl-	ks	op	g_1-	g_1-	g_2	bi	bi	pl	pl-	pl-	ks
SiO	38.	38.	38.	38	60.	61.	64.	50	39.	38.	39.	36	37	56	60.	60	64
TiO	0.0	0.1	0.0	5.	0.0	0.0	0.0	0.	0.0	0.0	0.0	5.	5.	0.	0.0	0.	0.
Al_2	21.	22.	22.	14	25.	23.	18.	1.	22.	22.	22.	16	15	27	24.	24	18
Cr_2	0.0	0.0	0.0	0.	0.0	0.0	0.0	0.	0.0	0.1	0.0	0.	0.	0.	0.0	0.	0.
Fe	30.	30.	30.	15	0.0	0.1	0.0	28	26.	27.	26.	10	14	0.	0.1	0.	0.
Mn	0.3	0.2	0.3	0.	0.0	0.0	0.0	0.	0.4	0.4	0.4	0.	0.	0.	0.0	0.	0.
Mg	6.9	6.3	5.2	12	0.0	0.0	0.0	17	10.	9.8	10.	16	13	0.	0.0	0.	0.
Ca	1.8	2.4	4.4	0.	6.7	5.5	0.0	0.	1.4	1.4	1.8	0.	0.	9.	6.5	6.	0.
Na_2	0.0	0.0	0.0	0.	8.3	8.6	1.0	0.	0.0	0.0	0.0	0.	0.	6.	7.9	7.	0.
K_2	0.0	0.0	0.0	10	0.2	0.2	15.	0.	0.0	0.0	0.0	10	10	0.	0.4	0.	16
Tot	10	10	10	96	10	10	10	99	10	10	10	96	96	99	10	99	99
O	12.	12.	12.	11	8.0	8.0	8.0	6.	12.	12.	12.	11	11	8.	8.0	8.	8.
Si	2.9	3.0	2.9	2.	2.6	2.7	2.9	1.	3.0	2.9	3.0	2.	2.	2.	2.6	2.	2.
Ti	0.0	0.0	0.0	0.	0.0	0.0	0.0	0.	0.0	0.0	0.0	0.	0.	0.	0.0	0.	0.
Al	2.0	2.0	2.0	1.	1.3	1.2	1.0	0.	2.0	2.0	1.9	1.	1.	1.	1.2	1.	1.
Cr	0.0	0.0	0.0	0.	0.0	0.0	0.0	0.	0.0	0.0	0.0	0.	0.	0.	0.0	0.	0.
Fe^3	0.0	0.0	0.0	0.	0.0	0.0	0.0	0.	0.0	0.0	0.0	0.	0.	0.	0.0	0.	0.
Fe^2	2.0	2.0	2.0	0.	0.0	0.0	0.0	0.	1.6	1.7	1.6	0.	0.	0.	0.0	0.	0.
Mn	0.0	0.0	0.0	0.	0.0	0.0	0.0	0.	0.0	0.0	0.0	0.	0.	0.	0.0	0.	0.
Mg	0.8	0.7	0.6	1.	0.0	0.0	0.0	1.	1.1	1.1	1.1	1.	1.	0.	0.0	0.	0.
Ca	0.1	0.2	0.3	0.	0.3	0.2	0.0	0.	0.1	0.1	0.1	0.	0.	0.	0.3	0.	0.
Na	0.0	0.0	0.0	0.	0.7	0.7	0.1	0.	0.0	0.0	0.0	0.	0.	0.	0.6	0.	0.
K	0.0	0.0	0.0	0.	0.0	0.0	0.9	0.	0.0	0.0	0.0	0.	0.	0.	0.0	0.	0.
X(p	0.2	0.2	0.2	0.	0.3	0.2	0.9	0.	0.4	0.3	0.3	0.	0.	0.	0.3	0.	0.
Y(p	0.0	0.0	0.1						0.0	0.0	0.0						

X(g) = Mg/(Fe^{2+} + Mg + Ca); Y(g) = Ca/(Fe^{2+} + Mg + Ca); X(bi) = Mg/(Fe^{2+} + Mg); X(opx) = Mg/(Fe^{2+} + Mg); X(pl) = Ca/(Ca + Na + K); X(ksp) = K/(Ca + Na + K); -C,core; -R, rim; -I, in garnet; -M, in matrix; -MC, grain core in matrix; -MR, grain rim in matrix. The mineral abbreviations are the same as those in Figs. 3-5.

Table 4. Selected microprobe analyses for sample JD15126 from eastern Hebei

JD15126								
Mineral	g_1-C	g_1-R	g_2	bi-M	pl-I	pl-MC	pl-MR	ksp
SiO_2	39.21	39.14	39.56	38.63	55.15	58.73	60.07	64.43
TiO_2	0.01	0.05	0.00	5.18	0.00	0.03	0.00	0.03
Al_2O_3	22.33	21.94	22.44	14.58	27.98	25.71	24.98	18.45
Cr_2O_3	0.04	0.06	0.06	0.04	0.04	0.00	0.00	0.02
FeO	26.13	26.90	26.39	9.43	0.00	0.04	0.10	0.00
MnO	0.33	0.29	0.34	0.04	0.01	0.01	0.00	0.05
MgO	10.01	10.28	10.28	16.29	0.03	0.00	0.00	0.00
CaO	1.88	1.22	1.43	0.00	10.22	7.45	6.64	0.07
Na_2O	0.00	0.02	0.00	0.05	5.96	7.73	8.24	1.16
K_2O	0.01	0.00	0.00	9.88	0.05	0.23	0.32	15.42
Totals	100.02	99.9	100.58	94.13	100.04	100.00	100.35	99.65
O	12.00	12.00	12.00	11.00	8.00	8.00	8.00	8.00
Si	3.00	3.00	3.00	2.84	2.49	2.63	2.67	2.99
Ti	0.00	0.00	0.01	0.29	0.00	0.00	0.00	0.00
Al	2.01	1.98	2.01	1.27	1.49	1.36	1.31	1.01
Cr	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Fe^{3+}	0.00	0.02	0.00	0.00	0.02	0.00	0.00	0.00
Fe^{2+}	1.67	1.70	1.67	0.58	0.00	0.00	0.00	0.00
Mn	0.02	0.02	0.02	0.00	0.00	0.00	0.00	0.00
Mg	1.14	1.17	1.16	1.79	0.00	0.00	0.00	0.00
Ca	0.15	0.10	0.12	0.00	0.49	0.36	0.32	0.00
Na	0.00	0.00	0.00	0.01	0.52	0.67	0.71	0.10
K	0.00	0.00	0.00	0.93	0.00	0.01	0.02	0.91
X(phase)	0.38	0.39	0.39	0.75	0.49	0.34	0.30	0.90
Y(phase)	0.05	0.03	0.04					

Details are the same as those in Table 3.

Table 5. Comparisons of the calculated mineral compositions and modal proportions (on one-oxide basis) with those measured in samples

	P(kbar)	T(°C)	X(py)	X(gr)	X_{An}	$X_{Mg\text{-}bi}$	Ti-bi	$X_{Mg\text{-}opx}$	g	opx	ksp	pl	q	bi	ilm	liq	ru	sill	cd
JD1334 in Fig.6																			
P	9.78	1050	0.30	0.05	0.31			0.46	30.65	2.96	14.89	19.88	18.65		1.86	11.12			
A	9.10	880	0.28	0.06	0.30			0.48	33.00	4.18	17.23	20.81	20.40		1.88	12.50			
B	9.00	865	0.26	0.07	0.28	0.58	0.16	0.47	33.00	2.16	15.31	25.22	22.13	3.41	1.72	0.05			
Measured			0.25-0.27	0.05-0.07	0.26-0.31	0.59-0.66	0.26-0.32	0.52	32.60	2.65	12.72	25.69	21.79	3.28	1.26				
JD1453 in Fig.8																			
M	3.00	1010	0.35	0.03	0.43			0.58	0.83	12.61		30.14	1.99		1.80	50.06			3.26
N	9.00	1130	0.52	0.04	0.43			0.69	2.96	8.05		21.35	2.24		0.56	4.55	2.78		
P	9.00	1000	0.45	0.05	0.29			0.71	14.62	6.54	1.00	44.51	4.76		0.86	7.61	0.10		
A	9.00	900	0.43	0.06	0.30			0.63	17.65	5.60	8.05	41.90	7.34		0.58	8.66	0.23		

B	9.00	865	0.37	0.06	0.29	0.68	0.19	15.71	2.26	46.93	23.40	11.09	0.47	0.14	
Measured			0.38-0.41	0.04-0.08	0.29-0.31	0.62-0.73	0.28-0.33	16.40	4.15	44.78	22.81	11.47	0.40		
JD15126 in Fig.9															
M	3.00	1030			0.47		0.48	16.85	18.56	11.67	42.40		0.68		9.84
N	9.50	1154	0.50	0.05	0.47		0.64	12.59	5.17	6.96	12.04	62.56	0.68		0.00
P	9.50	1070	0.45	0.03	0.29			28.89	3.36	27.54	24.51	14.46	0.68		0.57
A	9.50	895	0.43	0.04	0.31			30.41	1.09	25.63	27.91	2.69	0.68		0.59
B	9.50	882	0.41	0.04	0.31	0.73	0.19	27.70	9.15	27.33	29.67	3.53	0.14	0.59	0.89
Measured			0.37-0.40	0.03-0.05	0.30-0.34	0.75-0.81	0.23-0.29	28.55	8.90	27.99	28.99	3.38	0.58		0.76

Measured $X_{(py)}$ such as 0.25-0.27 refers to the range of pyrope contents in garnet grains; P, A, B, M and N correspond to labels in Figs 6, 8 and 9.

Highlights

►Neoarchean anticlockwise P–T path with peak condition of > 950 °C

►Paleoproterozoic metamorphism with P–T condition of 10–12 kbar/830–880 °C

►An interval of ~20 Ma between deposition and metamorphism of pelitic granulites