HT–MP metamorphism in Central Qilian Block, NE Tibet Plateau: Implications on the tectonic evolution of the Qilian Orogen

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Abstract:
The metamorphism of the Central Qilian Block in the northeastern Tibetan Plateau records a complete tectonic history of the Qilian Orogen. Here, results of structural measurements, geochronological data, and thermobarometry for metamorphic rocks in the Central Qilian Block were presented, which trace the tectonic evolution of the Qilian Orogen. New U–Pb dating of detrital zircon from one paragneiss shows a main age population between c. 1500 and c. 1250 Ma, with the youngest age of 1085 Ma. This unit was intruded by an orthogneiss which has a U–Pb weighted mean age of 920 ± 18 Ma. Together with c. 1200–1000 Ma ages from inherited zircon cores in amphibolites, these results indicate that the protoliths of the Huangyuan Group were formed during the Mesoproterozoic and Neoproterozoic. Rims of these zircons obtain tightly constrained and concordant ages ranging from c. 459 to c. 427 Ma, with a weighted mean age of c. 450 Ma. Phase equilibrium modelling and conventional thermobarometry jointly indicate high-temperature/medium-pressure HT–MP metamorphism along a clockwise pressure–temperature (P–T) path at c. 450 Ma, passing through prograde conditions of 7.8–8.0 kbar and 620–650°C to peak conditions of ~7 kbar and ~780–800°C. Together with documented widespread metamorphism, magmatism, and ductile shear belts, these new results reassert that the Central Qilian Block experienced a three-stage tectono-metamorphic evolution during the Early Palaeozoic and the HT–MP metamorphism of Huangyuan Group was developed by a continental collision geodynamic setting with coeval mafic and granitic magmatism.

Key words: Metamorphism; *P–T–t* Path; Proto-Tethys ocean; Huangyuan Group; Central Qilian block

1 INTRODUCTION

Studies on the Palaeozoic orogenic collages in East Asia, such as the Qinling, Qilian and Kunlun orogenic belts indicated that several Precambrian blocks were located on the periphery of the northern margin of eastern Gondwana, which were involved in the evolution of the Proto-Tethys Ocean during the Early Palaeozoic (e.g., Cawood & Korsch, 2008; Fu et al., 2018, 2019; Li, Niu, et al., 2019; Sun, Dong, et al., 2020; Sun et al., 2022; Xiao et al., 2009; Yan et al., 2015; Yan, Fu, Aitchison, Buckman, et al., 2019, Yan, Fu, Aitchison, Niu, et al., 2019, Yan et al., 2022; Yu et al., 2017, 2019, 2021; Zhang, Zou, et al., 2013; Zhang, Gong, et al., 2013; Zhao et al., 2012; Zhao & Xiao, 2018). The spatial–temporal relationship of these blocks is vital to reasonably understand the evolution of these orogenic belts and the Proto-Tethys Ocean in the northeastern (NE) Tibet Plateau.

The Qilian orogenic belt in the NE Tibet Plateau (Figure 1a) is a typical Phanerozoic accretionary-to-collisional orogenic collage, which records the rifting of the Rodinia supercontinent through multiple episodes of subduction–accretion and continent–continent collision. As an important portion of this orogen, the Central Qilian Block—dominated by Precambrian rocks—is generally interpreted based on the geochemical and geochronologic data to have rifted from the Yangtze Craton in response to breakup of the supercontinent Gondwana and formation the Proto-Tethys Ocean during the latest Neoproterozoic (e.g., Fu et al., 2018; Tung et al., 2013, 2016; Wan et al., 2006; Xu et al., 2006; Yan et al., 2015; Zhang et al., 2017). Accretion of these blocks to Gondwana was driven by the closure of the Proto-Tethys Ocean during the Early Palaeozoic, thus creating Proto-Pangea: an early stage in the formation of the supercontinent Pangea (e.g., Li, Suo, et al., 2018; Li, Zhao, et al., 2018; Zhao & Xiao, 2018). Recently, most research performed on the Central Qilian Block has focused on the protolith age of Precambrian metamorphosed rocks (e.g., Huang et al., 2015; Tung et al., 2013; Yan et al., 2015), but the evolution and timing of metamorphism are still uncertain.

Conventional thermobarometry results demonstrate that metamorphic rocks of the Huangyuan Group along the northern margin of the Central Qilian Block experienced amphibolite-facies metamorphism, but their absolute peak pressure–temperature (P–T) conditions remain debated (Bai et al., 1998; Qi et al., 2004). This discrepancy significantly hinders understanding of the tectonic evolution of the Central Qilian Block and its associated ocean basin. In this contribution, we are exploring the P–T–time evolution of the Huangyuan Group by combining geothermobarometrical (P–T pseudosection modelling and conventional thermobarometry), and geochronological (zircon U–Pb dating) methods. This will lead to a better understanding of the tectonic processes involved in the evolution of the Qilian Orogen during the Early Palaeozoic period.

2 REGIONAL GEOLOGIC SETTING

2.1 The Qilian Orogen

The Qilian Orogen is a part of the orogenic collage of the Tethyan domain and is surrounded by several continental blocks, including the North China, Tarim, and Qaidam blocks (Figure 1a). A series of long-lived subduction and collision events in this region are documented in the Qilian Orogen by numerous island arc complexes, accretionary prisms, ophiolites, high-pressure (HP), ultrahigh-pressure (UHP) metamorphic rocks, and associated sedimentary basins (Fu et al., 2018, 2019; Li, Suo, et al., 2018; Song et al., 2013, 2017; Xia et al., 2011; Xu et al., 2016; Yan et al., 2015; Yan, Fu, Aitchison, Buckman, et al., 2019; Yan, Fu, Aitchison, Niu, et al., 2019; Zhang et al., 2017). The Qilian Orogen is traditionally divided into the North Qilian belt, the Central Qilian block, and the South Qilian belt based on distinctive lithographic units and geological structures (Figure 1b; e.g., Song et al., 2006; Xiao et al., 2009).

The North Qilian belt separates the Alxa Block to the NE from the Central Qilian Block to the southwest (Figure 1a,b). It consists of Early Palaeozoic arc-trench system rocks and Silurian-Triassic siliciclastic and carbonate rocks (Song et al., 2013, 2017; Xiao et al., 2009; Zhang et al., 2015), representing a typical accretionary orogeny. The Central Qilian belt, which is generally named the Central Qilian Block, is dominated by Precambrian rocks intruded with Early Palaeozoic granitoid plutons. It has been previously considered a micro-continental block rifted from the North China Block (Feng & He, 1996; Zuo & Liu, 1987). However, recent geochemistry and geochronology studies indicate a close affinity with the Yangtze Block (e.g., Tung et al., 2013; Wan et al., 2006; Yan et al., 2015). The South Qilian belt, separating the Central Qilian Block and the North Qaidam belt, is composed of Cambrian to Ordovician volcano-siliciclastic rocks and Late Ordovician–Silurian alluvial sediments (Yan, Fu, Aitchison, Buckman, et al., 2019). It records the subduction and closure of the Proto-Tethys Ocean during the Cambrian to the Early Silurian (Song et al., 2006; Fu et al., 2018, 2019; Fu, Yan, et al., 2020; Yan, Fu, Aitchison, Buckman, et al., 2019; Yan, Fu, Aitchison, Niu, et al., 2019).

The Precambrian rocks in the Central Qilian Block were subdivided into two units based on rock assemblage and metamorphism (BGMR-GP, 1989; BGMR-QP, 1991). The oldest subdivision includes the Yemananshan, Huangyuan, and Maxianshan groups, respectively exposed in the western, central, and eastern parts of the Central Qilian Block (BGMR-GP, 1989; BGMR-QP, 1991). Rocks of this subdivision were originally regarded as Palaeoproterozoic and consist of greenschist- to amphibolite-facies gneisses, migmatites, schists, Mg-enriched marbles, quartzites, amphibolites, and metavolcanic rocks intruded by Neoproterozoic (940–788 Ma) granitoids and mafic intrusive rocks. A younger subdivision was assigned as the Mesoproterozoic Danghe Group in the Yemananshan area, Huangzhong and Huashishan groups in the Huangyuan area, and Xinglongshan Group in the Maxianshan area. Rocks of this subdivision consist of low-grade metavolcanic and metasedimentary rocks.

2.2 The Huangyuan Group of the Central Qilian belt

The Huangyuan Group is part of the Precambrian basement in the Qilian Orogen spread over Huangyuan County, Huangzhong County, and Ledu County of Qinghai Province. The Huangyuan Group can be subdivided into the Liujiatai Formation and the Dongchagou Formation (BGMR-QP, 1991). Its lowermost unit, the Liujiatai Formation is ca. 1200–2450 m thick and mainly consists of greenschist- to amphibolite-facies schists, gneisses, marbles, quartzites, amphibolites, and minor volcano-sedimentary rocks. The upper unit, the Dongchagou Formation, consists of garnet-bearing quartz–mica schist and two-mica quartz schist intercalated with minor chlorite-quartz schist, amphibolite schist, marble, and quartzite in the lower part; quartz–mica schist, phyllite, thinner quartzite horizons, and minor chlorite-quartz schist and amphibolite schist in the upper part (Yan et al., 2022). Geochemical data and geological surveys indicate that their protoliths are intermediate-basic volcano-sedimentary rocks deposited in an epicontinental back-arc basin (Guo et al., 2000; Tung et al., 2013; Wan et al., 2006). Granulite-facies to amphibolite-facies metamorphism were reported affected the rocks (Bai et al., 1998; Peng et al., 2017).

The Huangyuan Group exposed in the Zhongniuchang area was taken as the study area in this contribution (Figure 2). Gneiss and amphibolite are exposed widely. The foliation strikes predominantly ENE-WSW and dips at 42°–59° toward the SSW. Many E-W-trending thrust faults with throws to the north occur in this area and juxtapose different lithostratigraphic units together. Amphibolite often occurs as lenses within paragneiss or mica–quartz schist, with all units overlain conformably by marble (Figure 3c, Figure 4a). Orthogneiss intruded into the paragneiss (Figure 3d). Meanwhile, deformations develop well within the Huangyuan Group in the Zhongniuchang area. One shear belt crosses Huangyuan Group and is parallel to the trend of Qilian orogenic belt. The outcrops are characterized by small-scale folds, S/C (schistosité–cisaillement) shear fabrics, rotated feldspar porphyroblasts, and asymmetric folds (Figure 4c,d), showing a dextral sense of motion with minor reverse components. Furthermore, microscopic observations confirm the widespread mylonitization of the dextral ductile shear belts (Figure 4e). The mylonitization is characterized by the dynamic recrystallization of quartz, and it produced a range of mylonitized rocks, protomylonites, mylonites, and ultramylonites, with increasing intensity of deformation (Figure 5a–f).

3 ANALYTICAL METHODS

Six samples, including one orthogneiss (18ST01), two paragneiss samples (18WJF18, 19ZN27), and three amphibolites (19ZN09, 19ZN22, and 19ZN29), were collected from the Huangyuan Group in the Zhongniuchang area (Figure 2; GPS: 37° 20′ 50″ N, 108° 11′ 12″ E).

Electron Probe Microanalysis (EPMA) was conducted on amphibolite samples 19ZN22 and 19ZN29, and paragneiss 19ZN27 to obtain mineral compositions for thermobarometry. This was conducted at the School of Resources and Environmental Engineering, Hefei University of Technology (China), by using a SHIMADZU JXA-8230 electron microprobe. The following oxides were analysed: SiO2, TiO2, Al2O3, Cr2O3, FeO, MnO, MgO, CaO, Na2O, and K2O. The working conditions and analytical methods followed those of Li, Li, et al. (2019). These data are reported in full in Table S1.

All six amphibolite and gneiss samples were subjected to zircon U–Pb geochronology. Zircon was separated by standard heavy liquid and magnetic separation techniques, followed by hand-picking under a binocular microscope. Selected grains were mounted in an epoxy resin, polished to expose the grain centre, photographed in transmitted and reflected light, and imaged using cathodoluminescence (CL). Zircon U–Pb geochronology was performed at the School of Resources and Environmental Engineering, Hefei University of Technology (China) by Laser Ablation Inductively Coupled Plasma Mass Spectrometry (LA-ICP-MS). The LA-ICP-MS system comprises an Agilent 7500a ICP-MS and laser ablation system with a COMPex PRO 102 ArF-Excimer laser source (λ = 193 nm). The analytical procedures used are described by Sun, Niu, et al. (2020). The weighted mean ages and concordia plots of the zircon were calculated using Isoplot v. 3.23 (Ludwig, 2003). These data are reported in Table S2.

Whole-rock compositions were obtained for amphibolite sample 19ZN29 and paragneiss 19ZN27 via a P61-XRF26S X-ray fluorescence (XRF) spectrometer fusion method at the laboratory of ALS Chemex (Guangzhou) Co. Ltd., Guangzhou, China. The following oxides were measured: SiO2, TiO2, Al2O3, Fe2O3T, MnO, MgO, CaO, Na2O, K2O, P2O5, and loss-on-ignition (LOI). The proportion of ferric and ferrous iron was determined by titration. The sample digestion procedure and analytical accuracy for major elements were described in detail by Li, Li, et al. (2019). These data are reported in Table 1.

4 PETROGRAPHY AND MINERAL CHEMISTRY

4.1 Petrography

4.1.1 Amphibolite

Sample 19ZN09 consists mainly of garnet (20%), amphibole (30%), plagioclase (35%), quartz (12%), and opaque minerals (<1%) (Figure 5a). The equidimensional garnet porphyroblasts with a diameter ranging from 0.5 to 1.0 mm contain inclusions of clinopyroxene, amphibole, epidote, plagioclase, quartz, rutile, and ilmenite (Figure 5a). Occasionally, matrix plagioclases with lengths ranging from 0.5 to 1.5 mm are retrogressed and exhibit sericitic alteration. Amphibole crystals with lengths of 0.3–1.2 mm in the matrix have inclusions of plagioclase, quartz, ilmenite, and magnetite.

Sample 19ZN22 is composed of garnet (35%–40%), amphibole (30%–35%), plagioclase (3%–10%), biotite (3%–10%), and quartz (3%–10%) with minor zoisite (up to 2%) (Figure 5b). Accessory minerals include late ilmenite, pyrite, and zircon. The garnet porphyroblasts occur as fine- to medium-grained (0.5–1 mm) subhedral crystals and host very few inclusions. Large plagioclase occurs as single crystals or banded aggregates in the matrix, with crystal sizes of 0.2–1.8 mm. Meanwhile, a few plagioclase crystals are retrogressed with sericitic alteration. The amphibole occurs as a fine-grained matrix (0.2–1.5 mm) coexisting with biotite, or as a fine-grained relict showing a round reaction boundary with biotite (Figure 5b). It commonly shows straight and regular contact boundaries with garnet, amphibole, and biotite, indicating textural equilibrium between all minerals. Minor fine-grained magnetite coexists with chlorite, possibly formed at a very late period.

Sample 19ZN29 mainly consists of garnet (24%–44%), amphibole (17%–35%), plagioclase (5%–12%), biotite (5%–10%) and quartz (2%–10%), and minor opaque minerals (Figure 5c). Garnet porphyroblasts with lengths of 0.1–1.0 mm contain a few small inclusions of amphibole, plagioclase, quartz, rutile, and ilmenite. Some garnet porphyroblasts are enveloped by biotite and amphibole. Amphibole porphyroblasts with lengths of 0.5–2.0 mm usually shows euhedral to subhedral. Plagioclase occurs as single crystals or banded aggregates in the matrix, with crystal sizes of 0.2–1.5 mm. Plagioclase is sometimes replaced by clay minerals. The biotite occurs as subhedral, fine-grained (0.2–0.5 mm) matrix grains intergrown with plagioclase, amphibole, and quartz, indicating a thermal equilibrium state. Minor calcite, titanite, and pyrite occur in the matrix, which may occur in the late period (Figure 5c). They demonstrate that the peak mineral assemblage is Grt + Hbl + Bt + Pl + Qz.

4.1.2 Gneiss

Sample 18ST01 is an orthogneiss with granoblastic texture and is composed of plagioclase (55%), biotite (10%), and quartz (30%) (Figure 5d). Most alkali feldspar grains are microcline, 0.2–2 mm in width, with well-developed grid twinning. The crystals of plagioclase and quartz show dynamic recrystallization and the biotite presented mica-fish, indicating the deformation in late stage of the orogenesis (Figure 5d). Titanite, apatite, and zircon are the common accessory minerals.

Sample 18WJF18 is a paragneiss that contains garnet (6%), plagioclase (35%), biotite (10%), quartz (45%), and minor ilmenite (Figure 5e). Garnet is euhedral or subhedral, with diameters varying between 1.0 and 4.0 mm. They are in contact with biotite, quartz, and plagioclase, and the cores of the garnets contain inclusions of ilmenite, biotite, and quartz. Biotite occurs as small grains aligned in the foliation or as coarse prismatic plates that occur at steep angles to the foliation planes; however, some large biotite grains in the matrix are rimmed by ilmenite. Plagioclase occurs throughout the matrix. Quartz polycrystalline aggregate bands with superimposed oblique oriented fabric, indicating the ductile shear deformation.

Sample 19ZN27 is a paragneiss that consists of plagioclase (27%), quartz (45%), biotite (15%), garnet (8%), sillimanite (5%), and small amounts of ilmenite (Figure 5f). Euhedral garnet grains are 3–4 mm in width and contain inclusions of biotite and quartz in their cores, whereas the rims have no inclusions. The biotite occurs as small relict grains, with length of 0.1–0.5 mm. Sillimanite mostly occurs as coarse-grained prismatic needles that define the matrix foliation and is commonly spatially associated with biotite. Plagioclase occurs throughout the matrix, 0.5–1 mm in width. Some large biotite grains in the matrix are replaced by ilmenite. Garnet porphyroclasts and biotite, and sillimanite aggregates, oriented, and arranged, indicating the characteristics of high-temperature (HT) and superimposed low-temperature deformed microstructure. The peak assemblage is interpreted to be Grt + Bt + Sil + Ilm + Pl + Qz.

4.2 Mineral compositions

4.2.1 Garnet

Garnet porphyroblasts in amphibolite samples 19ZN22 and 19ZN29, and paragneiss sample 19ZN27 mostly exhibit high almandine [Fe2+/(Fe2+ + Mg + Ca + Mn)] and pyrope [Mg/(Fe2+ + Mg + Ca + Mn)] contents, but low grossular [Ca/(Fe2+ + Mg + Ca + Mn)] and spessartine [Mn/(Fe2+ + Mg + Ca + Mn)] contents (Table S1). In most cases, cores are characterized by broad zones of homogeneous composition, with rims showing inflections in Fe and Mg, indicative of equilibration with matrix minerals during retrograde cooling.

Fine-grained garnet of amphibolite 19ZN22 has a high content of almandine of 70–88 mol%, pyrope of 7–18 mol%, grossular of 4–10 mol%, and little spessartine (Figure 6a). Garnet porphyroblasts in 19ZN22 show some changes in almandine, pyrope, and grossular, from core to rim, with gradually decreased pyrope and grossular and increased almandine. It indicates a preserved peak metamorphic composition in the core and a retrograde metamorphic composition in the rim possibly (e.g., Caddick et al., 2010; Florence & Spear, 1991; Spear & Wolfe, 2019).

Garnet porphyroblasts in amphibolite 19ZN29 have high almandine contents of 74–86 mol% and grossular contents of 12–20 mol%, low pyrope of 2–5 mol% and negligible spessartine (<1 mol%), with the increase of almandine and the reduction of the grossular and pyrope from core to rim (Figure 6b). Garnet porphyroblasts in this sample show some relatively obvious changes in almandine and pyrope from core to rim, with gradually decreased pyrope and increased almandine. It is speculated that the garnet core records the composition of the metamorphic peak stage, while the garnet rim resulted from diffusion and dissolution, formed at decompression stage.

Garnet porphyroblasts in paragneiss sample 19ZN27 obtain high contents of almandine (92–95 mol%) and pyrope (4–7 mol%) and low contents of grossular (0.01–0.05 mol%), with spessartine nearing zero (Figure 6c). Compared with amphibolites, paragneiss possess higher content of almandine and lower content grossular and spessartine. In 19ZN27, from the core to the rim, almandine and pyrope show slight changes. It indicates a preserved peak metamorphic composition in the core and a retrograde metamorphic composition in the rim possibly.

4.2.2 Other minerals

Brown amphibole in the matrix from amphibolite 19ZN22 is categorized as magnesiohornblende (Ti = 0.20–0.23; Si = 6.53–6.63; p.f.u.), with Mg# [Mg/(Mg + Fe^{2+})] of 0.55–0.58 (Table S1; Figure 7a). There is no clear difference in composition between amphibole cores and rims, indicating a single stage of growth and/or compositional equilibration with the surrounding rock matrix. Correspondingly, the amphibole in the matrix from amphibolite 19ZN19 is also classified as magnesiohornblende with Ti (p.f.u) contents of 0.20–0.22 and Si (p.f.u) contents of 6.53–6.56 (Table S1; Figure 7a). The little difference in amphibole composition between cores with rims reveals homogeneous character.

Plagioclase feldspar in the matrix of amphibolite 19ZN22 could be classified as bytownite (Smith, 1984), with XAn = Ca/(Ca + Na + K) of 0.85–0.90, XAb = Na/(Na + Ca + K) of 0.09–0.15, and negligible XOr = K/(Na + Ca + K) of 0.00–0.01, and exhibits no compositional zoning across individual grains (Table S1). Plagioclase in gneiss 19ZN27 is oligoclase, with XAn mostly between 0.15 and 0.16, XAb between 0.82 and 0.84 (Figure 7b). Plagioclase of the two samples is relatively homogeneous.

Biotite in paragneiss sample 19ZN27 could be classified as ferro-biotite and has a Mg# of 0.49–0.50 and Ti = 0.28–0.44 p.f.u. for a 22 oxygen calculation (Foster et al., 2001). Biotite in 19ZN27 shows little change in composition, indicating biotite is homogeneous.

5 U–Pb GEOCHRONOLOGY

5.1 Amphibolite

Sample 19ZN09 Zircon grains from this sample are dark grey to colourless, rounded to ovoid in shape, and mostly subhedral (Figure 8a) with an average length of c.150 μm and aspect ratios of 1.5–2. Most zircon grains in the sample are very low or low luminescent, although when this is present, it reveals an indistinct planar zonation. Some core-rim textures occur, although most grains are homogenous. Eighteen concordant (concordance ≥90%) analyses were obtained from the core and rims of sample 19ZN09, and 14 analyses in the rims form a concordant cluster yielding a mean 206Pb/238U age of 445 ± 6 Ma (MSWD = 0.4) (Figure 8a; Table S2). The six core analyses yield older ages of ~1.0–1.2 Ga and so likely represent inherited/xenocrystic domains (Figure 8a). Most zircon rims exhibit highly variable U and Th contents of 242–1577 ppm and 12–1253 ppm, respectively. The 14 analyses in rims show low Th/U ratios of 0.01–0.33, and the six inherited zircons exhibit higher ratios of 0.71–0.91. Rare earth element (REE) profiles in zircons show different characters between the cores and rims. The cores exhibit relatively high ∑REE contents, steep HREE patterns with negative Eu anomalies and positive Ce anomalies. While the rims have a relatively low content of ∑REE with weakly negative Eu anomalies (Figure 10a).

Sample 19ZN22 contains dark zircons that are euhedral to subhedral in shape, measure up to ~300 μm in length, and show aspect ratios of 2:1–3:1. Most zircon grains in the sample lack luminescence, although if present, it reveals indistinct planar zonation (Figure 8b). Twenty-five concordant analyses were obtained from this sample, and all have variable Th (10.0–84.4 ppm) and U (35.70–1710 ppm) contents. All analysed spots have low Th/U ratios, ranging from 0.02 to 0.28. Twenty-five analyses yield a weighted mean 206Pb/238U age of 453 ± 4 Ma (MSWD = 0.29; concordance ≥90%; Figure 8b; Table S2). No inherited zircon grains were collected.

Sample 19ZN29 contains grey zircons (Figure 8c) that are rounded to ovoid in shape. They have lengths of 100–300 μm with aspect ratios of 1:1–3:1 and most zircon grains in the sample lack luminescence. They show planar growth banding, firtree sector zoning, weakly oscillatory zoning, or no zoning. Twenty-three spots were analysed and they have Th/U ratios of 0.02–0.28. Most of them exhibit low Th/U ratios, indicating that the zircons are metamorphic (Rubatto & Hermann, 2007). All analyses form a concordant cluster yielding a mean 206Pb/238U age of 459 ± 6 Ma (MSWD = 0.73; concordance≥90%; Figure 8c; Table S2).

5.2 Gneiss

Sample 18ST01 contains zircons with an average of 200–250 μm in their longest dimension and aspect ratios mostly between 2.5:1 and 3:1 (Figure 9a). In this sample, most samples exhibit narrow rims, representing the latter magmatism, and the rims are hard to be analysed by LA-ICP-MS. Zircons in 18ST01 display typical oscillatory growth zoning, indicating a magmatic origin (Hoskin & Schaltegger, 2003). Seventeen concordant ages are obtained, ranging from 963 to 899 Ma. All analyses form a concordant cluster yielding a mean 206Pb/238U age of 920 ± 8 Ma (MSWD = 3.7; concordance ≥90%) (Figure 9a; Table S2). Seventeen concordant analyses exhibit variable Th (55.6–182.1 ppm) and U (187.5–670.1 ppm) contents, with Th/U between 0.05 and 0.57.

Sample 18WJF18 contains zircon grains with an average size of 100–200 μm in their longest dimension and aspect ratios mostly between 2:1 and 2.5:1 (Figure 9b). In this sample, rims are too narrow to analyse by LA-ICP-MS, even though most zircons exhibit core-rim structures. CL images show that most zircon displays oscillatory growth zoning, reflecting a magmatic origin (Hoskin & Schaltegger, 2003), and the cores have a lighter colour than the narrow rims. There are still some zircons displays irregular, subangular, and subrounded to rounded shapes with distinct erosional pits, suggesting they have experienced long-distance transportation in the processes of deposition. Fifty spot analyses were obtained from 50 zircon grain cores, which all show concordant (concordance ≥90%) Mesoproterozoic to Palaeoproterozoic ages (Figure 9b). All detrital zircons form a major age group at 1836–1085 Ma with two prominent peaks at ca. 1250 and ca. 1500 Ma, and subordinate peaks at ca. 1700 and ca. 1800 Ma (Table S2). The measured age of the youngest zircon yielded a 207Pb/206U age of 1085 Ma. The zircons display ratios of Th/U in the range 0.2–1.62. The 1085 Ma age thus represents the maximum depositional age, indicating the onset of protolith deposition.

Sample 19ZN27 contains zircon with an average length of 100–200 μm and aspect ratios mostly between 2:1 and 3:1. CL images reveal that most grains contain oscillatory-zoned or unzoned cores and are surrounded by CL-bright, banded or unzoned rims (Figure 9c). Analyses were performed on both the core and rim domains. Twenty-three concordant points were obtained from sample 19ZN27, which can be divided into three groups. Most zircon cores in this sample are too small to be analysed, but six oscillatory-zoned relict cores produced six concordant 207Pb/206U age of 1.29, 1.61, 1.45, 1.76, 1.71, and 1.31 Ga (Figure 9c; Table S2; spots 5, 11, 22, 26, and 28). Two cores gave two younger concordant 206Pb/238U ages of 506 Ma and 513 Ma (spot 08, spot 14). Eighteen analyses from the rims form a concordant (concordance≥90%) cluster yielding a mean 206Pb/238U age of 445 ± 6 Ma (MSWD = 0.4) (Figure 9c). Zircons exhibit various contents of U and Th, which fall within the ranges of 6.65–77.8 and 448–1652 ppm, respectively. The rim exhibits Th/U ratios lower than 0.1, indicating a metamorphic origin. The REE profiles are different for cores and rims. The cores exhibit relatively high HREE contents and steep HREE patterns with negative Eu anomalies and positive Ce anomalies. While the rims are characterised by relatively depleted and flat chondrite-normalized HREE patterns with weakly negative Eu anomalies (Figure 10b). Combined with the zircon morphology, geochemistry, and age, we propose that 19ZN27 records metamorphism at 445 Ma. Ages of 0.51–1.76 Ga are from the inherited zircons, indicating the protolith of the 19ZN27 was a sedimentary rock with a long history of sedimentation.

6 METAMORPHIC HISTORY

To constrain the metamorphic evolution in north margin of Central Qilian Block, a combined approach of using conventional thermobarometry and phase diagram based thermobarometry has been applied to representative samples 19ZN22, 19ZN27, 19ZN29. Together, these results constrain various parts of an extended metamorphic evolution.

6.1 Conventional thermobarometry

Temperature and pressure estimations were calculated for the amphibolite sample 19ZN22 and paragneiss sample 19ZN27. Mineral compositions used for calculations and results are given in Table S1. Garnet-biotite-plagioclase-quartz (GBPQ) geothermobarometry (Wu et al., 2004) was used to constrain the metamorphic P–T conditions for gneiss sample 19ZN27, and garnet-hornblende–plagioclase-quartz (GHPQ) geothermobarometry (Eckert et al., 1991) was used for amphibolite sample 19ZN22. For hornblende and plagioclase, core compositions were used to minimize the effect of diffusion during cooling and retrograde metamorphism. In addition, the titanium-in-biotite thermometer of Henry et al. (2005) was applied to gneiss sample 19ZN27 to determine the range of conditions over which matrix biotite growth occurred.

For amphibolite sample 19ZN22, a mean P–T condition of 8.2 kbar and 800°C was obtained using compositions for porphyroblasts with cores of garnet, hornblende, and plagioclase. The total range of results was 6.0–9.5 kbar and 708–890°C. For the paragneiss sample 19ZN27, the compositions of garnet core, biotite, and plagioclase porphyroblasts in the matrix were used in GBPQ geothermobarometry to produce mean P–T conditions of 7.3 kbar at 730°C, with a range of 6.3–7.7 kbar and 703–762°C. This mean P–T point is taken to represent peak conditions of paragneiss sample 19ZN27. In addition, the Ti-in-biotite thermometer for matrix biotite in paragneiss sample 19ZN27 produced a range of crystallization temperatures of 652–722°C and a mean value of 697°C. These constraints from conventional geothermobarometry from both rock types thus indicate HT and medium-pressure (MP) metamorphic conditions.

6.2 Petrological modelling

While conventional thermobarometry is useful for constraining P–T conditions based on mineral compositions, cation diffusion at upper-amphibolite and granulite-facies metamorphic conditions is common and may alter the geochemical characteristics of minerals that formed during prograde or peak metamorphism through intracrystalline diffusion (Caddick et al., 2010; Yardley, 1977). In most cases, cation exchange does not alter the mineral assemblage itself—only their equilibrium compositions—such that matching observed parageneses against the stability fields of calculated parageneses derived via petrological modelling can provide a complementary and potentially more robust way to constrain the thermal evolution of high-grade metamorphic rocks (Diener et al., 2008). These data may be interpreted alongside the results of conventional thermobarometry.

Phase diagram modelling was applied to paragneiss sample 19ZN27 and amphibolite sample 19ZN29 from the Zhongniuchang sequence in the Central Qilian Block at P–T conditions of 4–10 kbar and 600–850°C. All calculations were performed using the petrological modelling programme Theriak-Domino (de Capitani & Petrakakis, 2010) and internally consistent thermodynamic dataset ds-62 of Holland and Powell (2011), updated February sixth, 2012. Pelitic paragneiss sample 19ZN27 was modelled in the MnO–Na2O–CaO–K2O–FeOtotal–MgO–Al2O3–SiO2–H2O–TiO2–O2 (MnNCKFMASHTO) system using activity-composition (a–x) relations for garnet, biotite, muscovite, epidote, plagioclase–K-feldspar, staurolite, chlorite, ilmenite–haematite, magnetite-spinel, orthopyroxene, cordierite, and silicate melt from White et al. (2014). Pure phases included rutile, andalusite, sillimanite, kyanite, quartz, and aqueous fluid (H2O). Amphibolite sample 19ZN29 was modelled in the NCKFMASHTO system using the following a–x relations: melt, augite, and hornblende (Green et al., 2016); garnet, orthopyroxene, biotite, and chlorite (White et al., 2014); olivine and epidote (Holland & Powell, 2011); magnetite-spinel (White et al., 2002); ilmenite-haematite (White et al., 2000); C-bar1 plagioclase and K-feldspar (Holland & Powell, 2003). Pure phases comprised albite, quartz, rutile, sphene, and aqueous fluid (H2O).

The bulk-rock compositions used for modelling (Table 1) were calculated by conversion of the whole-rock ICP-MS analyses to model-ready bulk compositions

(Palin, Weller, Waters, & Dyck, 2016). Individual FeO and Fe2O3 contents for samples were determined via Fe-Vol05 titration, and the amount of aqueous fluid (H2O) was determined by LOI measurements performed during XRF analysis. Mineral abbreviations are after Whitney and Evans (2010). The amount of H2O plays an important role in HT phase equilibria system (Diener et al., 2008). To estimate the amount of H2O, T–M(H2O) pseudosections are constructed below to assess the effect of reduced water content on the peak assemblage. T–M(H2O) pseudosections were computed at 7.5 kbar, corresponding to the pressure conditions calculated by the conventional thermobarometry.

6.2.1 Paragneiss sample 19ZN27

The anhydrous and total LOI value are taken as the starting point and terminal point respectively. The T–M(H2O) pseudosection for pelitic sample 19ZN27 allows the effect of variations in H2O to be modelled over the range from 0 (M(H2O) = 0) to 0.94 (M(H2O) = 1). The predicted mineral proportions of the 19ZN27 is stable above the solidus over a wide range of H2O contents. A value of 0.75 mol% H2O (M(H2O) = 0.80) was chosen as an appropriate value for this sample. A pseudosection calculated for sample 19ZN27 is shown in Figure 11a. The phase equilibria correlate well with those predicted for metamorphosed high-Al sediments (Palin & Dyck, 2020), indicating that the protolith for sample 19ZN27 was likely mudstone or siliceous shale. The LOI-derived H2O content predicts a fluid-saturated solidus over the entire pressure range of consideration at a temperature of 670–680°C, suggesting that sample 19ZN27 was a first-generation sediment and did not experience melt loss, even though previously determined P–T conditions from the GBPQ thermobarometer are high enough for partial melting to begin.

The observed mineral assemblage Grt–Bt–Pl–Ilm–Sil–Qtz is stable over a broad range of P–T conditions at >7 kbar and >700°C (Figure 11a). The low-temperature (>700°C) limit of this calculated assemblage field is defined by the destabilization of garnet, and at HP (>7.5 kbar), this is limited by the kyanite–sillimanite polymorphic transition. When pressure is lower than ~7.8 kbar, cordierite is calculated to stabilize, although this mineral is not observed within sample 19ZN27, indicating a minimum peak pressure of 7–7.5 kbar at 750–850°C. It is also noteworthy that magnetite (Mt) is predicted to be stable up to 8.5 kbar at these temperatures; however, its calculated proportion comprises less than 0.5 vol%, which lies within the magnitude of error of mineral volume proportions observed in thin sections that is associated with “geological” uncertainty (e.g., natural heterogeneity; Palin, Weller, Waters, & Dyck, 2016). The composition isopleths of XMg = 0.28 and XCa = 0.02 (Table S1), chose from the garnet mantle, intersect exactly at the upper stability field of the observed early mineral assemblage, indicating the peak metamorphic P–T conditions of ~730°C and ~7.9 kbar. The garnet exhibits the obvious growth zoning, with the diverse composition content from the core to rim, indicating the content of the rim represents the retrograde metamorphism. The composition isopleths of XMg = 0.25 and XCa = 0.02, chosen from the garnet rim, intersect at a P–T condition of ~650°C and ~7.2 kbar, representing the retrograde metamorphic P–T conditions.

Conventional thermobarometric estimates of peak metamorphism obtained via the GBPQ geothermometer (6.3–7.7 kbar and 703–762°C) show a degree of overlap with the peak assemblage field defined on this calculated pseudosection, indicating that final textural equilibrium prior to solidification of incipient partial melt occurred at around 7–7.5 kbar and 700–750°C. Petrographic evidence of feldspar in paragneiss contains abundant inclusions of rounded plagioclase, quartz, and biotite, as well as rare inclusions of amphibole that could reveal the partial melting (Figure 5f). Experimental studies suggest that the presence of an H2O-bearing fluid can decrease the temperature of the solidus, and melting of biotite-bearing rocks in the presence of hydrous fluid at 6 kbar begins at temperatures of about 680°C (Watkins et al., 2007). This assemblage is compatible with the generalized fluid-present melting reaction: Bt + Pl + Qtz + H2O-rich fluid = Hbl + Pl + Ttn + melt (e.g., Lappin & Hollister, 1980; Li, Zhao, et al., 2018). Therefore, water-fluxed biotite-breakdown melting is likely to account for partial melting.

The general lack of microstructural evidence for prograde or retrograde mineralogical change in sample 19ZN27 does not allow tight constraints to be placed on the early or late thermal evolution, although the retrograde metamorphic path is expected to have involved considerable cooling in tandem with exhumation; otherwise, cordierite would be expected to form. The results of Ti-in-biotite thermometry further indicate recrystallization of matrix biotite across the retrograde suprasolidus-subsolidus transition through the temperature range 652–722°C, as summarized in Figure 12.

6.2.2 Amphibolite sample 19ZN29

A pseudosection for sample 19ZN29 is shown in Figure 11b. The phase equilibria closely match those expected in metamorphosed mid-ocean ridge basalt (MORB) and similar mafic precursor lithologies at amphibolite- and granulite-facies conditions (Diener et al., 2007; Palin, White, & Green, 2016; Palin, White, Green, Diener, et al., 2016), although sample 19ZN29 is less aluminous, which expands the stability field of biotite to a slightly higher grade. The calculated solidus is fluid saturated at high and low pressures, but slightly undersaturated at medium pressures (5–7.5 kbar), indicating that incipient partial melting may have occurred in the sample. This is supported by the presence of minor leucosome in outcrops (Figure 3f), although there are no microstructural or outcrop-scale indications of melt loss in sample 19ZN29. Calcic amphibole (hornblende, sensu lato) and quartz are stable across the P–T range of interest, and garnet is stable at P > 5 kbar above the solidus. Plagioclase becomes unstable at P > 8 kbar in the suprasolidus regime.

The observed mineral assemblage Grt–Hbl–Pl–Qtz–Bt–Ilm in sample 19ZN29 is calculated to be stable at P–T conditions of 6–8 kbar and 710–800°C, which is constrained at HT by the loss of biotite and the stabilization of clinopyroxene, which are not observed in the sample. The HP limit of this assemblage field is defined by the breakdown of plagioclase, and to low pressure by the stabilization of magnetite and the loss of garnet. This peak assemblage field overlaps with P–T conditions determined from GHPQ conventional thermobarometry at ~7.5–8 kbar and 770–800°C, which may be taken as a robust estimate of the conditions of peak metamorphism, which fall within the upper amphibolite facies. Observed versus calculated mineral proportions for all major phases closely correlate within this peak assemblage field, indicating that although mineral compositions may have changed during retrograde metamorphism (e.g., Fe–Mg exchange), volume proportions have remained unchanged. The compositions of the garnet core (XMg = 0.13, XCa = 0.27), and the garnet rim are taken to further calculate the peak and retrograde metamorphism P–T conditions, but the predicted XMg values in garnet do not intersect the interpreted peak field. These inconsistencies may reflect partial diffusional resetting during retrogression from the metamorphic peak or could be due to inappropriate effective bulk compositions related to the assumed equilibrium volume. Garnet porphyroblasts are partly replaced by cordierite and biotite, which are inconsistent with isobaric heating but compatible with decompression at HT. The lack of orthopyroxene in the sample suggests that the rock did not decompress into orthopyroxene-bearing fields during exhumation, which requires pressures of >3 kbar for the retrograde evolution.

7 DISCUSSION

7.1 Age of the Huangyuan Group

The age of the Huangyuan Group was originally defined as early Sinian, although from 1978, it was reassigned to be Early Palaeoproterozoic, based on the single 2469 Ma zircon U–Pb age of migmatitic granite in the Huangyuan area (Wang et al., 1983). However, whole-rock Rb-Sr isochron ages and detrital zircon U–Pb dating results suggest the Mesoproterozoic (BGMR-QP, 1991; Lu et al., 2009). Guo et al. (2000) and Tung et al. (2013) suggested that the age of the Huangyuan Group is Neoproterozoic based on TIMS and SHRIMP zircon dating results. Wan et al. (2006)

suggested the basement of the Qilian Block was formed in the period of 0.8–1.0 Ga (the Jinningian Period) by the zircon dating results. While Zhang et al. (2021) obtain LA-ICPMS U–Pb dating of 1176–2492 Ma from paragneiss in the Huangyuan Group and deduce that the Huangyuan Group was formed during the Mesoproterozoic period. Thus, precise geochronology studies are essential to resolve this uncertainty.

In this contribution, we analysed some inherited zircons from samples 19ZN09 and 19ZN27, which mainly produced ages between 1.7 and 1.0 Ga. In addition, a paragneiss sample (18WJF18) was sampled from the upper part of the Huangyuan Group and was used for U–Pb dating of the Huangyuan Group. These obtained ages define prominent peaks at 1.25 and 1.5 Ga, with most ages mainly ranging from 1085 to 1600 Ma. The youngest detrital zircon age (1085 Ma) represents the maximum age of deposition in the Huangyuan Group. Further, orthogneiss 18ST01, an intruding gneiss, was dated and produced a mean 206Pb/238U age of 920 ± 8 Ma, representing the protolith age of the granitoid. Tung et al. (2007) suggested that the Huangyuan Group was younger than 882 Ma based on the youngest zircon grains of a schist sample. Lu et al. (2009) suggested that deposition of the Huangyuan Group occurred from 1456 to 917 Ma according to the youngest detrital zircon grains of a mylonitized quartz–mica schist sample from the Dongchagou Formation in the Baoku River area and associated oldest gneissic granitoid intrusion. Combining these results with the 930–910 Ma ages of granites that intruded into the Huangyuan Group and the detrital zircon ages from the schist and the metasedimentary gneiss (Guo et al., 1999, 2000; Tung et al., 2013; Zhang et al., 2016; unpublished data), we propose that the Huangyuan Group experienced a long period of deposition and that the Huangyuan Group is the Mesoproterozoic to Neoproterozoic basement. The deposition likely lasted from 1.6 to 1.0 Ga, and the maximum depositional ages are c. 1.0 Ga, which may locally have continued to c. 884 Ma, as indicated by the reported youngest detrital age from the group (Tung et al., 2013, 2016). It is inferred that diagenesis mainly occurred in the late Mesoproterozoic to early Neoproterozoic, and the Huangyuan Group is the Mesoproterozoic to Neoproterozoic basement.

7.2 Metamorphic evolution of Huangyuan Group

Phase equilibrium modelling indicates that the peak P–T conditions of pelitic paragneiss sample 19ZN27 and amphibolite sample 19ZN29 are ~7–7.5 kbar and 700–750°C and ~7.5–8 kbar and 770–800°C, respectively. For both samples, these conditions are associated with an approximate uncertainty of 1 kbar and 50°C (Palin, Weller, Waters, & Dyck, 2016). These conditions overlap with the results of conventional thermobarometry for each sample, which define a peak-to-near-peak (early retrograde) exhumation path along a linearized geothermal gradient of ~40°C/km (Figure 12). This is supported by the temperatures of crystallization of matrix biotite in sample 19ZN27 (652–722°C; brown band in Figure 12) overlapping the fluid saturated solidus. These data suggest that the northern margin of the Huangyuan Group experienced upper amphibolite facies metamorphic conditions of around 7 kbar and 780–800°C (yellow star; Figure 12) at an effective geothermal gradient of ~35°C/km, which reveals that the 450 Ma metamorphic rocks in Huangyuan is at medium pressure-high temperature (MP/HT) condition. In addition, the phase equilibrium modelling indicates a retrograde stage of 7.2 kbar, 650°C by the rim of the garnets in sample 19ZN27.

Constraints on the prograde metamorphic P–T evolution can be provided by the mineral inclusions in garnet cores. The garnet core of gneiss 19ZN27 contains biotite, quartz, and kyanite, revealing the prograde conditions in the amphibolite facies pressure at around 7.8–8.0 kbar and 620–650°C in the pseudosection, indicating a clockwise P–T path (Figure 12). Amphibolite 19ZN29 contains ilmenite, rutile, hornblende, and biotite in the core of the garnet. This assemblage is also consistent with a clockwise P–T path (Figure 12), which is characteristic of heating and compression followed by decompression. These sequences of tectonic processes are probably linked to metamorphism at convergent margins, starting with subduction, arc-related magmatism, and leading to continental collision, driven by crustal thickening, thermal relaxation, and subsequent exhumation (e.g., Vissers, 1992; Zhang et al., 2018).

The Central Qilian Block, is one of the small Precambrian microcontinents in East Asia that has been extensively metamorphosed. However, reported P–T conditions show a wide variation and apparently conflicting evolutions. Geothermobarometry performed by Bai et al. (1998) obtained an anticlockwise P–T path with peak P–T conditions of 3.8–4.8 kbar and 558–645°C. Qi et al. (2004) obtained peak P–T conditions of 6.2–7.7 kbar and 651–763°C from a felsic mylonite in the Huangyuan Group of the Central Qilian Block. Recently, Peng et al. (2017) used phase equilibrium modelling to obtain peak metamorphic P–T conditions for Huangyuan Group 500 Ma granulite phase rocks of ~700°C and ~6.5 kbar, which lay along an anticlockwise P–T path. By contrast, used the same technique, but reported a clockwise P–T path with a peak P–T stage (7.8–8.5 kbar, 660–690°C).

When combined with the peak P–T conditions in this contribution (7 kbar and 780–800°C), two different P–T conditions can be recognized: low pressure-high temperature metamorphism (LP/HT) and MP/HT metamorphism. Interestingly, the LP/HT conditions represent the upper amphibolite to granulite facies at c. 500 Ma, while the MP/HT amphibolite facies conditions occurred at c. 450 Ma separately (Li, Li, et al., 2019; Peng et al., 2017; Qi et al., 2004; this contribution). This interpretation is confirmed by the U–Pb zircon ages presented in this work, which demonstrate that the rocks underwent two stages of high-grade metamorphism during the Cambrian and Ordovician, respectively. The two-stage metamorphism is also verified by two stages micro-deformation recorded in garnet and plagioclase prophyroblasts in the Central Qilian Block (Cao et al., 2015). The early deformation events recorded in garnet give Lu-Hf ages of 512 Ma. While the younger record in plagioclase porphyroblast without monazite grains was dated by the monazite grains in the matrix gave the U–Pb ages of no earlier than 481 Ma, with NE–SW horizontal bulk shortening.

Previous studies have revealed that ductile shear zones developed on the north margin of the Central Qilian Block, with individual widths of tens of metres in a band of ~800 km long and 5–6 km wide (e.g., Qi et al., 2004; Sun et al., 2022; Xu et al., 2006). Gneisses and amphibolite have been deformed into protomylonite, mylonite, and locally ultramylonite, which can also be verified by the samples in our manuscript under the microstructure and petrography (Figure 4b–d). The late ductile shearing overprinted the early stage of metamorphism, and the minerals in the matrix, such as quartz, biotite, and mica have deformed and even possessed dynamic recrystallization. The results of 40Ar/39Ar age dating on K-feldspar and biotite from mylonites, and the relationships among the ductile shear zone, strata, and magmatic intrusions indicate that ductile deformation occurred between 440 and 390 Ma (Qi et al., 2004; Yong et al., 2008; Zhang & Xu, 1995). Quartz c-axis plots of mylonite samples from the north margin of the Central Qilian Block indicate deformation temperatures of 350–500°C (Qi et al., 2004; Sun et al., 2022). Thus, this ductile shear zone in Central Qilian was formed by large-scale transpression that resulted from an oblique continental collision. In other words, the Huangyuan Group experienced low temperature (LT) deformation/metamorphism, which could be verified by the micro-deformation in our samples (18ST01-19ZN27; Figure 5d–e). As such, metamorphism in the Huangyuan Group is multi-stage: (M1) a c. 500 Ma upper amphibolite- to granulite facies-metamorphism (HT-LP; ~700°C, ~6.5 kbar), (M2) a c. 450 Ma amphibolite-facies metamorphism (HT-MP; 780–800°C, 7 kbar), and (M3) a later 440–390 Ma LT metamorphism associated with dextral ductile shearing (350–500°C).

7.3 Mechanisms of MP/HT metamorphism and tectonic implications

The MP/HT metamorphic rocks (~450 Ma) exposed in the Central Qilian Block are characterized by a sequence of mineral zones—chlorite, biotite, garnet, and sillimanite—which record a clockwise pressure–temperature–time (P–T–t) path. The Huangyuan ~450 Ma HT–MP metamorphic rocks experienced typical clockwise P–T paths, which are assumed to be typical of the continental mass in a crustal section doubled in thickness by collision (Brown, 1993; Thompson & Ridley, 1987; Wei et al., 1998). Thus, collisional thickening could reasonably account for the formation of Huangyuan HT–MP metamorphic rocks. Meanwhile, anatexis and migmatization can also be observed in the field (Figure 3f), indicating that conditions locally reached at least 700°C (Brown, 2010). The age of MP/HT metamorphism spanned 460–440 Ma, which indicates that it was a progressive process. Furthermore, the previous study indicated that the Huangyuan Group could be divided into different metamorphic zones by metamorphic mineral assemblage (BGMR-QP, 1991; Li, Suo, et al., 2018). One zone is the garnet metamorphic zone, including garnet-bearing micaschist with layers of garnet-bearing plagioclase amphibolite. And another zone is the sillimanite metamorphic zone, which includes garnet-bearing biotite schist and sillimanite-bearing biotite-plagioclase gneiss. These characteristics indicate that the MP/HT metamorphism could be regarded as a Barrovian-type metamorphism (Barrow, 1912). The presence of a Barrovian metamorphism (medium pressure/low–high temperature) is typical of many collisional orogens (e.g., Burg & Gerya, 2005; Liou et al., 2004), revealing the MP/HT metamorphism in Huangyuan Group was formed by the collisional thickening during the Late Ordovician.

The metamorphic heating of Barrovian metamorphism was generally taken to result from thermal equilibration of an overthickened crust, involving distributed internal heating by radioactive decay in addition to conduction of heat from the base of the crust (Bickle et al., 1975; England & Thompson, 1984; Oxburgh & Turcotte, 1974; Viete et al., 2013). However, this model requires long time scales (10 s of Myr) to produce significant heat for metamorphism. Recently, models for the Barrovian metamorphism have combined short-term heating with a broad regional heating event that resulted from thermal relaxation (Ague & Baxter, 2007; Baxter et al., 2002). They argued that the Barrovian metamorphism resulted from the overprinting of a significant and widespread metamorphic thermal regime relating to overthickening and thermal relaxation by a brief (<1 Myr), episodic heating event. Regionally, coeval mafic and granitic rocks were broadly found in the Qilian orogenic belt, including arc-related volcanism with ages of 470–440 Ma in the North and South Qilian belts (Fu, Yan, et al., 2020; Song et al., 2017; Yan, Fu, Aitchison, Niu, et al., 2019; Yan et al., 2022; Yang et al., 2019; Yong et al., 2008; Zhang & Xu, 1995) and syn-collisional magmatism in the Qilian orogenic belt (460–450 Ma; Figure 1b; e.g., Chen et al., 2016, 2018; Huang et al., 2015; Song et al., 2013; Yong et al., 2008; Yu et al., 2021). This mafic and granitic magmatism is widespread in the Qilian orogenic belt, but is small-scale. As a result, the mafic and granitic magmatism coeval with MP–HT metamorphism may have provided an additional heat contribution to develop and maintain an elevated geotherm. A continental collision geodynamic setting with coeval mafic and granitic magmatism most likely accounts for MP–HT metamorphism in the Central Qilian Block, which was associated with the closure of the Proto-Tethys Ocean.

8 CONCLUSIONS

The maximum age of deposition of the Huangyuan Group is 1085 Ma, and the orthogneiss intruded into the paragneiss yields an age of 920 Ma, indicating that the Huangyuan Group represents Mesoproterozoic to Neoproterozoic rocks.

Phase equilibrium modelling and conventional thermobarometry suggest that this region experienced HT–MP metamorphism at 443–459 Ma along a clockwise P–T path, with peak conditions of 780–800°C and 7 kbar, and retrograde conditions of ~650°C and 7.8 kbar.

HT–MP metamorphism in Central Qilian Block resulted from continental collision geodynamic setting with coeval mafic and granitic magmatism.

Credit authorship contribution statement

Yi Sun: Formal analysis, Investigation, Writing-original draft. Manlan Niu: Supervision, Funding acquisition, Writing-review. Richard M. Palin: Writing-review & editing. Xiucai Li: Investigation, Project administration, review. Changlei Fu: Investigation. Chen Li: Investigation.

Declaration of competing interest

The authors declare that they have no known competing financial interests or personal relationships that could have appeared to influence the work reported in this paper.

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Figure captions

Fig. 1. (a) Simplified tectonic map of China. (b) Simplified geological map of Altun-Qilian-Qaidam (AQQ) orogenic system, showing the main lithotectonic units (modified from Fu et al., 2018).

Fig. 2. Geological map of the Zhongniuchang metamorphic Complex and locality of sampling profile (modified from BGMR-QP, 1991).

Fig. 3. Representative field photographs of representative samples from the Zhongniuchang area. (a) Marble. (b) Amphibolite (19ZN29). (c) Amphibolite as lenses lying in the gneiss. (d) Fine grained granitoid gneiss (18ST01). (e) Garnet grains in gneiss. (f) Leucosome can be occasionally found in the outcrop.

Fig. 4. (a) Simplified cross section in Zhongniuchang area. (b) Mylonite contains rotated feldspar porphyroclasts indicating the dextral sense of shear. (c) The arkose quartzite vein indicating the dextral sense of shear. (d) The dextral Fold in the outcrop. (e) A rotated dextral mica porphyroclast.

Fig. 5. Representative photomicrographs in this contribution. (a) Amphibolite 19ZN09. (b) Amphibolite 19ZN22. (c) Amphibolite 19ZN29. (d) Granitoid gneiss 18ST01. (e) Psammitic paragneiss 18WJF18. (f) Pelitc paragneiss 19ZN27.

Fig. 6. Zircon U-Pb concordia and weighted mean age diagrams for 19ZN09 (a, b), 19ZN22 (c, d), 19ZN29 (e, f). (g) Representative cathodoluminescence (CL) images of zircon grains.

Fig. 7. Zircon U-Pb concordia and weighted mean age diagrams for 18ST01 (a, b), 18WJF18 (c, d), 19ZN27 (e, f). (g) Representative cathodoluminescence (CL) images of zircon grains.

Fig. 8. Chondrite-normalized REE patterns for zircon from sample 19ZN09 (a) and 19ZN27 (b).

Fig. 9. Compositional zoning profiles of major minerals in amphibolite samples 19ZN09 (a), 19ZN22 (b), 19ZN29 (c) and gneiss sample 19ZN27 (d).

Fig. 10. (a) Composition of amphibole. (b) Composition of feldspar.

Fig. 11. Calculated P–T pseudosections for samples (a) 19ZN27 and (b) 19ZN29. Mineral abbreviations are after Whitney and Evans (2010). Bulk compositions used for modeling are given in Table S3. GHPQ – garnet-hornblende-plagioclase-quartz; GBPQ – garnet-biotite-plagioclase-quartz. Numbered assemblages are as follows: (a) 1 – Ms Ab Pg Ilm Ky Bt Pl Qtz; 2 – Ms Ilm Sil Bt Pl Qtz H_2O; 3 – Ilm Ky Bt Pl Qtz H_2O; 4 – L Ilm Ky Bt Pl Qtz H_2O; 5 – L Ilm Sill Bt Pl Qtz; 6 – Ilm Mt Sill Bt Pl Qtz H_2O; 7 – L Crd Ilm Bt Pl Qtz; 8 – L Crd Mt Ilm Opx Bt Pl Qtz; 9 – L Crd Ilm Opx Pl Qtz; (b) 1 – L Grt Ilm Ep Bt Hbl Qtz H_2O; 2 – Grt Ilm Ep Pl Bt Hbl Qtz; 3 – L Grt Ilm Hbl Qtz; 4 – L Grt Ilm Rt Bt Hbl Qtz; 5 – L Grt Ilm Hbl Qtz; 6 – L Grt Ilm Pl Hbl Qtz; 7 – L Grt Ilm Pl Aug Bt Hbl Qtz; 8 – L Grt Ilm Pl Bt Hbl Qtz Mt; 9 – L Ilm Mt Pl Bt Hbl Qtz H_2O; 10 – L Grt Ilm Aug Mt Pl Hbl Qtz; 11 – L Grt Ilm Aug Opx Hbl Pl Qtz.

Fig. 12. Summary *P*–*T* grid showing the aggregated results of petrological modeling and conventional thermobarometry. Shaded regions mark the interpreted peak *P*–*T* conditions for samples 19ZN27 and 19ZN29. Results of Ti-in-biotite thermometry are shown by a brown band, with a bold line representing the mean temperature. Aluminosilicate polymorph equilibria are from Pattison (1992), and abbreviations are as follows: And – andalusite; Ky – kyanite; Sil – sillimanite; GHPQ – garnet-hornblende-plagioclase-quartz; GBPQ – garnet-biotite-plagioclase-quartz.

Fig. 13. Tectonic model illustrating the evolution of the North Qilian Ocean: (a) The initial subduction of the North Qilian Ocean (Early-Middle Ordovician); (b) the closure time of the North Qilian Ocean (Late Ordovician).

Captions for Supplementary material

Table S1. LA-ICPMS zircon U-Pb data for metamorphic rocks from the Huangyuan group.

Table S2. EPMA data for metamorphic rocks from the Huangyuan group.

Table S3. Whole-rock composition from the Huangyuan group.