Journal of Metamorphic Geology Accepted manuscript

DOI: 10.1111/jmg.12647

Barrovian and Buchan metamorphic series in the Chinese Altai: P−T−t−D evolution and tectonic implications

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Running title: P−T−t−D evolution of the Altai Orogen

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Abstract

The relations between Barrovian- and Buchan-type metamorphic series in the Chinese Altai remain obscure, and hence a representative region of the central part of the Chinese Altai was investigated to address this issue, using combined microstructural, petrological and geochronological methods. In the region, Barrovian-type garnet, staurolite, kyanite and sillimanite zones are locally overprinted by Buchan-type andalusite- and cordierite-bearing domains. Microstructural analysis shows that Barrovian garnet, staurolite and kyanite grew synchronously with the earliest regional metamorphic foliation $S1_{B}$. A sillimanite-bearing assemblage locally overprinted the assemblage of the staurolite zone in a foliation parallel with $S1_{B}$, assigned as $S1_{M}$. The originally subhorizontal $S1_{B-M}$ foliation, metamorphic zones and mineral isograds were folded by F2 upright folds, leading to their inclination and juxtaposition to upper crustal levels. Subsequent D3 deformation affected heterogeneously all previous structures producing vertical high-strain zones around low-strain domains. The D3 high-strain zones in the vicinity of Permian pegmatites are associated with Buchan-type metamorphism and are characterized by syn-D3 growth of andalusite and cordierite. Phase equilibria modelling of the staurolite/ kyanite-bearing assemblages suggests a prograde **P–T** path with an apparent

DOI: 10.1111/jmg.12647

thermal gradient of _23_C/km associated with the S1$_{B}$ fabric. The partial re-equilibration occurred in the sillimanite stability at _670_C, corresponding to an apparent thermal gradient of _34_C/km in the S1$_{M}$ fabric. Garnet to sillimanite metamorphic zones were subsequently exhumed without apparent re-equilibration during the D2 event. The interpreted pressure–temperature (**P–T**) evolution of the Buchan-type metamorphism on the basis of thermobarometry and phase equilibria modelling suggests significant heating processes, corresponding to apparent thermal gradient of 41_C/km or more. In situ U–Pb dating of monazite inclusions in staurolite revealed predominantly 280–260 Ma ages and minor older ages scattering between 350 and 290 Ma, interpreted as important monazite recrystallization during the D3 event. Monazite in andalusite and cordierite yielded only ages of 280–260 Ma, interpreted as dating the growth of these minerals. Lu–Hf garnet–whole rock isochron of a garnet– cordierite– chlorite schist gave an age of ca. 262 Ma, overlapping in time with the age of monazite in the cordierite. Combined with available regional data, the results suggest that the Barrovian-type metamorphic cycle reflects a continued burial heating followed by decompression, probably connected with the Devonian suprasubduction tectonic switching between shortening and extension events. In contrast, the Permian Buchan-type metamorphism documents an important heat input associated with regional NE–SW shortening, probably related to the Early Permian collision between the Chinese Altai and the southerly Junggar arc system.

KEYWORDS

Barrovian metamorphism, Buchan metamorphism, Chinese Altai, monazite U–Pb, P–T–D–t Path

1 | INTRODUCTION

Barrovian-type (also known as middle-pressure or kyanite-type) and Buchan-type (also termed as lowpressure or andalusite-type) metamorphic series are commonly considered as two endmembers of regional metamorphism (Barrow, 1893; Harte & Hudson, 1979; Hudson, 1980). The Barrovian series characterized by a moderate apparent thermal gradient around 20_C/km is considered to reflect crustal thickening, whereas the Buchan series characterized by apparent thermal gradients of 30_C/km or more is interpreted to indicate

DOI: 10.1111/jmg.12647

elevated heat flux in part associated with crustal thinning (e.g. Brown & Johnson, 2018; England & Thompson, 1984; Lyubetskaya & Ague, 2010; Miyashiro, 1994). However, close geographic proximity or even overlap of Barrovian- and Buchan-type metamorphic assemblages has been documented in a number of orogens, for example, from the Variscan Bohemian Massif (Košulicˇov_a & Štípsk_a, 2007), the Central Alps (e.g. Berger et al., 2011), the Grampian Terrane and Southern Highlands of Scotland (e.g. Oliver et al., 2000; Tilley, 1925), the pan-African Kaoko Belt (Will et al., 2004) or the Omineca Belt of British Columbia (Webster & Pattison, 2018). It has been increasingly noted that a simple relationship between metamorphism and tectonics is not always observed.

The Chinese Altai represents a high-grade core of the Central Asian Orogenic Belt (CAOB) where metamorphic zones are grouped into Barrovian- and Buchan-type series (Zhuang, 1994). The former type includes classically biotite, garnet, staurolite and kyanite zones, whereas the latter is characterized by the andalusitesillimanite and, less commonly, garnet–cordierite or cordierite-bearing assemblages (e.g. Wei et al., 2007). Geographically, the Barrovian-type metamorphism is predominantly preserved in the northern part of the Chinese Altai, whereas the Buchan-type is more common in the south (Jiang et al., 2010; Zhuang, 1994). However, such a simple spatial distribution pattern is not definitive, because remnants of Barrovian-type assemblages are also reported from cordierite-bearing (Buchan-type) migmatites in the south (Broussolle et al., 2018; Jiang et al., 2015). Time scales of the development of these two metamorphic series also remain controversial. Numerous metamorphic U–Pb zircon, U–Pb monazite and $^{40}Ar/^{39}Ar$ micas ages fall into Devonian (420–360 Ma) and Permian (290–250 Ma) groups (see summary in Broussolle et al., 2019). Because the Barrovian-type series were overprinted by Buchan-type metamorphism in places (e.g. Wei et al., 2007), some authors have assigned the

DOI: 10.1111/jmg.12647

Devonian ages to the Barrovian-type metamorphism and the Permian ages to the Buchan-type metamorphism (e.g. Li et al., 2014). However, others have shown that in places, seemingly Buchan-type, cordierite-bearing assemblages are associated with ca. 390 Ma metamorphic zircon U–Pb ages (Jiang et al., 2010). Therefore, the main question is whether these two metamorphic series reflect different stages of one metamorphic cycle or two chronologically distinct events. These uncertainties are responsible for a range of contradictory tectonometamorphic models of the Chinese Altai (He et al., 1990; Jiang et al., 2010; Li et al., 2014; Tong et al., 2014; Wang et al., 2009), which cannot be resolved without new approaches involving combined petrological, structural and geochronological investigations.

In order to address this problem, we selected a representative region of the central part of the Chinese Altai, which is known for a juxtaposition of both Barrovian- and Buchan-type metamorphic series (Wei et al., 2007) and a polyphase structural evolution (Jiang et al., 2019; Li, Sun, Rosenbaum, Jiang, et al., 2016). The petrogenetic and temporal relationships of the two metamorphic series have so far not been well constrained because the related metamorphic and structural features have been described in separate studies. In this study, we use detailed microstructural analysis in conjunction with phase equilibria modelling, U– Pb monazite and Lu– Hf garnet geochronological investigations to define a complete **P–T–t–** D path. By doing this, we propose a new geodynamic model describing the tectonometamorphic evolution associated with the Barrovian- and Buchantype metamorphism.

2 | GEOLOGICAL FRAMEWORK

The CAOB represents a Phanerozoic accretionary system located between the Siberian craton in the north and the Tarim– North China cratons in the south (Figure 1). An important part of the system is represented by a Cambro-Ordovician volcano-sedimentary unit called the Altai

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accretionary wedge, extending for more than 2500 km from eastern Kazakhstan via Russia, through north-west China to Mongolia (Figure 1; Jiang et al., 2017). The Altai accretionary wedge was reworked during a Devono-Carboniferous orogenesis that led to significant crustal anatexis and magmatism culminating at 400– 380 Ma (Broussolle et al., 2019; Cai et al., 2011; Hanžl et al., 2016; Huang et al., 2020; Jiang et al., 2016; Wang et al., 2006; Yuan et al., 2007).

The Chinese Altai forms the southern part of the Altai accretionary wedge and is separated in the south from the Devono-Carboniferous Junggar arc system by the NW-trending Erqis Fault (Figure 1b). The Chinese Altai is characterized by migmatite-granite complexes surrounded by the variably metamorphosed Ordovician volcano-sedimentary Habahe Group and low-grade Devonian sedimentary sequences (Broussolle et al., 2019). The Habahe Group consists mainly of quartzo-feldspathic clastic turbidites and pyroclastic rocks with subordinate pelitic interlayers, which are interpreted as active margin sediments (e.g. Long et al., 2008). The Devonian sedimentary sequence consists of weakly to unmetamorphosed pyroclastic and volcanic rocks with bimodal geochemical characteristics (Cui et al., 2020; Wan et al., 2011; Xu et al., 2003).

The Altai accretionary wedge recorded polyphase Palaeozoic deformation, metamorphic and magmatic events leading to transformation of an immature sedimentary wedge into a mature orogenic crust (Huang et al., 2020; Jiang et al., 2016). The main orogenic event in the Chinese Altai is recorded in a subhorizontally foliated deep migmatite-granite unit (e.g. Jiang et al., 2015, 2019; Zhang et al., 2015), which represents anatectic orogenic lower crust derived from the Habahe Group due to extensive Middle Devonian crustal thinning and anataxis (Jiang et al., 2016). This extended crust was later affected by NE– SW-trending upright folding and flow of anatectic orogenic lower crust, assisted by the interplay between upright folding and magma diapirism,

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forming cogenetic migmatite-granite complexes and/or domes in the Middle– Late Devonian (e.g. Wang et al., 2021). The last orthogonal folding event affected mainly the southern Chinese Altai and formed NW– SE-trending upright folds and heterogeneous NW– SEtrending subvertical deformation zones (e.g. Broussolle et al., 2018), in response to the collision between the Chinese Altai and the southerly Junggar arc domain (Guy et al., 2020; Jiang et al., 2019; Li et al., 2017; Li, Sun, Rosenbaum, Cai, et al., 2016). Some of these deformation zones are cored by sillimanite– spinel– orthopyroxene-bearing granulites and garnet– cordierite migmatites dated at 290– 260 Ma (Li et al., 2014; Liu et al., 2020; Tong et al., 2014; Wang et al., 2009), but 400– 390 Ma metamorphic zircons were also preserved (e.g. Broussolle et al., 2018; Jiang et al., 2010).

2.1 | Geology of the study area

This study focuses on a specific area located west of Aletai City (Figure 1b). This region is composed of key lithological and metamorphic sequences of the Chinese Altai varying from structurally deep to shallow levels, consisting of (1) a high-grade Devonian migmatitegranite complex to the west, (2) medium-grade metamorphosed but non-migmatitic Ordovician Habahe Group in the centre and (3) a weakly to unmetamorphosed volcanoclastic sedimentary Devonian succession to the east (Figure 2a). These three rock packages were further defined as roughly orogenic lower crust, orogenic middle crust and orogenic upper crust (Jiang et al., 2019). The Devonian migmatite-granite complex (the anatectic section of the Ordovician Habahe Group) is composed mainly of granites associated with migmatitic paragneiss and rare amphibolite lenses. The granites are represented by medium- to coarse-grained two-mica granites and texturally inhomogeneous schlieren-rich granites. Zircon U– Pb studies of the granites gave ages ranging from 432 to 382 Ma, with a dominant peak at ca. 400 Ma (Jiang et al., 2019; Wang et al., 2006). The

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migmatitic paragneisses correspond to partially molten Habahe Group rocks (Jiang et al., 2019). They are stromatitic to nebulitic, defined by quartz– feldspar _ cordierite-bearing melanosome layers.

The medium-grade metamorphosed Habahe Group consists mainly of amphibolite facies coarse-grained micaschists interlayered with quartzo-feldspathic rocks and quartzites. This sequence preserves a record of overlapping Barrovian- and Buchan-type metamorphism of poorly constrained ages. It is characterized by older middle-pressure/middle-temperature (M**P** /M**T**) staurolite- and kyanite-bearing and younger low-pressure/ high-temperature (L**P** /H**T**) andalusite- and cordierite-bearing metamorphic assemblages (Wei et al., 2007). The timing of Barrovian-type metamorphism has not been determined yet. A metamorphic zircon U– Pb age of 299.2 _ 3.4 Ma was reported from a sillimanite– biotite– plagioclase– quartz schist in a highly deformed ‘ Permian H**T** zone’ along the contact with the Devonian migmatite-granite complex (Figure 2a; Wang et al., 2014) and interpreted as the timing of Buchan-type metamorphism. Monazite Th– U– total Pb isochron ages of 264– 262 Ma reported from similar rocks nearby were also considered as the timing of Buchan-type metamorphism (Zheng et al., 2007).

The Devonian succession overlies the Ordovician Habahe Group with a disconformity and is made up of volcanic and pyroclastic rocks in the lower part (the Kangbutiebao Formation) and siliciclastic sediments, fossiliferous limestones and volcanoclastic rocks in the upper part (the Altai Formation; BGMRX, 1993). Zircons from felsic volcanic rocks document an eruption age of ca. 396 Ma (Jiang et al., 2019), similar to the major age population of schlieren-rich granites from the west. The eruption age is consistent with published geochronological results for volcanic rocks in the Altai Formation from other parts of the Chinese Altai (e.g. Chai et al., 2009).

The sequence of deformation in the study area was

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previously described by Li, Sun, Rosenbaum, Jiang, et al. (2016) and Jiang et al. (2019). The oldest fabric is an amphibolite facies foliation ($S1_B$) with staurolite- and/or kyanite-bearing assemblages in micaschists of the metamorphosed Habahe Group, interpreted as a consequence of crustal thickening (Figure 2b; Jiang et al., 2019). A subhorizontal migmatitic/magmatic foliation ($S1_M$) in the Devonian migmatite-granite complex is geometrically parallel to the $S1_B$. This foliation is considered as a successor of $S1_B$, because the $S1_B$ is crosscut by granites of the same ages as the Devonian migmatite-granite complex (Jiang et al., 2019). The $S1_M$ foliation is commonly associated with extensional lock-up shear bands filled with granitic leucosomes and interpreted as a result of melt-assisted horizontal extension (Jiang et al., 2019). Accordingly, the tectonometamorphic evolution is divided into a burial stage associated with Barrovian metamorphism ($D1_B$– $M1_B$), followed by an extensional stage associated with partial melting ($D1_M$– $M1_M$), responsible for the formation of the structures $S1_B$ and $S1_M$ (Figure 2b). The timing of $D1_M$– $M1_M$ was constrained to ca. 400 Ma by zircon U– Pb ages of the associated S-type granites (Jiang et al., 2019). A sedimentary bedding and subhorizontal greenschist facies metamorphic foliation (up to biotite zone conditions) developed in the Devonian succession and are also considered as reflecting stretching of the orogenic upper crust during the $D1_M$ extension (Jiang et al., 2019).

The subhorizontal structures were heterogeneously affected by crustal-scale NE– SW striking upright F2 folding and were rotated into moderately dipping to subvertical positions, resulting in variable D2 structural features in the three lithological sequences of the region (Figure 2b). In the migmatite-granite complex, the D2 structures are exemplified by large-scale upright F2 folds associated with a steep transposed S2 foliation, whereas in the medium-grade metamorphosed Habahe Group and the Devonian sequence, the outcrop-scale F2 folds are only locally preserved (Figure 2b). Juxtaposition of

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different crustal levels is also taken as the consequence of the F2 upright folding. The juxtaposed metamorphic zones in the studied area are quite narrow, which is due to thinning of crust related to the $D1_M$ extension and further thinning of the F2 limbs through buckling and flattening (Jiang et al., 2019). An abrupt transition from the staurolite zone of the Habahe Group to the biotite zone of the overlying Devonian succession to the south and east (Figure 2a,c) may be explained by an extensional detachment due to thinning of the F2 fold limbs. The exact timing of the D2 episode remains unconstrained, but it occurred soon after $D1_M$ melting as indicated by the observations that $S1_M$ parallel leucosomes are in continuity with the vertical axial planar syn-F2 leucosomes in the migmatite-granite complex (Jiang et al., 2019).

The area was further affected by a large number of NW– SE-trending open to closed upright F3 folds associated with subvertical, generally greenschist facies, axial planar cleavage S3, dipping predominantly to the NE (Figure 2b). As a consequence, the steep oldest fabric S1 (verticalized due to D2 folding) was heterogeneously affected by D3 deformation, resulting in either transposition of S1 into S3 foliation in high-strain zones or just in milimetric D3 crenulations to dozens-of-meters-scale folding of the S1 foliation in low-strain domains (Figure 2a,b). The D3 shortening was also related to important magmatic activity. This activity is characterized by syn-D3 emplacement of numerous pegmatite dykes orthogonal to the XY plane of the D3 strain ellipsoid (i.e. parallel to principal compressive stress), accompanied and/or followed by tight folding and significant fabric transposition (Jiang et al., 2019). In some places, it is also characterized by kilometre-scale syntectonic intrusions of S3-parallel muscovite granites. The timing of D3 shortening was constrained by 280– 273 Ma zircon and monazite U– Pb ages of these pegmatite dykes and muscovite granites (Jiang et al., 2019).

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3 | CRYSTALLIZATION–DEFORMATION RELATIONSHIPS

The overall microstructural features associated with growth of index minerals in the study area were previously investigated in Jiang et al. (2019). In the current study, spatial distributions of mineral isograds and metamorphic zones in the metamorphosed Ordovician Habahe Group and the Devonian succession are further investigated. A map and a block diagram showing the distribution of biotite, garnet, staurolite, kyanite and sillimanite zones and corresponding mineral isograds (Figure 2a,c) are based on study of mineral assemblages from 50 representative samples detailed in Figure S1. The mineral isograds form dome-and-basin isograd interference pattern in the field due to orthogonal D2 and D3 folding processes (Figure 2a– c). Unlike continuous nature of these metamorphic zones, andalusite/ cordierite-bearing regions form spatially discontinuous domains within the metamorphosed Habahe Group (Figure 2a, c). These domains vary in size from a few tens to several hundred meters and are always associated with accumulations of centimetre to tens of meter-wide syn-D3 pegmatite dykes and quartz veins (Figure 2d). These pegmatite dykes and quartz veins locally contain up to 10-cm large crystals of andalusite (Figure 2e). Local replacement of staurolite by quartz pseudomorphs (Figure S2), and growth of large chlorite and muscovite crystals in the surrounding rocks can be observed. The contiguous biotite, garnet, staurolite, kyanite and sillimanite metamorphic zones were previously assigned as Barrovian-type series, whereas the isolated andalusitebearing domains were considered as having a Buchantype metamorphic regime. In order to further address the structural relationship between these two types of metamorphic assemblages, we present a detailed microstructural analysis of four representative samples, followed by a general interpretation of crystallization– deformation relations of the region. Three samples (Grt-St-Ky-And

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micaschist 15CA34, Grt-St-Sil-And micaschist 15CA32 and Grt-St-Sil-And micaschist 16CA52) are collected from low-strain domains, and one sample (Grt-Crd-Chl schist 15CA27) is collected from the D3 high-strain zone.

3.1 | Grt-St-Sil-And micaschist, sample 15CA34

The sample was collected from the kyanite zone (Figure 2a,c). At the outcrop, the steep S1 foliation (rotated from subhorizontal to subvertical attitudes due to D2 deformation) is heterogeneously folded by the F3 folds (Figure 3a). The S1 foliation shows alternation of quartz– feldspar-rich and mica-rich bands, whereas the non-penetrative S3 foliation is defined by the reoriented biotite that is preferentially aligned along the F3 axial planes. The sample is composed of kyanite, staurolite, garnet, andalusite, muscovite, biotite, plagioclase, quartz and ilmenite. Both kyanite and staurolite porphyroblasts, ranging from several millimetres to several centimetres, contain straight inclusion trails of biotite, plagioclase, quartz and rare ilmenite, which are continuous to the external S1 fabric (Figures 3b and 4a– c). Moreover, both kyanite and staurolite porphyroblasts show strong preferred orientations of long axes and are elongated parallel with the surrounding S1 foliation. In places, the external S1 foliation wraps around these porphyroblasts and is locally associated with the development of strain shadows (Figure 4c). Garnet (1– 3 mm) is present inside staurolite porphyroblasts and in the matrix and has quartz and ilmenite inclusions (e.g. Figure 4b). Muscovite and biotite are preferably aligned within the S1 foliation. Andalusite porphyroblasts up to few centimetres long contain inclusion trails of relict S1 foliation and F3 crenulations (Figure 4d). F3 micro-folds are open in the matrix and tight at the contact with the andalusite porphyroblasts (Figure 4e). Large chlorite crystals are adjacent to the andalusite porphyroblasts or occur in the matrix. In places, they overgrow the folded S1 fabric.

3.2 | Grt-St-Sil-And micaschists, samples

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15CA32 and 16CA52

Two samples from the sillimanite zone (Figure 2a,c) are presented. At the outcrop, they have steeply inclined S1 foliation due to F2 folding and weakly refolded by D3, compatible with macrostructures of the kyanite-bearing sample 15CA34. Sample 15CA32 preserves a better record of the garnet– staurolite– sillimanite-bearing assemblage crystallization– deformation relations, whereas sample 16CA52 exhibits a better record of crystallization– deformation relations of andalusite.

In sample 15CA32, the S1 foliation is composed of recrystallized quartz ribbons, biotite, muscovite, garnet, staurolite, sillimanite, ilmenite and rare plagioclase. Garnet (2– 4 mm) is euhedral and has sparse aligned quartz and ilmenite inclusions (Figure 5a). Staurolite porphyroblasts contain straight inclusion trails of ilmenite and quartz, which are continuous to the external S1 foliation (Figure 5b,c). Staurolite, sillimanite and biotite crystals show shape preferred orientation parallel to the S1 foliation (Figure 5a,b). Fibrolitic sillimanite and biotite form aggregates partially replacing staurolite crystals but still oriented parallel to the S1 foliation (Figure 5b). These aggregates were folded during D3 deformation. Randomly distributed andalusite porphyroblasts 2– 4 mm in size overgrow the F3 crenulation (Figure 5d). The textural features related to the growth of andalusite are better developed in sample 16CA52. In this sample, andalusite includes former Barrovian index minerals and the F3 crenulations, or it is aligned with the newly formed S3 foliation (Figure 5e,f). In the outcrop, strain shadows adjacent andalusite related to the formation of penetrative S3 foliation can be locally observed (Figure 5g). In both samples, andalusite porphyroblasts are commonly surrounded by an aggregate of large muscovite, biotite and chlorite crystals. These crystals are in sharp contact with andalusite.

3.3 | Grt-Crd-Chl schist, sample 15CA27

The D3 high-strain zone is characterized by a penetrative

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subvertical S3 foliation defined by recrystallized ribbons of quartz and plagioclase alternating with biotite-rich domains. Relics of large staurolite porphyroblasts are only locally preserved, and less commonly, they contain S1 inclusion trails. One garnet– cordierite– chlorite schist sample 15CA27 was selected for further investigation. The sample contains garnet and cordierite porphyroblasts in a matrix composed of biotite, plagioclase, quartz, chlorite, ilmenite and accessory minerals (Figure 6a– c). Unlike the above two samples, this sample does not contain muscovite and aluminosilicate minerals (Ky/Sil/And). Garnet (2– 4 mm in size) is either without inclusions or with inclusions of quartz and ilmenite close to the margins, oriented parallel to the external S3 fabric (Figure 6a,c). Cordierite porphyroblasts (4– 8 mm in size) contain numerous inclusions of ilmenite, quartz, biotite and chlorite (Figure 6b– d). Cordierite shows complex geometrical relations with respect to the S3 fabric. In some places, cordierite crystals show straight inclusion trails composed of fine-grained quartz, biotite and chlorite in the inner part and coarse-grained inclusions of biotite and quartz along cordierite margins (Figure 6b). The internal mineral inclusion trails in the inner part of the cordierite occur at high angle to the coarse-grained external S3 fabric, whereas the coarse biotite and quartz inclusions in the marginal parts are continuous with the matrix (Figure 6b). In other places, cordierite crystals overgrow the F3 crenulations in the core and the S3 foliation at the rim and are also further wrapped by the same foliation with development of strain shadows (Figure 6e,f). Similar crenulations are also preserved in the matrix at the contact with cordierite and are associated with development of a weak crenulation cleavage S3 (Figure 6e,f). The crenulation axial planes as well as the S3 cleavage are both parallel with the external S3 foliation in the matrix. Apart from chlorite inclusions in the cordierite, large chlorite crystals overgrowing the S3 foliation are locally present and are considered as a result of post-D3 growth.

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3.4 | Interpretation of crystallization–deformation relations

Microstructural analysis indicates shape preferred orientations of staurolite and kyanite porphyroblasts lying within the S1 foliation. These porphyroblasts commonly contain straight inclusions trails that are in continuity with the external S1 foliation (e.g. Figures 3b, 4a,b and 5c), suggesting either post-tectonic or syntectonic growth with respect to the S1 foliation. Despite significant deformation during D2 and D3, some staurolite porphyroblasts are wrapped by the same S1 foliation and exhibit strain shadows (e.g. Figure 4c). These microstructural features are compatible with dynamic growth rather than static crystallization of the porphyroblasts and fit the diagnostic scheme of Zwart (1962) of syntectonic growth. Therefore, we interpret the growth of staurolite and kyanite as synchronous, at least in part, with the D1 deformation. This interpretation is also advocated by the modelled prograde **P–T** evolution of the staurolite– kyanite-bearing S1 fabric, which is in favour of dynamic rather than static growth of these porphyroblasts, as introduced in Wei et al. (2007) as well as in Section 5.

The relative timing of sillimanite growth is more ambiguous. In the S1-parallel sillimanite zone, sillimanite crystals aligned within the S1 foliation, and the sillimanite isograd was rotated into a subvertical orientation together with garnet-staurolite– kyanite isograds during the subsequent D2 evolution (Figure 2a,c). This implies that sillimanite growth predated the D2 episode of deformation. On the other hand, crystallization of sillimanite was associated with breakdown of staurolite (Figure 5a– c), implying that the sillimanite post-dated the main Barrovian-type S1 fabric. In other words, the S1-parallel sillimanite zone most likely formed by sillimanite grade overprinting of the pre-existing staurolite zone at an interval post-dating the Barrovian metamorphism but predating the D2– M2 tectonometamorphic event. Apart from the S1-parallel

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sillimanite aggregates, the sillimanite occurs also parallel to S3 at the contact between the migmatite-granite complex and the metamorphosed Habahe Group. As introduced above, a metamorphic zircon U– Pb age of ca. 299 Ma was interpreted as the timing of sillimanite growth (Wang et al., 2014). In this regard, the sillimanite should be considered as syntectonic with D3, because the age overlaps with the timing of the regional D3 event (Jiang et al., 2019). Alternatively, if the zircon age is taken as the timing of recrystallization during a later thermal perturbation, the S3-parallel sillimanite could be literally considered as the counterpart of S1-parallel sillimanite but reoriented during D3. In these regards, the nature of the S3-parallel sillimanite zone at the contact between the migmatite-granite complex and the metamorphosed Habahe Group remains undetermined and calls for further investigation.

Andalusite porphyroblasts exhibit variable relationships with respect to the heterogeneous and progressive development of the S3 foliation. This is documented in andalusite overgrowths of former Barrovian index minerals, overgrowths of F3 crenulations and S3 cleavage planes and development of the S3 strain shadows around the andalusite porphyroblasts (e.g. Figures 4d and 5e– g). Intensification of the F3 crenulations at the margins of some andalusite grains can also be observed (e.g. Figure 4e). These features suggest that andalusite growth occurred during variable stages of D3 deformation; it started after the beginning of the F3 crenulation, and ended while the D3 deformation still continued, thereby implying syn-D3 crystallization. By contrast, the cordierite shows static overgrowth of the relatively finegrained S1 foliation (Figure 6a), followed by rotation of crystals and by further overgrowth of the F3 crenulations during the D3 (Figure 6e,f). These features suggest that the cordierite started to grow statically and continued its growth during the progression of the D3 deformation.

The growth of cordierite over a progressively coarsened matrix is most likely a consequence of heating. The

DOI: 10.1111/jmg.12647

growth of andalusite and less commonly of cordierite is also spatially related to pegmatite veins (Figure 2e), suggesting an important role of magmatic activity on crystallization of these porphyroblasts. Andalusitebearing quartz veins within the andalusite-bearing metamorphic domains (Figure 2e) suggest circulation of hydrothermal fluids during syn-metamorphic veining, similar to a mechanism described in Cesare (1994). Presence of large idiomorphic crystals of hydrous minerals such as chlorite and muscovite and development of quartz pseudomorphs after staurolite (Figure S2) are also interpreted as a result of fluid circulation and therefore as the effects of hydrothermal activity.

Garnet grains from the kyanite zone (sample 15CA34) and the sillimanite zone (sample 15CA32) contain mineral inclusion trails that are in continuity with the external S1 foliation and show equilibrium texture with the associated staurolite and kyanite (e.g. Figures 3b and 5a). The garnet from sample 15CA27 contains rare inclusion trails that are in continuity with the external S3 foliation and displays equilibrium texture with the associated cordierite (Figure 6a). These features suggest that the garnet grains in samples 15CA34 and 15CA32 grew during D1– M1 and in sample 15CA27 during D3– M3.

4 | MINERAL CHEMISTRY

Three samples (Grt-St-Ky-And micaschist 15CA34, Grt-St-Sil-And micaschist 15CA32 and Grt-Crd-Chl schist 15CA27) were selected for further mineral chemical analysis. Mineral composition analyses and garnet composition mapping of samples 15CA34 and 15CA27 were performed on a JEOL FEG-EPMA JXA-853 electron microprobe at the Institute of Petrology and Structural Geology (IPSG, Charles University in Prague). The compositional analyses were performed in point beam mode at 15 kV and 30 nA with a 5 **μ** m beam diameter and 30 s counting time. Garnet compositional mapping was acquired with 20 kV accelerating voltage and 70 nA beam current (dwell time 40 ms per point) for typically 6– 8 h.

DOI: 10.1111/jmg.12647

The analysis of sample 15CA32 was performed using a JEOL JXA-8100 electron microprobe at the Key Laboratory of Mineralogy and Metallogeny, Guangzhou Institute of Geochemistry, Chinese Academy of Science (GIG-CAS). The operating conditions for compositional analysis were 15 kV accelerating voltage, 20 nA beam current, 3– 5 **μ** m beam diameter and 20 s counting time. The operation conditions for garnet mapping are accelerating voltage of 20 kV, a probe current of 480 nA and a 4 **μ** m beam diameter (dwell time 100 ms per point) for about 4– 6 h. The analyses were calibrated using multiple natural reference materials and were reduced using the ZAF correction routines. The precision of the mineral composition analyses is better than 2% (relative) for most oxides. In addition, x-ray compositional map of staurolite and cordierite was obtained using a Carl Zeiss SUPRA55SAPPHIRE field-emission scanning electron microscope (FE-SEM) at GIG-CAS. The mapping was conducted with a scanning resolution of 1,024 _ 768 pixel in a roughly 2.8 _ 2.1 mm area at 20 kV. The dwell time of each pixel was 300 ms, and the mapping process was operated 6– 8 h per sample.

Representative mineral analyses are summarized in Table 1. Trends in mineral composition or zoning quoted in the text are marked with '!' , and '–' designates a range of mineral composition. Mineral abbreviations are after Whitney and Evans (2010). Calculated endmember proportions and cation ratios are defined as alm = $Fe^{2+}/(Ca + Fe^{2+} + Mg + Mn)$, sps = $Mn/(Ca + Fe^{2+} + Mg + Mn)$, py = $Mg/(Ca + Fe^{2+} + Mg + Mn)$, grs = $Ca/(Ca + Fe^{2+} + Mg + Mn)$, $X_{Fe} = Fe^{2+}/(Fe^{2+} + Mg)$.

4.1 | Grt-St-Ky-And micaschist, sample 15CA34

Garnet porphyroblasts in staurolite and in the matrix are compositionally zoned with an increase in almandine and pyrope and decrease in spessartine and flat grossular from core to rim. Additionally, the matrix garnet shows decreasing pyrope and increasing

DOI: 10.1111/jmg.12647

almandine, spessartine and $\mathbf{X}_{Fe}$ values at the very rim (alm$_{0.62\,!\,0.67\,!\,0.68}$ sps$_{0.20\,!\,0.14\,!\,0.15}$ py$_{0.12\,!\,0.14\,!\,0.12}$ grs$_{0.05–0.06}$, $\mathbf{X}_{Fe}$ = 0.83– 0.85 ! 0.86; Figure 7a). The garnet included in staurolite has slightly narrower almandine, spessartine and pyrope ranges (alm$_{0.64\,!\,0.67}$ sps decreasing Mn intensity from c$_{0.18\,!\,0.14}$ py$_{0.13\,!\,0.14}$ grs$_{0.05–0.06}$, $\mathbf{X}_{Fe}$ = 0.83– 0.84). The matrix garnet x-ray map corroborates decreasing Mn intensity from core to rim (Figure 7a). Staurolite has a constant $\mathbf{X}_{Fe}$ = 0.75– 0.76 in the interior part, and at the very rim, $\mathbf{X}_{Fe}$ reaches 0.79 (Figure 7d), consistent with the Mg x-ray map (Figure 7e). Biotite crystals parallel to the S1 and S3 foliations have the same mineral composition with $\mathbf{X}_{Fe}$ = 0.41– 0.42 and Ti = 0.07– 0.09 p.f.u., whereas biotite in contact with garnet has $\mathbf{X}_{Fe}$ = 0.43 and Ti = 0.09 p.f.u. Plagioclase is not zoned and contains 30%– 34% of anorthite.

4.2 | Grt-St-Sil-And micaschist, sample 15CA32

Garnet shows a weak compositional zoning with slightly increasing $\mathbf{X}_{Fe}$, almandine and pyrope and decreasing spessartine and flat grossular components from core to rim (alm$_{0.61\,!\,0.63}$ sps$_{0.22\,!\,0.20}$ py$_{0.14\,!\,0.15}$ grs$_{\sim 0.03}$, $\mathbf{X}_{Fe}$ = 0.81– 0.82; Figure 7b). At the very rim, spessartine increases to 0.22, whereas pyrope decreases to 0.13 (Figure 7b). The x-ray garnet map corroborates decreasing Mn intensity from core to rim and the increase at the very rim (Figure 7b). Staurolite grains show an increase in $\mathbf{X}_{Fe}$ from core to rim ($\mathbf{X}_{Fe}$ = 0.77– >0.80) (Figure 7d), consistent with the Fe x-ray map (Figure 7f). Matrix biotite parallel to S1 and S3 has the same composition with $\mathbf{X}_{Fe}$ = 0.44– 0.45 and Ti = 0.08– 0.09 p.f.u. Biotite in contact with garnet has similar composition ($\mathbf{X}_{Fe}$ = 0.44 and Ti = 0.09 p.f.u.). Biotite crystals in contact or included in andalusite show slightly different compositional ranges ($\mathbf{X}_{Fe}$ = 0.44– 0.48 and Ti = 0.07– 0.09 p.f.u.). Muscovite in contact with andalusite has X_{Fe} = 0.23 and Si = 3.07 p.f.u.

DOI: 10.1111/jmg.12647

4.3 | Grt-Crd-Chl schist, sample 15CA27

Garnet shows a slight increase in almandine, decrease in spessartine and stable pyrope, grossular and $\mathbf{X}_{Fe}$ values from core to rim (alm$_{0.56\,!\,0.60}$ sps$_{0.20\,!\,0.14}$ py$_{0.18–0.19}$ grs$_{0.06–0.07}$, $\mathbf{X}_{Fe}$ = 0.75– 0.76; Figure 7c). Cordierite porphyroblasts overgrowing the straight S1 inclusion trails have a weak compositional zoning (Figure 7g), characterized by decreasing $\mathbf{X}_{Fe}$ from core to rim ($\mathbf{X}_{Fe}$ = 0.18 ! 0.15) (Figure 7h). Cordierite porphyroblasts overgrowing the F3 crenulation or the S3 foliation have $\mathbf{X}_{Fe}$ = _ 0.17. Coarse biotite included in the rims of cordierite crystals has similar composition as the matrix biotite ($\mathbf{X}_{Fe}$ = 0.31– 0.34 and Ti = 0.05– 0.06 p.f.u.). Chlorite crystals included in cordierite and the matrix chlorite show similar compositional ranges ($\mathbf{X}_{Fe}$ = 0.29– 0.30). Plagioclase is not zoned and has around 38% of anorthite.

5 | METAMORPHIC P–T EVOLUTION

The metamorphic **P–T** evolution of the above three samples were investigated. Given that these samples show diverse microstructural features, different strategies were applied. Samples 15CA34 and 15CA32 preserved S1 assemblages and show heterogeneous overgrowths of D3 andalusite. Their **P–T** evolution with respect to S1 assemblage was investigated using phase equilibria modelling. Instead of using whole rock compositions, effective bulk compositions, that is, compositions of local domains, were applied for the phase diagram modelling, similar to those described in previous studies (e.g. Evans, 2004; Marmo et al., 2002; Zeh, 2006). To approach as closely as possible the bulk composition effective at the scale of the thin section containing the mineral assemblage of interest, whole rock composition used for phase diagram modelling was obtained by quantitative analysis of a continuous and representative area of the thin section. The analysis was conducted by using a Bruker micro-XRF (**µ** -XRF) spectrometer with poly-capillary x-ray optics of

DOI: 10.1111/jmg.12647

the type M4 Tornado equipped with two XFlash® silicon drift x-ray detectors at the Tuoyan Laboratories Ltd in Guangzhou, China, operated at 50 kV and 300 **μ** A. The selected areas were scanned at a pixel resolution of 12 **μ** m using a dwell time of 5 ms per pixel for about 1– 2 h. Semi-quantitative data for major elements were calculated from x-ray intensities via a standard-less model with the software M4 TORNADO and then normalized to 100% (Table S1). Several studies have documented the fractionation effect of zoned garnet porphyroblasts on the changes of effective bulk compositions (e.g. Evans, 2004; Spear, 1988; Zeh, 2006). These authors further suggested an ongoing assessment of effective bulk composition as more appropriate when modelling the **P–T** path associated with the growth of garnet. However, as shown in Zeh (2006), the garnet fractionation influenced the phase diagram topology, but its effect on the resulting interpreted **P–T** path was insignificant. Because the garnets in this study are weakly zoned, their fractionation effect on effective bulk composition was not considered. For the Buchan-type metamorphism in these two samples, the heterogeneous overgrowth of andalusite porphyroblasts makes a quantitative evaluation of effective bulk composition less reliable. Therefore, the **P–T** conditions were estimated using conventional geothermobarometers. The remaining sample 15CA27 shows development of the S3 assemblage associated with pervasive growth of cordierite. Its **P–T** evolution was investigated by phase equilibria modelling using a XRF whole rock composition.

Phase equilibria modelling was conducted by using PerpleX Version 6.9.1 software package (Connolly, 2005) with the upgraded thermodynamic database DS62 from Holland and Powell (2011) in the MnO– Na_2O– CaO– K_2O– FeO– MgO– Al_2O_3– SiO_2– H_2O– TiO_2– Fe_2O_3 (MnNCKFMASHTO) system. The activity– solution TABLE models applied are chlorite, chloritoid, muscovite, biotite, garnet, cordierite, orthopyroxene and melt (White et al., 2014), plagioclase (Newton et al., 1980), epidote

DOI: 10.1111/jmg.12647

(Holland & Powell, 2011) and ilmenite (White et al., 2000). Phase diagrams for all samples were calculated in subsolidus conditions where H_2O was set in excess. Fe_2O_3 (O) was determined using calculated **T–M** (O) pseudosections. Mineral composition and molar isopleths are plotted for the phases of interest in order to determine the **P–T** evolution of the studied samples.

5.1 | Phase equilibria modelling for the Barrovian-type M1 assemblage

5.1.1 | P–T phase diagram for Grt-St-Ky-And micaschist, sample 15CA34

A bulk composition obtained from a 17.9 _ 5.9 mm large continuous area covering typical Barrovian-type S1 fabric, that is, M1 mineral assemblage, in the thin section (Figure S3a) was used for the calculation. The composition is plotted in an AFM diagram in Figure S3a, where it plots close to the average 'high-Al pelites' domain of Spear (1993). The effect of ferric iron was first investigated in a T–**M**(O) phase diagram (not shown). The observed assemblage is stable with O = 0.12–0.20 mol.%, and an average value (O = 0.16 mol.%) was chosen for the construction of **P–T** phase diagram. The M1 assemblage kyanite–staurolite–garnet–biotite–muscovite–plagioclase observed in the S1 foliation is comparable with the stability field of St-Ky-Grt-Bt-Ms-Pl-Qz-H_2O in a **P–T** range of 6.5–9.0 kbar and 640–650_C (Figure 8a). Garnet rim composition (alm$_{0.66–0.67}$sps$_{0.14–0.15}$grs$_{0.05–0.06}$) broadly matches the modelled isopleths within this field at 6.8–7.2 kbar and 645–655_C (the upper right 'circle' in Figure 8b), which is interpreted as the peak **P–T** range of the Barrovian-type M1 assemblage.

The compositions of the garnet core (alm$_{0.62–0.63}$sps$_{0.20}$grs$_{0.05–0.06}$) and the staurolite interior ($\mathbf{X}_{Fe}$ = 0.76) are consistent with the calculated isopleths at 5.6–6.4 kbar and 570–580_C in the stability field of St-Grt-Bt-Ms-Chl-Pl-Qz-H_2O (the lower left 'circle' in Figure 8b,c), which is taken as a **P–T** estimation for the early garnet and staurolite growth. Therefore, a prograde **P–T** path can be inferred, as indicated by the arrow in Figure 8b, compatible with an increase of garnet volume isopleths (Figure 8d). Staurolite volume proportions increase significantly in the chlorite-present field (Figure 8e), implying that growth of staurolite was initially related to continuous chlorite breakdown reactions. By contrast, staurolite volume proportions decrease significantly in the kyanite-present field (Figure 8e), suggesting that kyanite growth would occur at the expense of staurolite along a prograde path. Recent findings suggested that growth of kyanite in typical Barrovian-type pelitic sequences could be approximately contemporaneous with staurolite through chloriteconsuming rather than staurolite-consuming reactions (Pattison & Spear, 2018). However, such view is unlikely the case for the current sample, because the garnet compositional zoning is consistent with a prograde **P–T** evolution through the conditions of the staurolite zone and ending at the conditions of the kyanite zone (Figure 8b). Even if replacement textures of staurolite by kyanite have so far not been observed, kyanite could form at one place, whereas staurolite could dissolve at another place, by a similar mechanism as described in Carmichael (1969).

5.1.2 | P–T phase diagram for Grt-St-Sil-And micaschist, sample 15CA32

The sample shows alternations of garnet–staurolite–sillimanite–biotite-rich and quartz–feldspar-rich layering, which defines the S1 foliation. A representative area of 9.9 _ 7.6 mm covering mainly the biotite–sillimanite-rich layer was selected for the analysis of the effective bulk composition (Figure S3b), which plots in the average ‘high-Al pelites’ domain of Spear (1993) (see AFM diagram in Figure S3b). The Fe_2O_3 amount for the construction of the **P–T** phase diagram was inferred from the T–**M**(O) phase diagram (not shown). The observed peak assemblage (Grt-Sil-Ms-Bt-Pl-Qz-Ilm) is stable

DOI: 10.1111/jmg.12647

with O = 0.02–0.2 mol.%. A nearly average value (O = 0.10 mol.%) was chosen for the **P–T** phase diagram construction, and such value is also comparable with a generally reduced character of the metapelite protoliths. With this value, the calculated **P–T** phase diagram is characterized by sillimanite stability field at high temperature conditions, staurolite stability field at medium temperature conditions and garnet stability starting at low temperature conditions (Figure 9a). Garnet rim composition ($alm_{0.62–0.63}sps_{0.20–0.22}grs_{0.03}$) broadly matches the modelled isopleths at 5.6–6.0 kbar and 660–670_C in the stability field of Sil-Grt-Bt-Ms-Pl-Qz-H_2O, as indicated by a circle in Figure 9b, suggesting equilibration of the peak assemblage under typical sillimanite zone conditions. The $\mathbf{X}_{Fe}$ = 0.77 in staurolite interiors is comparable with the modelled isopleth in the stability fields of St-Grt-Bt-Ms-Pl-Qz-H_2O and St-Grt-Bt-Ms-Chl-Qz-H_2O. Because the staurolite interior in the Grt-St-Ky-And micaschist sample 15CA34 has a similar $\mathbf{X}_{Fe}$ that constrains initial staurolite growth to the stability field of St-Grt-Bt-Ms-Chl-Pl-Qz-H_2O, we assume that staurolite growth in sample 15CA32 could also start at the conditions within the chlorite-bearing stability field ('circle' in Figure 9c). However, to reveal the prograde **P–T** evolution is not straightforward. A weakly zoned spessartine resembles typical prograde garnet zoning in metapelites and therefore is an indicator of originally prograde growth of the garnet, even if affected by significant diffusional homogenization (e.g. Atherton, 1968; Hollister, 1966). However, the overall flat compositional pattern of the garnet (Figure 7b), commonly interpreted as a result of re-equilibration by diffusion at high-grade metamorphic conditions ($\mathbf{T}_{max}$ > 650 C, e.g. Anderson & Olimpio, 1977), prevents a deeper investigation of the prograde **P–T** path. Nevertheless, microstructural observations indicate that staurolite was partly replaced by sillimanite, and, hence, an evolution from the staurolite stability field to the sillimanite stability field ('dashed arrow' in Figure 9d– f) is established, up to the **P–T**

DOI: 10.1111/jmg.12647

conditions inferred previously from the garnet rim composition. Along such a **P–T** path, the volume proportions of staurolite decrease, whereas those of sillimanite and biotite increase, in agreement with the microstructural observations.

5.2 | P–T estimation for the Buchan-type M3 assemblage

In samples 15CA34 and 15CA32, growth of andalusite porphyroblasts is very heterogeneous and probably associated with hydrothermal activity. In this regard, phase equilibrium modelling is not straightforward because the estimation of the equilibrium bulk composition is difficult. An alternative way to estimate **P–T** conditions is the use of conventional geothermobarometers, because they do not require an estimation of the composition of the equilibrium volume (Wu, 2017). The very rim portions of matrix garnets in both samples show significant reverse trends in spessartine and pyrope contents (Figure 7a,b), which suggests re-equilibration of the garnet outermost rim. The biotite in contact with the garnet in both samples commonly has $\mathbf{X}_{Fe}$ values higher compared with biotite farther from garnet, which likely reflects reequilibration with the garnet rim. These compositional features can be most easily explained as a result of M3 Buchan-type metamorphism, because the rocks were thereafter not affected by any further metamorphic overprinting. Therefore, these features are used as the basis for **P–T** estimations of the M3 event. Among the many geothermometers that can be applied to metapelitic rocks, the garnet– biotite (GB) geothermometer of Holdaway (2000) has been suggested as one of the most robust (see summary in Wu & Cheng, 2006). A recent calibration of garnet– biotite– $Al_2 SiO_5$– quartz (GBAQ) geobarometer (Wu, 2017) was proved useful for estimating pressures in particular in CaO-deficient metapelitic rocks. Using the above GB– GBAQ geothermobarometers, the outermost garnet rim and biotite in contact with garnet were combined and

DOI: 10.1111/jmg.12647

gave results of _ 532_ C and 3.0 kbar for sample 15CA34, and _ 595_ C and 3.6 kbar for sample 15CA32. The suggested uncertainties are around _ 30_ C and _ 1.8 kbar (Wu, 2017; Wu & Cheng, 2006). These **P–T** conditions are consistent with the previous estimates of andalusite crystallization in the staurolite (_ 540_ C and 3 kbar) and sillimanite (_ 560_ C and 3 kbar) zone rocks in the study area (Wei et al., 2007).

5.3 | Phase equilibria modelling for Grt-Crd-Chl schist, sample 15CA27

The sample is characterized by a well-developed S3 fabric with negligible relics of the S1 fabric and contains the assemblage garnet, cordierite, biotite, plagioclase and quartz (Figure S4). Small chlorite flakes are present as inclusions in cordierite, whereas large post-tectonic chlorite laths locally occur in the matrix. An initially inferred bulk rock composition from thin section showed unusually low $\mathbf{X}_{Fe}$ and SiO_2 contents compared with samples 15CA34 and 15CA32. Therefore, we performed also an XRF whole rock composition of the sample (Table S1), which showed similarly low $\mathbf{X}_{Fe}$ and SiO_2 contents, and this new composition is used for the **P–T** phase equilibria calculation. The effect of ferric iron content was evaluated in a **T–M** (O) phase diagram, in which the observed assemblage (Grt-Crd-Chl-Bt-Pl-Qz-Ilm) is stable with O = 0.01– 0.2 mol.% (not shown). An average value (O = 0.10 mol.%) was chosen for the **P–T** phase diagram construction, which is also consistent with a generally reduced character of the metapelite protoliths and would allow the discussion of stability of the observed assemblage in the sample in general.

The resulting **P–T** phase diagram (Figure 10a) shows a wide stability field of chlorite, which extends to abnormally high temperature conditions of about 630_ C. Because a possible reason for such extended chlorite stability is the high magnesium content of the rock (e.g. Thompson, 1976), we evaluated its effect on the stability of chlorite in a **T–M** (Mg) phase diagram

DOI: 10.1111/jmg.12647

(Figure 10b). From the diagram, it is apparent that increasing magnesium content in the system increases significantly the upper temperature stability of chlorite.

The observed assemblage can be correlated with the stability field of Grt-Crd-Bt-Chl-Pl-Qz-H_2O in the calculated **P–T** phase diagram (Figure 10a). Coarsening of inclusions from cordierite core to rim suggests increasing temperature during cordierite growth. Occurrence of chlorite crystals in the core of the cordierite and their absence in the external S3 foliation suggest the rocks reached the chlorite-free Grt-Crd-Bt-Pl-Qz-H_2O stability field (Figure 10c). This is consistent with cordierite compositional zoning (Figure 7h), suggesting a prograde **P–T** path starting from modelled $\mathbf{X}_{Fe} =$ 0.18 in the Chl-present stability field of Grt-Crd-Bt-Chl-Pl-Qz-H_2O, to $\mathbf{X}_{Fe} =$ 0.15– 0.16 in the Chl-absent stability field of Grt-Crd-Bt-Pl-Qz-H_2O, as indicated by the arrow in Figure 10c. Garnet core (alm$_{0.56}$ sps$_{0.20}$ grs$_{0.07–0.06}$) and garnet rim (alm$_{0.60}$ sps$_{0.14}$ grs$_{0.06}$) compositions broadly match the modelled isopleths at relatively lower pressure and temperature and higher pressure and temperature, respectively (Figure 10d). Accordingly, a prograde **P–T** path beginning at approximately 3.9– 4.2 kbar and 590– 610_ C and ending at approximately 4.4– 5.0 kbar and 630– 650_ C can be established (the arrow in Figure 10c– e). Along such prograde **P–T** path, molar proportions of garnet are increasing (Figure 10e), supporting prograde garnet growth.

6 | U–PB MONAZITE AND LU–HF GARNET GEOCHRONOLOGY

6.1 | Methods and sample preparation

Y, La, Th and U compositional mapping of monazite was performed on a JEOL JXA-8230 electron microprobe at the Key Laboratory of Mineralogy and Metallogeny, GIG-CAS. The operation conditions of an accelerate voltage of 20 kV, a probe current of 700 nA and a beam size of 1 **μ** m were adopted for mapping. Elemental K**α** line

DOI: 10.1111/jmg.12647

was chosen for all elements during analyses. The dwell time was set to 100 ms for each point. Typically, 4– 6 h was required to accomplish mapping for one grain. In situ monazite U– Pb and simultaneous trace-element analysis were acquired at the State Key Laboratory of Geological Processes and Mineral Resources, China University of Geosciences, Wuhan, using a ArF excimer GeoLas 2005 193 **μ** m connected to an Agilent 7500a ICPMS instrument. The isotopic results were corrected and calculated using ICPMSDataCal (Version 7.0, Liu et al., 2008). The age calculations and concordia plots were done using Isoplot 3.75 (Ludwig, 2003). The dating results and trace elements are given in Table S2. All ages are quoted at the 2s level of uncertainty. Among the above studied samples, only sample 15CA27 contains inclusion-poor clear garnet crystals, which are ideally suited for Lu– Hf garnet geochronology. The Lu– Hf isotopic analyses on this sample were conducted at the Krakow Research Center, Institute of Geological Sciences, Polish Academy of Sciences. Sample digestions, columns chemistry and mass spectrometry procedures are given in Thirlwall and Anczkiewicz (2004) and Anczkiewicz et al. (2004). The dating results are summarized in Table 2. Analytical procedures for U– Pb monazite and Lu-Hf garnet geochronology are available in Appendix S1.

6.2 | U–Pb ages and REE patterns of monazite from the D3 low-strain domain

Samples 15CA34 and 15CA32 from the D3 low-strain domain were selected for in situ monazite U– Pb dating and trace element analysis. These samples show comparable structures that are characterized by well-preserved S1 fabric that was folded but not transposed by D3. The analysis was applied only to the monazite grains included in either staurolite or andalusite porphyroblasts. Staurolite porphyroblasts in these samples show syn-D1 growth, whereas andalusite experienced syn-D3 growth. Monazite grains included in both staurolite and andalusite were

DOI: 10.1111/jmg.12647

analysed in sample 15CA34. In sample 15CA32, only monazites included in staurolite were large enough for analysis, whereas those in andalusite were too small. Instead, we studied a texturally and compositionally similar sample 16CA52 collected from the same metamorphic zone as sample 15CA32 (Figure 2a), as it preserves large monazite in andalusite.

6.2.1 | Monazite in syn-D1 staurolite: Samples 15CA34 and 15CA32

Numerous monazite grains were found in samples 15CA34 and 15CA32, which are distributed in the matrix as well as in staurolite and kyanite porphyroblasts; no monazite was found in garnet, and those in kyanite were not selected for analysis due to small grain size. Seven and four grains included in staurolite porphyroblasts of samples 15CA34 and 15CA32, respectively, were studied. Monazite grains in these two samples exhibit similar morphological features characterized by either subhedral or irregular shape. In x-ray images, most grains do not show significant zoning of Y, except for a few grains that exhibit high Y, U and Th patches (Figure 11a).

Thirty-nine analyses were performed on monazite in sample 15CA34. The majority of data yield concordant ages ranging from 280 to 260 Ma (Figure 11a) with a weighted mean $^{206}Pb/^{238}U$ age of 270 _ 3 Ma (MSWD = 3.8). Several analyses on the high-Y domains yield relatively older ages spreading from 350 to 300 Ma (Figure 11a). All analyses show right-dipping REE distribution patterns with negative Eu anomalies, irrespective of their ages (Figure 11a).

For sample 15CA32, 13 analyses were performed. The resulting data show a similar age pattern as in sample 15CA34. Ten analyses yield similar U– Pb ages falling in a range of 270– 260 Ma, corresponding to a weighted mean $^{206}Pb/^{238}U$ age of 262 _ 3 Ma (MSWD = 0.9; Figure 11b). Two remaining analyses on the high-Y portions give $^{206}Pb/^{238}U$ ages of 309 and 294 Ma, respectively (Figure 11b). Chondrite-normalized REE patterns exhibit

DOI: 10.1111/jmg.12647

also decrease towards the HREE. However, when compared with sample 15CA34, they show weaker negative Eu anomalies and less depleted heavy REE patterns (Figure 11b).

6.2.2 | Monazite in syn-D3 andalusite: Samples 15CA34 and 16CA52

Six and eight analyses were performed on monazite grains included in syn-D3 andalusite from samples 15CA34 and 16CA52, respectively. These grains either sit within the micro-folded S1 fabric or along the S3 fabric (e.g. Figure 11b). Locally, monazite grains occur in clusters of multiple tiny grains (Figure 11a), morphologically similar to hydrothermal monazite or monazite altered by metamorphic fluids (e.g. Schandl & Gorton, 2004; Williams et al., 2011). Compared with monazite included in staurolite, monazite crystals from andalusite porphyroblasts have irregular shape and lack Y zoning (Figure 11a,b).

Six analyses in sample 15CA34 yield consistent $^{206}Pb/^{238}U$ ages ranging from 290 to 270 Ma. These data form a tight cluster of concordant points, giving a weighted mean $^{206}Pb/^{238}U$ age of 279 _ 7 (MSWD = 1.6; Figure 11a). Eight analyses in sample 16CA52 give $^{206}Pb/^{238}U$ ages ranging between 280 and 260 Ma, with a weighted mean $^{206}Pb/^{238}U$ age of 269 _ 5 Ma (MSWD = 1.5; Figure 11b). These analyses have higher heavy REE contents and show relatively weaker Eu negative anomalies compared with analyses from monazite in staurolite, despite their similar ages.

6.3 | Monazite in the D3 high-strain zone

Monazite grains from the garnet– cordierite– chlorite schist, sample 15CA27, were analysed. Monazite grains occur both in the matrix and in cordierite porphyroblasts. Most grains are too small for analysis, but three relatively large grains from cordierite porphyroblasts have been measured. These monazite crystals are all xenomorphic and do not show notable Y zoning but have patchy

DOI: 10.1111/jmg.12647

zoning in Th and U (Figure 11c). Four analyses give nearly equivalent $^{206}Pb/^{238}U$ ages in a range of 270–265 Ma (Figure 11c), forming a single age population with a weighted mean $^{206}Pb/^{238}U$ age of 268 _ 5 Ma (MSWD = 0.26).

6.4 | Lu–Hf garnet geochronology

Lu– Hf isotope dilution analyses were conducted on one whole rock and three garnet fractions derived from garnet– cordierite– chlorite schist (sample 15CA27). The analyses yielded very similar $^{176}Lu/^{177}Hf$ ratios for the garnet fractions, which range between 0.749 and 0.763. This suggests well-mixed aliquots or homogeneous distribution of Lu and Hf in garnet. Concentration of Hf is higher than that in typical metamorphic garnet (e.g. Anczkiewicz et al., 2007), which indicates the presence of some Hf-rich inclusion(s) in garnet fractions. Likely, ilmenite inclusions are enriched in Hf relative to garnet and therefore could contribute to the Hf concentration budget, thus lowering $^{176}Lu/^{177}Hf$ ratios of the garnet dissolutions. Importantly, ilmenite is distributed evenly throughout the garnet and the rest of the rock, forming part of the paragenesis of the sample (Figure 6c). As documented in Scherer et al. (2000), when Hf-rich accessory phases are present in both garnet and the matrix, their effects on Lu– Hf garnet dating accuracy are insignificant. Lu– Hf isochron defined by three garnet fractions and a whole rock yields an age of 261.9 _ 3.4 (MSWD = 3.5; Figure 12). This age is interpreted as reflecting the time of garnet formation.

7 | DISCUSSION

7.1 | Interpretations of geochronological results

The Permian ages obtained mostly ranging from 280 to 260 Ma are the most prominent age population in the study area. A large number of concordant 279– 268 Ma U– Pb ages were obtained from the preferentially aligned S3-parallel monazite grains in the andalusite porphyroblasts, from monazite grains in the micro-folded

DOI: 10.1111/jmg.12647

S1 fabric in andalusite porphyroblasts and from monazite grains in the syn-D3 cordierite porphyroblasts (Figure 11). These data are interpreted as the timing of M3 Buchan-type metamorphism. This interpretation is supported by the fact that growth of the typical Buchantype index minerals (andalusite and cordierite) was connected with the D3 event that was previously constrained at 280– 273 Ma (Jiang et al., 2019).

A large number of monazite inclusions parallel to S1 fabric in the staurolite porphyroblasts also gave predominant 270– 262 Ma ages and minor older ages scattering at 350, 330, 300 and 290 Ma (samples 15CA34 and 15CA32; Figure 11a,b). These ages are much younger than supposed timing of the S1 fabric formation, which should be at least older than 400 Ma (Jiang et al., 2019), implying later equilibration or recrystallization of the monazite. Irrespective of their ages, REE patterns of all these monazite spots exhibit similar distributions with moderate to significant Eu troughs in the chondrite-normalized distribution patterns (Figure 11a,b). These features are different from those of monazite spots in the andalusite porphyroblasts, which are marked by insignificant negative Eu anomalies and systematically higher heavy REE contents (Figure 11a,b). This apparent difference implies that monazite grains included in the M1 staurolite are originally dissimilar from those included in the M3 andalusite. In addition, the analyses giving >290 Ma ages were mainly obtained from the high-Y monazite domains, whereas the analyses with uniform Permian ages were obtained from the low-Y domains (Figure 11a,b).

Although monazite included in porphyroblasts is expected to be protected from later equilibration or recrystallization (e.g. Foster et al., 2004; Martin et al., 2007), it has been documented that monazite inclusions can undergo partial or complete recrystallization when they are connected to the matrix via microcracks (e.g. Hoisch et al., 2008; Martin et al., 2007; Montel et al., 2000). When a fluid is present, such recrystallization,

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for example, by dissolution– precipitation processes, can occur, even at temperatures far below the diffusional closure temperature of the U– Th– Pb system in the monazite (Kelly et al., 2012; Taylor et al., 2014; Williams et al., 2011). In this study, the staurolite texture is poikiloblastic and always contains microcracks (e.g. Figure 4b,c). Such textural features indicate that monazite inclusions were not sufficiently isolated from the matrix and fluids associated with the D3– M3 event could cause recrystallization of monazite. Altogether, the Permian monazite ages in the staurolite porphyroblasts are interpreted to reflect almost complete re-equilibration of U– Th– Pb system during the Buchan-type metamorphic overprint. Likewise, the few >290 Ma dates scattering along the concordia are probably attributed to partial re-equilibration of the U– Th– Pb system in the monazite, and these ages are therefore geologically meaningless.

Lu– Hf garnet– whole rock isotopic data of sample 15CA27 formed a regression line with an isochron age of 261.9 _ 3.4 Ma. This age is identical within the analytical uncertainties to that of monazite included in the cordierite (268 _ 5 Ma; Figure 12) from the same sample. These ages are nearly overlapping with the age populations (270 _ 3, 279 _ 7 Ma, 269 _ 5 and 262 _ 3 Ma; Figure 11a,b) reported from the majority of monazite grains from the D3 low-strain domains. Such an age range is comparable with the zircon ages from the pegmatite dykes in the region, that is, 277– 273 Ma (Jiang et al., 2019). The close spatial relation between arrays of pegmatite dykes and development of Buchan-type andalusite and cordierite domains (Figure 2d,e) attests to close genetic links. Jiang et al. (2019) outlined that the pegmatite dykes emplaced in tensional fractures were considered to form parallel to principal compressive stress (Mode I fractures) of the D3. Emplacement of pegmatite dyke is always thought to connected with propagating hydrofractures (e.g. Mériaux et al., 1999; Weinberg, 1999). The intrusion of magma would enhance wedging-apart of the fracture walls and cause overpressure

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effects at the fracture tips, thus leading to further fracturing (Clemens & Mawer, 1992; Rubin, 1993). This process would also heat, soften and enhance the matrix permeability of the wall rocks in the presence of volatilerich fluids (Connolly et al., 1997). In the study area, we suggest that such a process would account for advecting heat flux to the hosting rocks via emplacement of pegmatite dykes, resulting in widespread growth of andalusite domains (Figure 2d) and growth of new monazite included by the andalusite and cordierite porphyroblasts, and also to nearly complete recrystallization of monazite in staurolite porphyroblasts. The geochronological relationships presented here can be interpreted as a result of progressive deformation and heterogeneous heat and fluid transfer related to emplacement of pegmatites (Thompson & Connolly, 1992). In this regard, the current geochronological work provides constraints on duration and progression of Permian deformation and metamorphism in the region, that is, starting from at ca. 280 Ma and ending at ca. 260 Ma.

7.2 | P–T–D constraints of tectonometamorphic evolution of the Chinese Altai

Microstructural data combined with phase equilibria modelling allow deciphering the **P–T–D** histories of Barrovian- and Buchan-type metamorphism in the region. Data from this study together with available regional data are summarized in Figure 13, and their significance for regional tectonometamorphic evolution is portrayed in Figure 14.

7.2.1 | P–T path of D1–M1

It has been shown above that growth of the index minerals garnet (except for cordierite-bearing schists), staurolite, kyanite and sillimanite is associated with the D1– M1 event. Phase equilibria modelling constrained simultaneous increase of pressure and temperature associated with a Barrovian-type M**P** /M**T** thermal gradient of approximately 23_ C/km for the kyanitebearing

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assemblage in sample 15CA34, in agreement with the **P–T** evolution modelled previously for kyanite zone in the region (see details in Wei et al., 2007). For the staurolite-bearing assemblage in sample 15CA32, its prograde **P–T** evolution could not be currently constrained; however, a **P–T** path beginning at approximately 6– 7 kbar and 560_ C/and ending at approximately 7.3 kbar and 593_ C (with uncertainties of _ 10_ C and kbar) was reported for the staurolite zone in the study area (Wei et al., 2007). If this value is taken into account, a prograde **P–T** path associated with a thermal gradient of approximately 23– 24_ C/km for the staurolite-bearing assemblage can be constructed. In these regards, it can be concluded that the D1– M1 event accounting for the generation of staurolite- and kyanite-bearing is marked by Barrovian-type M**P** /M**T** prograde **P–T** evolution with a maximum pressure of 7.3 _ 1.0 kbar or higher (arrow ' D1B' in Figure 13a). Such a prograde **P–T** evolution is very similar to that deduced from the adjacent staurolite– kyanite-bearing migmatitic complex; the latter reaches a maximum pressure of 8.0 _ 1.0 kbar or higher (arrow ' D1B' in Figure 13b; Jiang et al., 2019). These features indicate that both the orogenic middle crust (e.g. the medium-grade metamorphosed Habahe Group) and the orogenic lower crust (e.g. the migmatite-granite complex) were affected by a Barrovian metamorphism and hence progressive burial during the course of D1B.

The replacement of staurolite by sillimanite-biotite aggregates in sample 15CA32 (Figure 5d) indicates reequilibration of the peak assemblage from the staurolite stability field to a lower pressure and higher temperature sillimanite stability field. Development of sillimanitebearing assemblages would therefore reflect a **P–T** evolution beginning at approximately 7.3 kbar and 593_ C and ending at approximately 5.6– 6.0 kbar and 660– 670_ C, the latter corresponding to a L**P** /H**T** thermal gradient of approximately 34_ C/km. The switch from M**P** /M**T** to L**P** / H**T** thermal gradients is similar to that deduced from structurally deeper migmatites from the adjacent

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regions (Figure 13b; Jiang et al., 2015). There, the relics of staurolite- and/or kyanite-bearing assemblages equilibrated at approximately 8 kbar and 650_ C were overprinted by garnet– K-feldspar– sillimanite-bearing assemblage, corresponding to a re-equilibration at approximately 7.0– 8.0 kbar and 770– 800_ C during D1M (Jiang et al., 2015; see also the arrow ' D1M' in Figure 13b). In addition, the fabric in the deepest orogenic lower crust shows extensional structures (Broussolle et al., 2018; Jiang et al., 2019), which, together with simultaneously formed volcano-sedimentary basins in the orogenic upper crust (Wan et al., 2011), indicate a generalized extensional regime during D1M (Figure 14a). It is therefore most likely that the growth of S1-parallel sillimanite reflects the effects of D1M extension on the orogenic middle crust leading to partial L**P** /H**T** reequilibration of the former Barrovian-type assemblage (the arrow ' D1M' in Figure 13a).

7.2.2 | P–T path of D2–M2

The **P–T** evolution related to the D2– M2 tectonometamorphic event is not constrained because no M2 metamorphic assemblages were observed in the investigated samples. However, garnet, staurolite, kyanite and sillimanite schists are sharply juxtaposed with the neighbouring biotite schists of the Devonian sequence prior to the F3 folding (Figure 2a,b), suggesting that the orogenic middle crust was exhumed to shallower levels already during the D2 upright folding (Figure 14a; see also Jiang et al., 2019). Such large-scale upright folding described in other parts of the Altai was responsible for the extrusion of deep-seated orogenic lower crust in the cores of gneiss domes (Broussolle et al., 2015; Jiang et al., 2015; Lehmann et al., 2017; Zhang et al., 2015). The biotite schist has a peak **P–T** condition of approximately 4 _ 1 kbar and 500 _ 10_ C (Wei et al., 2007). It is hence most likely that the kyanite and sillimanite schists might probably at least pass to 4 kbar and 500_ C and shared exhumation with the biotite schist prior to their juxtaposition. In other words,

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the studied kyanite and sillimanite schists were probably exhumed from approximately 20 km depth (**P** = 6– 5 kbar) to at least 10 km depth (**P** = 4– 2 kbar) during the D2 event (the arrow ' D2' in Figure 13a). The preservation of kyanite- and sillimanite-bearing assemblages during the D2 exhumation was probably due to passive elevation of deep crustal rocks without deformation and in the absence of aqueous fluids (Wei et al., 2007). The **P–T** evolution related to the D2– M2 event has been constrained for the orogenic lower crust where K-feldspar– sillimanite-bearing assemblage was reequilibrated in the cordierite stability field, as exemplified by the preservation of this mineral in the F2 planar leucosomes (e.g. Jiang et al., 2015; Wei et al., 2007; see also Figure 13b). **P–T** phase equilibria modelling revealed that this evolution was associated with an exhumation path from 8 to 9 kbar to at least 4– 5 kbar (arrow ' D2' in Figure 13b; Jiang et al., 2015; Wei et al., 2007). Altogether, the D2– M2 event is characterized by a significant decompression from different pressure peaks to 2– 4 kbar. This is in agreement with the model of growth of a hot migmatite-granite dome responsible for vertical transposition, elevation and thinning of originally subhorizontal metamorphic isograds (Figure 14a).

7.2.3 | P–T path of D3–M3

The D3 folding is responsible for the development of biotite- and muscovite-bearing S3 foliation zones, suggesting overall biotite-zone regional metamorphic conditions. However, the S3 was locally associated with formation of low-pressure andalusite- and cordieritebearing assemblages in the metamorphosed Habahe Group. As it has been suggested above, growth of andalusite in the D3 low-strain domains was developed at approximately 3– 4 kbar and 530– 600_ C, corresponding to a thermal gradients of 48_ C/km or more (e.g. samples 15CA34 and 15CA32, arrow ' D3' in Figure 13c), and growth of cordierite in the D3 high-strain zones was associated with a prograde **P–T** path ending at approximately

DOI: 10.1111/jmg.12647

4.3– 5.0 kbar and 630– 650_ C, corresponding to a thermal gradient of approximately 41_ C/km (e.g. sample 15CA27, the arrow ‘ D3’ in Figure 13d). The prominent feature of the M3 **P–T** evolution is associated with abnormally high thermal gradients, implying local additional input of heat. The **P–T** path developed in pressure conditions of approximately 4.0– 5.0 kbar, within the uncertainties of the **P–T** estimations, suggesting M3 might have been nearly isobaric heating or even was associated with slight burial or exhumation during D3 shortening.

Localized heat supply in the region is indicated by close spatial and temporal relationships between lowpressure assemblages and intrusions of pegmatite dykes. This is best exemplified by the deformation– crystallization relationships and geochronology of sample 15CA27 in a D3 high-strain zone, where growth of cordierite and garnet porphyroblasts and emplacement of pegmatite dykes are dated at ca. 268– 262 Ma (Figures 11 and 12) and 277– 273 Ma (Jiang et al., 2019), respectively. The first static overgrowth of the S1 by the cordierite porphyroblasts (e.g. Figure 6a) may reflect input of heat advected from incipient intrusion of pegmatite dykes (‘ Stage 1’ in Figure 13d). Ongoing intrusions of pegmatite dykes provided further heat, enhancing weakening and folding of S1 related with the sequential growth of the cordierite (e.g. Figure 6e,f) and garnet porphyroblasts during the D3 transposition (‘ Stage 2’ in Figure 13d).

7.3 | Barrovian- and Buchan-type metamorphism: Proxies for tectonic switching

In the Chinese Altai, the earliest Barrovian metamorphism of presumably Late Silurian to Early Devonian age is characterized by regularly spaced metamorphic zones and increasing metamorphic grade with depth (e.g. Jiang et al., 2015; Wei et al., 2007; see also Figure 14a). This metamorphic phase results in homogeneous but moderate crustal thickening to approximately 30– 35 km along **MP** /HT thermal gradients of 20– 25_ C/km (Figure 14c).

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Similar Barrovian metamorphism was also reported from the neighbouring Mongolian Altai (e.g. Burenjargal et al., 2014; Zorigtkhuu et al., 2011), suggesting that the whole Altai wedge was first affected by such burial.

Replacement of the Barrovian-type assemblages by L**P** / H**T** ones during D1$_M$ is the hallmark of a significant switch of thermal and tectonic regimes affecting the whole Altai wedge crust during Middle Devonian. At depth, the bottom of the thickened orogenic lower crust was partially molten at low pressure (Figure 14a; Hanžl et al., 2016; Jiang et al., 2015; Zhang et al., 2015), leaving behind a garnet– orthopyroxene granulite residuum (Jiang et al., 2016; Kozakov et al., 2002). The orogenic middle crust was decompressed during this event as shown by partial re-equilibration of the Barrovian-type assemblages in the sillimanite stability field (sample 15CA32; Figures 13a and 14a). At the same time, extrusion of bimodal volcanic rocks occurs in Devonian extensional basins (e.g. Soejono et al., 2018; Wan et al., 2011).

This major thermal event is typical for large heat input from the mantle associated with thinning and horizontal stretching of the whole lithosphere (Jiang et al., 2016).

The second episode of deformation (D2) is a decompression process that is related to renewed shortening, leading to growth of large migmatite-granite domes during Middle– Late Devonian. During this event, the hot orogenic lower and middle crust were exhumed to shallow crustal levels, forming the domes (Figure 14a). In the Chinese Altai, this deformation episode is characterized by formation of L**P** /H**T** cordierite-bearing migmatites in the core of domes (e.g. Jiang et al., 2015, see also Figure 14c), whereas, in the neighbouring Mongolian Altai, it led to a Buchan metamorphism marked by growth of andalusite and cordierite in the metamorphic envelopes of the domes (Broussolle et al., 2015; Lehmann et al., 2017).

Jiang et al. (2019) showed that the succession of Devonian tectonometamorphic events affecting the Chinese Altai can be explained by the tectonic switching

DOI: 10.1111/jmg.12647

model of Collins (2002), which is characterized by alternation between shortening and extensional deformation. In theory, such tectonic switches would lead to formation of multiple short-lived burial– exhumation cycles in subduction orogens and result in multiple burial– exhumation **P–T** cycles (Beltrando et al., 2007). We argue that the succession of short-lived Silurian– Devonian M**P** / M**T** and Middle Devonian L**P** /H**T** metamorphic cycles in the Chinese Altai may define the tectonic switching characterized by shortening followed by extension and renewed shortening (Figure 14a). These metamorphic and tectonic cycles can be considered as proxies of alternating retreats and advances of the subduction slab of the Paleo-Pacific Ocean (Jiang et al., 2017), which is also typical for accretionary orogenic systems (Collins, 2002).

The most spectacular Buchan-type metamorphism in the region is related to the Permian D3 heterogeneous reworking of the whole southern Chinese Altai (Broussolle et al., 2019). This event is constrained here to have lasted mainly from 280 to 260 Ma. This highly heterogeneous metamorphic event is related to nearly isobaric heating. Locally, ultra-high-temperature, lowpressure granulite facies metamorphism develops in narrow NE-trending tabular extrusion zones (e.g. Broussolle et al., 2018; Liu et al., 2020; Tong et al., 2014; Wang et al., 2009, see also Figure 15b,c). This important Buchan-type metamorphic cycle is probably connected with the coeval collision between the Chinese Altai with the southerly Junggar arc domain (Guy et al., 2020; Jiang et al., 2019). The collisional event is considered as the consequence of an interplay of large-scale oroclinal bending, exemplified by the Mongolian Orocline (e.g. Guy et al., 2020), formation of narrow E– W trending alkaline magmatic provinces that are axial planar to the orocline (e.g. Kovalenko et al., 2004) and closure of oceanic basins further south (e.g. Xiao et al., 2015). The heat source of this Buchan-type metamorphism is most likely linked with narrow linear thermal perturbations of the mantle related to the above-mentioned processes. This is

DOI: 10.1111/jmg.12647

consistent with recent findings that a short-lived Early Permian extension existed during the collision cited above (Li et al., 2015, 2017), which might provide heat source for the high-temperature Buchan-type metamorphism in the southern Chinese Altai.

8 | CONCLUSIONS

We undertook a multidisciplinary investigation that allowed the tectonometamorphic evolution of spatially overlapping Barrovian- and Buchan-type metamorphic cycles in the Chinese Altai to be defined, summarized as follows:

1. Barrovian-type metamorphic series including garnet, staurolite and kyanite zones developed synchronously with the regional $S1_B$ foliation, followed by partial reequilibration of the staurolite zone, resulting in formation of a sillimanite-bearing $S1_M$ foliation, still in parallelism with $S1_B$. The $S1_{B-M}$ foliations were subsequently folded by the D2 upright folds and finally refolded orthogonally by the D3 folds. In contrast, the Buchan-type metamorphic series are represented by localized overgrowth of andalusite- and cordieritebearing assemblages associated with D3 deformation.

2. Barrovian-type staurolite– kyanite sequence shows a prograde **P–T** evolution along M**P** /M**T** thermal gradients. This event is followed by high-temperature partial re-equilibration in the sillimanite stability field under L**P** /H**T** thermal regime and further important decompression without apparent re-equilibration. This sequence of metamorphism corresponds to syn-$D1_B$ burial and thickening, syn-$D1_M$ decompression and heating, followed by D2 compressive doming and associated exhumation. The Buchan-type metamorphism is related to syn-D3 heating associated with a renewed L**P** /H**T** thermal regime.

3. U– Th– Pb system in monazite preserved as inclusions in the Barrovian index minerals was nearly completely re-equilibrated, and monazites in the main Buchan-type assemblage yield ages of 280– 260 Ma.

DOI: 10.1111/jmg.12647

These data together with Lu– Hf garnet– whole rock age of ca. 262 Ma constrain the duration of Permian Buchan-type metamorphism.

4. The sequence of tectonometamorphic events in the Chinese Altai reflects the Devonian D1– D2 coaxial suprasubduction sequence of tectonic switching, whereas the D3 shortening and main Permian Buchan-type metamorphism reflects complex deformation and heat transfer process related to Early Permian collision between the Chinese Altai and its southerly Junggar arc domain.

ACKNOWLEDGEMENTS

Roberto Weinberg is thanked for his valuable comments on an earlier version of this manuscript. Martin Racek, Linli Chen and Changming Xing are thanked for their assistance with the collection of EPMA data. This study was supported by the National Natural Science Foundation of China (42022017, 41872222 and 42021002), the International Partnership Program of CAS (132744KYSB20190039) and the Guangdong Basic and Applied Basic Research Foundation (2019A1515012190). Contribution of K.S., P.Š . and C.A. was funded by Czech Science Foundation (EXPRO 19-27682X to K.S.). The Thousand Youth Talents Plan and a project from Government of Guangdong Province (2019QN01H101) to P.F.L. are also acknowledged. Constructive reviews from D.R.M. Pattison, F. George and an anonymous reviewer and editorial suggestions from Bernardo Cesare resulted in significant improvements to the paper.

DATA AVAILABILITY STATEMENT

The data that supports the findings of this study are available in the supplementary material of this article

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