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1 **Barrovian and Buchan metamorphic series in the Chinese Altai:**
2 **P–T–t–D evolution and tectonic implications**

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13 *Running title: P–T–t–D evolution of the Altai Orogen*

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16 **Abstract**

17 The relations between Barrovian- and Buchan-type metamorphic series in the
18 Chinese Altai remain obscure, and hence a representative region of the central
19 part of the Chinese Altai was investigated to address this issue, using combined
20 microstructural, petrological and geochronological methods. In the
21 region, Barrovian-type garnet, staurolite, kyanite and sillimanite zones are
22 locally overprinted by Buchan-type andalusite- and cordierite-bearing
23 domains. Microstructural analysis shows that Barrovian garnet, staurolite and
24 kyanite grew synchronously with the earliest regional metamorphic foliation
25 S_{1B}. A sillimanite-bearing assemblage locally overprinted the assemblage of
26 the staurolite zone in a foliation parallel with S_{1B}, assigned as S_{1M}. The originally
27 subhorizontal S_{1B-M} foliation, metamorphic zones and mineral isograds
28 were folded by F₂ upright folds, leading to their inclination and juxtaposition
29 to upper crustal levels. Subsequent D₃ deformation affected heterogeneously
30 all previous structures producing vertical high-strain zones around low-strain
31 domains. The D₃ high-strain zones in the vicinity of Permian pegmatites are
32 associated with Buchan-type metamorphism and are characterized by syn-D₃
33 growth of andalusite and cordierite. Phase equilibria modelling of the staurolite/
34 kyanite-bearing assemblages suggests a prograde P–T path with an apparent

35 thermal gradient of ~ 23 °C/km associated with the S1_B fabric. The partial
36 re-equilibration occurred in the sillimanite stability at ~ 670 °C, corresponding
37 to an apparent thermal gradient of ~ 34 °C/km in the S1_M fabric. Garnet to
38 sillimanite metamorphic zones were subsequently exhumed without apparent
39 re-equilibration during the D2 event. The interpreted pressure–temperature
40 (P–T) evolution of the Buchan-type metamorphism on the basis of
41 thermobarometry and phase equilibria modelling suggests significant heating
42 processes, corresponding to apparent thermal gradient of 41 °C/km or more. In
43 situ U–Pb dating of monazite inclusions in staurolite revealed predominantly
44 280–260 Ma ages and minor older ages scattering between 350 and 290 Ma,
45 interpreted as important monazite recrystallization during the D3 event. Monazite
46 in andalusite and cordierite yielded only ages of 280–260 Ma, interpreted
47 as dating the growth of these minerals. Lu–Hf garnet–whole rock isochron of a
48 garnet– cordierite– chlorite schist gave an age of ca. 262 Ma, overlapping in
49 time with the age of monazite in the cordierite. Combined with available
50 regional data, the results suggest that the Barrovian-type metamorphic cycle
51 reflects a continued burial heating followed by decompression, probably
52 connected with the Devonian suprasubduction tectonic switching between
53 shortening and extension events. In contrast, the Permian Buchan-type metamorphism
54 documents an important heat input associated with regional NE–
55 SW shortening, probably related to the Early Permian collision between the
56 Chinese Altai and the southerly Junggar arc system.

57

58

59 **KEYWORDS**

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

61

62

63 **1 | INTRODUCTION**

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

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

89 The Chinese Altai represents a high-grade core of the
90 Central Asian Orogenic Belt (CAOB) where metamorphic
91 zones are grouped into Barrovian- and Buchan-type
92 series (Zhuang, 1994). The former type includes
93 classically biotite, garnet, staurolite and kyanite zones,
94 whereas the latter is characterized by the andalusitesillimanite
95 and, less commonly, garnet–cordierite or
96 cordierite-bearing assemblages (e.g. Wei et al., 2007).

97 Geographically, the Barrovian-type metamorphism is
98 predominantly preserved in the northern part of the
99 Chinese Altai, whereas the Buchan-type is more common
100 in the south (Jiang et al., 2010; Zhuang, 1994). However,
101 such a simple spatial distribution pattern is not definitive,
102 because remnants of Barrovian-type assemblages are
103 also reported from cordierite-bearing (Buchan-type)
104 migmatites in the south (Broussolle et al., 2018; Jiang
105 et al., 2015). Time scales of the development of these two
106 metamorphic series also remain controversial. Numerous
107 metamorphic U–Pb zircon, U–Pb monazite and ⁴⁰Ar/³⁹Ar
108 micas ages fall into Devonian (420–360 Ma) and Permian
109 (290–250 Ma) groups (see summary in Broussolle
110 et al., 2019). Because the Barrovian-type series were
111 overprinted by Buchan-type metamorphism in places
112 (e.g. Wei et al., 2007), some authors have assigned the

113 Devonian ages to the Barrovian-type metamorphism and
114 the Permian ages to the Buchan-type metamorphism
115 (e.g. Li et al., 2014). However, others have shown that
116 in places, seemingly Buchan-type, cordierite-bearing
117 assemblages are associated with ca. 390 Ma metamorphic
118 zircon U–Pb ages (Jiang et al., 2010). Therefore, the main
119 question is whether these two metamorphic series reflect
120 different stages of one metamorphic cycle or two chronologically
121 distinct events. These uncertainties are responsible
122 for a range of contradictory tectonometamorphic
123 models of the Chinese Altai (He et al., 1990; Jiang
124 et al., 2010; Li et al., 2014; Tong et al., 2014; Wang
125 et al., 2009), which cannot be resolved without new
126 approaches involving combined petrological, structural
127 and geochronological investigations.
128 In order to address this problem, we selected a
129 representative region of the central part of the Chinese
130 Altai, which is known for a juxtaposition of both
131 Barrovian- and Buchan-type metamorphic series (Wei
132 et al., 2007) and a polyphase structural evolution (Jiang
133 et al., 2019; Li, Sun, Rosenbaum, Jiang, et al., 2016).
134 The petrogenetic and temporal relationships of the two metamorphic
135 series have so far not been well constrained
136 because the related metamorphic and structural features
137 have been described in separate studies. In this study, we
138 use detailed microstructural analysis in conjunction with
139 phase equilibria modelling, U–Pb monazite and Lu–Hf
140 garnet geochronological investigations to define a
141 complete P – T – t – D path. By doing this, we propose a new
142 geodynamic model describing the tectonometamorphic
143 evolution associated with the Barrovian- and Buchan-type
144 metamorphism.

145 **2 | GEOLOGICAL FRAMEWORK**

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

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

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

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

190 forming cogenetic migmatite-granite complexes and/or
191 domes in the Middle– Late Devonian (e.g. Wang
192 et al., 2021). The last orthogonal folding event affected
193 mainly the southern Chinese Altai and formed NW–
194 SE-trending upright folds and heterogeneous NW– SEtrending
195 subvertical deformation zones (e.g. Broussolle
196 et al., 2018), in response to the collision between the
197 Chinese Altai and the southerly Junggar arc domain
198 (Guy et al., 2020; Jiang et al., 2019; Li et al., 2017; Li,
199 Sun, Rosenbaum, Cai, et al., 2016). Some of these
200 deformation zones are cored by sillimanite– spinel–
201 orthopyroxene-bearing granulites and garnet– cordierite
202 migmatites dated at 290– 260 Ma (Li et al., 2014; Liu
203 et al., 2020; Tong et al., 2014; Wang et al., 2009), but
204 400– 390 Ma metamorphic zircons were also preserved
205 (e.g. Broussolle et al., 2018; Jiang et al., 2010).

206 2.1 | Geology of the study area

207 This study focuses on a specific area located west of
208 Aletai City (Figure 1b). This region is composed of key
209 lithological and metamorphic sequences of the Chinese
210 Altai varying from structurally deep to shallow levels,
211 consisting of (1) a high-grade Devonian migmatitegranite
212 complex to the west, (2) medium-grade metamorphosed
213 but non-migmatitic Ordovician Habahe Group in
214 the centre and (3) a weakly to unmetamorphosed
215 volcanoclastic sedimentary Devonian succession to the
216 east (Figure 2a). These three rock packages were further
217 defined as roughly orogenic lower crust, orogenic middle
218 crust and orogenic upper crust (Jiang et al., 2019).

219 The Devonian migmatite-granite complex (the
220 anatectic section of the Ordovician Habahe Group) is
221 composed mainly of granites associated with migmatitic
222 paragneiss and rare amphibolite lenses. The granites are
223 represented by medium- to coarse-grained two-mica
224 granites and texturally inhomogeneous schlieren-rich
225 granites. Zircon U– Pb studies of the granites gave ages
226 ranging from 432 to 382 Ma, with a dominant peak at
227 ca. 400 Ma (Jiang et al., 2019; Wang et al., 2006). The

228 migmatitic paragneisses correspond to partially molten
229 Habahe Group rocks (Jiang et al., 2019). They
230 are stromatitic to nebulitic, defined by quartz–
231 feldspar _ cordierite-bearing melanosome layers.
232 The medium-grade metamorphosed Habahe Group
233 consists mainly of amphibolite facies coarse-grained
234 micaschists interlayered with quartzo-feldspathic rocks
235 and quartzites. This sequence preserves a record of
236 overlapping Barrovian- and Buchan-type metamorphism
237 of poorly constrained ages. It is characterized by
238 older middle-pressure/middle-temperature (MP /MT)
239 staurolite- and kyanite-bearing and younger low-pressure/
240 high-temperature (LP /HT) andalusite- and
241 cordierite-bearing metamorphic assemblages (Wei
242 et al., 2007). The timing of Barrovian-type metamorphism
243 has not been determined yet. A metamorphic zircon
244 U– Pb age of 299.2 _ 3.4 Ma was reported from a
245 sillimanite– biotite– plagioclase– quartz schist in a highly
246 deformed ‘ Permian HT zone’ along the contact with the
247 Devonian migmatite-granite complex (Figure 2a; Wang
248 et al., 2014) and interpreted as the timing of Buchan-type
249 metamorphism. Monazite Th– U– total Pb isochron ages
250 of 264– 262 Ma reported from similar rocks nearby were
251 also considered as the timing of Buchan-type metamorphism
252 (Zheng et al., 2007).
253 The Devonian succession overlies the Ordovician
254 Habahe Group with a disconformity and is made up of
255 volcanic and pyroclastic rocks in the lower part (the
256 Kangbutiebao Formation) and siliciclastic sediments,
257 fossiliferous limestones and volcanoclastic rocks in the
258 upper part (the Altai Formation; BGMRX, 1993). Zircons
259 from felsic volcanic rocks document an eruption age of
260 ca. 396 Ma (Jiang et al., 2019), similar to the major
261 age population of schlieren-rich granites from the
262 west. The eruption age is consistent with published
263 geochronological results for volcanic rocks in the Altai
264 Formation from other parts of the Chinese Altai
265 (e.g. Chai et al., 2009).
266 The sequence of deformation in the study area was

267 previously described by Li, Sun, Rosenbaum, Jiang,
268 et al. (2016) and Jiang et al. (2019). The oldest fabric is an
269 amphibolite facies foliation (S_{1B}) with staurolite- and/or
270 kyanite-bearing assemblages in micaschists of the metamorphosed
271 Habahe Group, interpreted as a consequence
272 of crustal thickening (Figure 2b; Jiang et al., 2019). A
273 subhorizontal migmatitic/magmatic foliation (S_{1M}) in the
274 Devonian migmatite-granite complex is geometrically
275 parallel to the S_{1B} . This foliation is considered as a successor
276 of S_{1B} , because the S_{1B} is crosscut by granites of
277 the same ages as the Devonian migmatite-granite complex
278 (Jiang et al., 2019). The S_{1M} foliation is commonly
279 associated with extensional lock-up shear bands filled
280 with granitic leucosomes and interpreted as a result of
281 melt-assisted horizontal extension (Jiang et al., 2019).
282 Accordingly, the tectonometamorphic evolution is
283 divided into a burial stage associated with Barrovian
284 metamorphism (D_{1B} – M_{1B}), followed by an extensional
285 stage associated with partial melting (D_{1M} – M_{1M}), responsible
286 for the formation of the structures S_{1B} and S_{1M}
287 (Figure 2b). The timing of D_{1M} – M_{1M} was constrained to
288 ca. 400 Ma by zircon U–Pb ages of the associated S-type
289 granites (Jiang et al., 2019). A sedimentary bedding and
290 subhorizontal greenschist facies metamorphic foliation
291 (up to biotite zone conditions) developed in the Devonian
292 succession and are also considered as reflecting
293 stretching of the orogenic upper crust during the D_{1M}
294 extension (Jiang et al., 2019).
295 The subhorizontal structures were heterogeneously
296 affected by crustal-scale NE–SW striking upright F2 folding
297 and were rotated into moderately dipping to subvertical
298 positions, resulting in variable D2 structural
299 features in the three lithological sequences of the region
300 (Figure 2b). In the migmatite-granite complex, the D2
301 structures are exemplified by large-scale upright F2 folds
302 associated with a steep transposed S_2 foliation, whereas
303 in the medium-grade metamorphosed Habahe Group
304 and the Devonian sequence, the outcrop-scale F2 folds
305 are only locally preserved (Figure 2b). Juxtaposition of

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

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

344 **3 | CRYSTALLIZATION–**345 **DEFORMATION RELATIONSHIPS**

346 The overall microstructural features associated with
347 growth of index minerals in the study area were previously
348 investigated in Jiang et al. (2019). In the current
349 study, spatial distributions of mineral isograds and
350 metamorphic zones in the metamorphosed Ordovician
351 Habahe Group and the Devonian succession are further
352 investigated. A map and a block diagram showing the
353 distribution of biotite, garnet, staurolite, kyanite and sillimanite
354 zones and corresponding mineral isograds
355 (Figure 2a,c) are based on study of mineral assemblages
356 from 50 representative samples detailed in Figure S1. The
357 mineral isograds form dome-and-basin isograd interference
358 pattern in the field due to orthogonal D2 and D3
359 folding processes (Figure 2a– c). Unlike continuous
360 nature of these metamorphic zones, andalusite/
361 cordierite-bearing regions form spatially discontinuous
362 domains within the metamorphosed Habahe Group
363 (Figure 2a, c). These domains vary in size from a few tens
364 to several hundred meters and are always associated with
365 accumulations of centimetre to tens of meter-wide syn-
366 D3 pegmatite dykes and quartz veins (Figure 2d). These
367 pegmatite dykes and quartz veins locally contain up to
368 10-cm large crystals of andalusite (Figure 2e). Local
369 replacement of staurolite by quartz pseudomorphs
370 (Figure S2), and growth of large chlorite and muscovite
371 crystals in the surrounding rocks can be observed.
372 The contiguous biotite, garnet, staurolite, kyanite and
373 sillimanite metamorphic zones were previously assigned
374 as Barrovian-type series, whereas the isolated andalusitebearing
375 domains were considered as having a Buchantype
376 metamorphic regime. In order to further address the
377 structural relationship between these two types of metamorphic
378 assemblages, we present a detailed microstructural
379 analysis of four representative samples, followed by
380 a general interpretation of crystallization– deformation
381 relations of the region. Three samples (Grt-St-Ky-And

382 micaschist 15CA34, Grt-St-Sil-And micaschist 15CA32
383 and Grt-St-Sil-And micaschist 16CA52) are collected from
384 low-strain domains, and one sample (Grt-Crd-Chl schist
385 15CA27) is collected from the D3 high-strain zone.

386 **3.1 | *Grt-St-Sil-And micaschist, sample*** 387 ***15CA34***

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

419 **3.2 | *Grt-St-Sil-And micaschists, samples***

420 15CA32 and 16CA52

421 Two samples from the sillimanite zone (Figure 2a,c) are
422 presented. At the outcrop, they have steeply inclined S1
423 foliation due to F2 folding and weakly refolded by D3,
424 compatible with macrostructures of the kyanite-bearing
425 sample 15CA34. Sample 15CA32 preserves a better record
426 of the garnet– staurolite– sillimanite-bearing assemblage
427 crystallization– deformation relations, whereas sample
428 16CA52 exhibits a better record of crystallization–
429 deformation relations of andalusite.

430 In sample 15CA32, the S1 foliation is composed of
431 recrystallized quartz ribbons, biotite, muscovite, garnet,
432 staurolite, sillimanite, ilmenite and rare plagioclase. Garnet
433 (2– 4 mm) is euhedral and has sparse aligned quartz
434 and ilmenite inclusions (Figure 5a). Staurolite
435 porphyroblasts contain straight inclusion trails of ilmenite
436 and quartz, which are continuous to the external S1
437 foliation (Figure 5b,c). Staurolite, sillimanite and biotite
438 crystals show shape preferred orientation parallel to the
439 S1 foliation (Figure 5a,b). Fibrolitic sillimanite and biotite
440 form aggregates partially replacing staurolite crystals but
441 still oriented parallel to the S1 foliation (Figure 5b).
442 These aggregates were folded during D3 deformation.

443 Randomly distributed andalusite porphyroblasts 2– 4 mm
444 in size overgrow the F3 crenulation (Figure 5d). The textural
445 features related to the growth of andalusite are better
446 developed in sample 16CA52. In this sample,
447 andalusite includes former Barrovian index minerals and
448 the F3 crenulations, or it is aligned with the newly
449 formed S3 foliation (Figure 5e,f). In the outcrop, strain
450 shadows adjacent andalusite related to the formation
451 of penetrative S3 foliation can be locally observed
452 (Figure 5g). In both samples, andalusite porphyroblasts
453 are commonly surrounded by an aggregate of large muscovite,
454 biotite and chlorite crystals. These crystals are in
455 sharp contact with andalusite.

456 3.3 | *Grt-Crd-Chl schist, sample 15CA27*

457 The D3 high-strain zone is characterized by a penetrative

458 subvertical S3 foliation defined by recrystallized ribbons of
459 quartz and plagioclase alternating with biotite-rich
460 domains. Relics of large staurolite porphyroblasts are only
461 locally preserved, and less commonly, they contain S1
462 inclusion trails. One garnet– cordierite– chlorite schist sample
463 15CA27 was selected for further investigation. The
464 sample contains garnet and cordierite porphyroblasts in a
465 matrix composed of biotite, plagioclase, quartz, chlorite,
466 ilmenite and accessory minerals (Figure 6a– c). Unlike the
467 above two samples, this sample does not contain
468 muscovite and aluminosilicate minerals (Ky/Sil/And).
469 Garnet (2– 4 mm in size) is either without inclusions or
470 with inclusions of quartz and ilmenite close to the margins,
471 oriented parallel to the external S3 fabric (Figure 6a,c).
472 Cordierite porphyroblasts (4– 8 mm in size) contain
473 numerous inclusions of ilmenite, quartz, biotite and chlorite
474 (Figure 6b– d). Cordierite shows complex geometrical
475 relations with respect to the S3 fabric. In some places, cordierite
476 crystals show straight inclusion trails composed of
477 fine-grained quartz, biotite and chlorite in the inner part
478 and coarse-grained inclusions of biotite and quartz along
479 cordierite margins (Figure 6b). The internal mineral inclusion
480 trails in the inner part of the cordierite occur at high
481 angle to the coarse-grained external S3 fabric, whereas the
482 coarse biotite and quartz inclusions in the marginal parts
483 are continuous with the matrix (Figure 6b). In other
484 places, cordierite crystals overgrow the F3 crenulations in
485 the core and the S3 foliation at the rim and are also further
486 wrapped by the same foliation with development of strain
487 shadows (Figure 6e,f). Similar crenulations are also
488 preserved in the matrix at the contact with cordierite and
489 are associated with development of a weak crenulation
490 cleavage S3 (Figure 6e,f). The crenulation axial planes as
491 well as the S3 cleavage are both parallel with the external
492 S3 foliation in the matrix. Apart from chlorite inclusions
493 in the cordierite, large chlorite crystals overgrowing the S3
494 foliation are locally present and are considered as a result
495 of post-D3 growth.

496 **3.4 | Interpretation of crystallization–**497 ***deformation relations***

498 Microstructural analysis indicates shape preferred orientations
499 of staurolite and kyanite porphyroblasts lying
500 within the S1 foliation. These porphyroblasts commonly
501 contain straight inclusions trails that are in continuity
502 with the external S1 foliation (e.g. Figures 3b, 4a,b and
503 5c), suggesting either post-tectonic or syntectonic growth
504 with respect to the S1 foliation. Despite significant deformation
505 during D2 and D3, some staurolite porphyroblasts
506 are wrapped by the same S1 foliation and exhibit strain
507 shadows (e.g. Figure 4c). These microstructural features
508 are compatible with dynamic growth rather than static
509 crystallization of the porphyroblasts and fit the diagnostic
510 scheme of Zwart (1962) of syntectonic growth. Therefore,
511 we interpret the growth of staurolite and kyanite as synchronous,
512 at least in part, with the D1 deformation. This
513 interpretation is also advocated by the modelled prograde
514 P–T evolution of the staurolite– kyanite-bearing S1 fabric,
515 which is in favour of dynamic rather than static growth
516 of these porphyroblasts, as introduced in Wei et al. (2007)
517 as well as in Section 5.

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

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

551 Andalusite porphyroblasts exhibit variable relationships
552 with respect to the heterogeneous and progressive
553 development of the S3 foliation. This is documented in
554 andalusite overgrowths of former Barrovian index minerals,
555 overgrowths of F3 crenulations and S3 cleavage
556 planes and development of the S3 strain shadows
557 around the andalusite porphyroblasts (e.g. Figures 4d
558 and 5e– g). Intensification of the F3 crenulations at the
559 margins of some andalusite grains can also be observed
560 (e.g. Figure 4e). These features suggest that andalusite
561 growth occurred during variable stages of D3 deformation;
562 it started after the beginning of the F3 crenulation,
563 and ended while the D3 deformation still continued,
564 thereby implying syn-D3 crystallization. By contrast, the
565 cordierite shows static overgrowth of the relatively finegrained
566 S1 foliation (Figure 6a), followed by rotation of
567 crystals and by further overgrowth of the F3 crenulations
568 during the D3 (Figure 6e,f). These features suggest that
569 the cordierite started to grow statically and continued its
570 growth during the progression of the D3 deformation.
571 The growth of cordierite over a progressively coarsened
572 matrix is most likely a consequence of heating. The

573 growth of andalusite and less commonly of cordierite is
574 also spatially related to pegmatite veins (Figure 2e),
575 suggesting an important role of magmatic activity on
576 crystallization of these porphyroblasts. Andalusite-bearing
577 quartz veins within the andalusite-bearing metamorphic
578 domains (Figure 2e) suggest circulation of
579 hydrothermal fluids during syn-metamorphic veining,
580 similar to a mechanism described in Cesare (1994).
581 Presence of large idiomorphic crystals of hydrous
582 minerals such as chlorite and muscovite and development
583 of quartz pseudomorphs after staurolite (Figure S2)
584 are also interpreted as a result of fluid circulation and
585 therefore as the effects of hydrothermal activity.
586 Garnet grains from the kyanite zone (sample
587 15CA34) and the sillimanite zone (sample 15CA32) contain
588 mineral inclusion trails that are in continuity with
589 the external S1 foliation and show equilibrium texture
590 with the associated staurolite and kyanite (e.g. Figures 3b
591 and 5a). The garnet from sample 15CA27 contains rare
592 inclusion trails that are in continuity with the external S3
593 foliation and displays equilibrium texture with the associated
594 cordierite (Figure 6a). These features suggest that
595 the garnet grains in samples 15CA34 and 15CA32 grew
596 during D1– M1 and in sample 15CA27 during D3– M3.

597 **4 | MINERAL CHEMISTRY**

598 Three samples (Grt-St-Ky-And micaschist 15CA34, Grt-
599 St-Sil-And micaschist 15CA32 and Grt-Crd-Chl schist
600 15CA27) were selected for further mineral chemical
601 analysis. Mineral composition analyses and garnet composition
602 mapping of samples 15CA34 and 15CA27 were
603 performed on a JEOL FEG-EPMA JXA-853 electron
604 microprobe at the Institute of Petrology and Structural
605 Geology (IPSG, Charles University in Prague). The compositional
606 analyses were performed in point beam mode
607 at 15 kV and 30 nA with a 5 μ m beam diameter and 30 s
608 counting time. Garnet compositional mapping was
609 acquired with 20 kV accelerating voltage and 70 nA beam
610 current (dwell time 40 ms per point) for typically 6– 8 h.

611 The analysis of sample 15CA32 was performed using a
612 JEOL JXA-8100 electron microprobe at the Key
613 Laboratory of Mineralogy and Metallogeny, Guangzhou
614 Institute of Geochemistry, Chinese Academy of Science
615 (GIG-CAS). The operating conditions for compositional
616 analysis were 15 kV accelerating voltage, 20 nA beam
617 current, 3– 5 μ m beam diameter and 20 s counting time.
618 The operation conditions for garnet mapping are accelerating
619 voltage of 20 kV, a probe current of 480 nA and a
620 4 μ m beam diameter (dwell time 100 ms per point) for
621 about 4– 6 h. The analyses were calibrated using multiple
622 natural reference materials and were reduced using the
623 ZAF correction routines. The precision of the mineral
624 composition analyses is better than 2% (relative) for most
625 oxides. In addition, x-ray compositional map of staurolite
626 and cordierite was obtained using a Carl Zeiss SUPRA55SAPPHIRE
627 field-emission scanning electron microscope
628 (FE-SEM) at GIG-CAS. The mapping was
629 conducted with a scanning resolution of 1,024 \times 768
630 pixel in a roughly 2.8 \times 2.1 mm area at 20 kV. The dwell
631 time of each pixel was 300 ms, and the mapping process
632 was operated 6– 8 h per sample.

633 Representative mineral analyses are summarized in
634 Table 1. Trends in mineral composition or zoning quoted
635 in the text are marked with ‘!’ , and ‘–’ designates a
636 range of mineral composition. Mineral abbreviations are
637 after Whitney and Evans (2010). Calculated endmember
638 proportions and cation ratios are defined as $alm = Fe_{2+} /$
639 $(Ca + Fe_{2+} + Mg + Mn)$, $sps = Mn / (Ca + Fe_{2+} + Mg$
640 $+ Mn)$, $py = Mg / (Ca + Fe_{2+} + Mg + Mn)$, $grs = Ca /$
641 $(Ca + Fe_{2+} + Mg + Mn)$, $X_{Fe} = Fe_{2+} / (Fe_{2+} + Mg)$.

642 **4.1 | Grt-St-Ky-And micaschist, sample**

643 **15CA34**

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

649 almandine, spessartine and X_{Fe} values at the very rim
 650 ($alm_{0.62!0.67!0.68}$ $sps_{0.20!0.14!0.15}$ $py_{0.12!0.14!0.12}$ $grs_{0.05-0.06}$, $X_{Fe} = 0.83-0.85!0.86$; Figure 7a). The garnet
 651 included in staurolite has slightly narrower almandine,
 652 spessartine and pyrope ranges ($alm_{0.64!0.67}$ sps decreasing
 653 Mn intensity from $co_{0.18!0.14}$ $py_{0.13!0.14}$ $grs_{0.05-0.06}$,
 654 $X_{Fe} = 0.83-0.84$). The matrix garnet x-ray map corroborates
 655 decreasing Mn intensity from core to rim
 656 (Figure 7a). Staurolite has a constant $X_{Fe} = 0.75-0.76$ in
 657 the interior part, and at the very rim, X_{Fe} reaches 0.79
 658 (Figure 7d), consistent with the Mg x-ray map
 659 (Figure 7e). Biotite crystals parallel to the S1 and S3
 660 foliations have the same mineral composition with
 661 $X_{Fe} = 0.41-0.42$ and $Ti = 0.07-0.09$ p.f.u., whereas biotite
 662 in contact with garnet has $X_{Fe} = 0.43$ and $Ti = 0.09$
 663 p.f.u. Plagioclase is not zoned and contains 30%–34% of
 664 anorthite.

665

666 **4.2 | Grt-St-Sil-And micaschist, sample**
 667 **15CA32**

668 Garnet shows a weak compositional zoning with
 669 slightly increasing X_{Fe} , almandine and pyrope
 670 and decreasing spessartine and flat grossular
 671 components from core to rim ($alm_{0.61!0.63}$ $sps_{0.22!0.20}$
 672 $py_{0.14!0.15}$ $grs_{\sim 0.03}$, $X_{Fe} = 0.81-0.82$; Figure 7b). At the
 673 very rim, spessartine increases to 0.22, whereas pyrope
 674 decreases to 0.13 (Figure 7b). The x-ray garnet map corroborates
 675 decreasing Mn intensity from core to rim and
 676 the increase at the very rim (Figure 7b). Staurolite grains
 677 show an increase in X_{Fe} from core to rim ($X_{Fe} = 0.77-0.80$)
 678 (Figure 7d), consistent with the Fe x-ray map
 679 (Figure 7f). Matrix biotite parallel to S1 and S3 has the
 680 same composition with $X_{Fe} = 0.44-0.45$ and $Ti = 0.08-0.09$
 681 p.f.u. Biotite in contact with garnet has similar composition
 682 ($X_{Fe} = 0.44$ and $Ti = 0.09$ p.f.u.). Biotite crystals
 683 in contact or included in andalusite show slightly
 684 different compositional ranges ($X_{Fe} = 0.44-0.48$ and
 685 $Ti = 0.07-0.09$ p.f.u.). Muscovite in contact with andalusite
 686 has $X_{Fe} = 0.23$ and $Si = 3.07$ p.f.u.

687 **4.3 | *Grt-Crd-Chl schist, sample 15CA27***

688 Garnet shows a slight increase in almandine, decrease in
689 spessartine and stable pyrope, grossular and X_{Fe}
690 values from core to rim ($alm_{0.56-0.60}$ $sps_{0.20-0.14}$ $py_{0.18-}$
691 $0.19}$ $grs_{0.06-0.07}$, $X_{Fe} = 0.75-0.76$; Figure 7c). Cordierite
692 porphyroblasts overgrowing the straight S1 inclusion
693 trails have a weak compositional zoning (Figure 7g),
694 characterized by decreasing X_{Fe} from core to
695 rim ($X_{Fe} = 0.18-0.15$) (Figure 7h). Cordierite
696 porphyroblasts overgrowing the F3 crenulation or the S3
697 foliation have $X_{Fe} = 0.17$. Coarse biotite included in the
698 rims of cordierite crystals has similar composition as the
699 matrix biotite ($X_{Fe} = 0.31-0.34$ and $Ti = 0.05-0.06$ p.f.u.).
700 Chlorite crystals included in cordierite and the matrix
701 chlorite show similar compositional ranges ($X_{Fe} = 0.29-$
702 0.30). Plagioclase is not zoned and has around 38% of
703 anorthite.

704 **5 | METAMORPHIC P-T**

705 **EVOLUTION**

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

725 the type M4 Tornado equipped with two XFlash® silicon
726 drift x-ray detectors at the Tuoyan Laboratories Ltd in
727 Guangzhou, China, operated at 50 kV and 300 μ A. The
728 selected areas were scanned at a pixel resolution of
729 12 μ m using a dwell time of 5 ms per pixel for about
730 1–2 h. Semi-quantitative data for major elements were
731 calculated from x-ray intensities via a standard-less
732 model with the software M4 TORNADO and then
733 normalized to 100% (Table S1). Several studies have
734 documented the fractionation effect of zoned garnet
735 porphyroblasts on the changes of effective bulk compositions
736 (e.g. Evans, 2004; Spear, 1988; Zeh, 2006). These
737 authors further suggested an ongoing assessment of
738 effective bulk composition as more appropriate when
739 modelling the **P–T** path associated with the growth of
740 garnet. However, as shown in Zeh (2006), the garnet fractionation
741 influenced the phase diagram topology, but its
742 effect on the resulting interpreted **P–T** path was insignificant.
743 Because the garnets in this study are weakly zoned,
744 their fractionation effect on effective bulk composition
745 was not considered. For the Buchan-type metamorphism
746 in these two samples, the heterogeneous overgrowth of
747 andalusite porphyroblasts makes a quantitative evaluation
748 of effective bulk composition less reliable. Therefore,
749 the **P–T** conditions were estimated using conventional
750 geothermobarometers. The remaining sample 15CA27
751 shows development of the S3 assemblage associated with
752 pervasive growth of cordierite. Its **P–T** evolution was
753 investigated by phase equilibria modelling using a XRF
754 whole rock composition.
755 Phase equilibria modelling was conducted by using
756 PerpleX Version 6.9.1 software package (Connolly, 2005)
757 with the upgraded thermodynamic database DS62
758 from Holland and Powell (2011) in the MnO– Na₂ O–
759 CaO– K₂ O– FeO– MgO– Al₂ O₃– SiO₂– H₂ O– TiO₂– Fe₂ O₃
760 (MnNCKFMASHTO) system. The activity– solution
761 TABLE models applied are chlorite, chloritoid, muscovite, biotite,
762 garnet, cordierite, orthopyroxene and melt (White
763 et al., 2014), plagioclase (Newton et al., 1980), epidote

764 (Holland & Powell, 2011) and ilmenite (White
765 et al., 2000). Phase diagrams for all samples were
766 calculated in subsolidus conditions where H₂O was set in
767 excess. Fe₂O₃ (O) was determined using calculated T–M
768 (O) pseudosections. Mineral composition and molar
769 isopleths are plotted for the phases of interest in order to
770 determine the P–T evolution of the studied samples.

771 **5.1 | Phase equilibria modelling for the**

772 **Barrovian-type M1 assemblage**

773 ***5.1.1 | P–T phase diagram for Grt-St-Ky-And*** 774 ***micaschist, sample 15CA34***

775 A bulk composition obtained from a 17.9 _ 5.9 mm large
776 continuous area covering typical Barrovian-type S1
777 fabric, that is, M1 mineral assemblage, in the thin
778 section (Figure S3a) was used for the calculation. The
779 composition is plotted in an AFM diagram in Figure S3a,
780 where it plots close to the average ‘high-Al pelites’
781 domain of Spear (1993). The effect of ferric iron was first
782 investigated in a T–M(O) phase diagram (not shown).
783 The observed assemblage is stable with O = 0.12–
784 0.20 mol.%, and an average value (O = 0.16 mol.%) was
785 chosen for the construction of P–T phase diagram. The
786 M1 assemblage kyanite–staurolite–garnet–biotite–muscovite–
787 plagioclase observed in the S1 foliation is comparable
788 with the stability field of St-Ky-Grt-Bt-Ms-Pl-Qz-H₂O
789 in a P–T range of 6.5–9.0 kbar and 640–650_C
790 (Figure 8a). Garnet rim composition (alm_{0.66–0.67}sp_{0.14–}
791 _{0.15}grs_{0.05–0.06}) broadly matches the modelled isopleths
792 within this field at 6.8–7.2 kbar and 645–655_C (the
793 upper right ‘circle’ in Figure 8b), which is interpreted as
794 the peak P–T range of the Barrovian-type M1
795 assemblage.
796 The compositions of the garnet core (alm_{0.62–}
797 _{0.63}sp_{0.20}grs_{0.05–0.06}) and the staurolite interior
798 (X_{Fe} = 0.76) are consistent with the calculated isopleths
799 at 5.6–6.4 kbar and 570–580_C in the stability field of
800 St-Grt-Bt-Ms-Chl-Pl-Qz-H₂O (the lower left ‘circle’ in
801 Figure 8b,c), which is taken as a P–T estimation for the

802 early garnet and staurolite growth. Therefore, a prograde
803 **P–T** path can be inferred, as indicated by the arrow in
804 Figure 8b, compatible with an increase of garnet volume
805 isopleths (Figure 8d). Staurolite volume proportions
806 increase significantly in the chlorite-present field
807 (Figure 8e), implying that growth of staurolite was initially
808 related to continuous chlorite breakdown reactions.
809 By contrast, staurolite volume proportions decrease significantly
810 in the kyanite-present field (Figure 8e),
811 suggesting that kyanite growth would occur at the
812 expense of staurolite along a prograde path. Recent findings
813 suggested that growth of kyanite in typical
814 Barrovian-type pelitic sequences could be approximately
815 contemporaneous with staurolite through chlorite-consuming
816 rather than staurolite-consuming reactions
817 (Pattison & Spear, 2018). However, such view is unlikely
818 the case for the current sample, because the garnet
819 compositional zoning is consistent with a prograde **P–T**
820 evolution through the conditions of the staurolite zone
821 and ending at the conditions of the kyanite zone
822 (Figure 8b). Even if replacement textures of staurolite by
823 kyanite have so far not been observed, kyanite could
824 form at one place, whereas staurolite could dissolve at
825 another place, by a similar mechanism as described in
826 Carmichael (1969).

827 **5.1.2 | *P–T phase diagram for Grt-St-Sil-And*** 828 ***micaschist, sample 15CA32***

829 The sample shows alternations of garnet–staurolite–sillimanite–
830 biotite-rich and quartz–feldspar-rich layering,
831 which defines the S1 foliation. A representative area of
832 9.9 _ 7.6 mm covering mainly the biotite–sillimanite-rich
833 layer was selected for the analysis of the effective bulk
834 composition (Figure S3b), which plots in the average
835 ‘high-Al pelites’ domain of Spear (1993) (see AFM
836 diagram in Figure S3b). The Fe₂O₃ amount for the
837 construction of the **P–T** phase diagram was inferred from
838 the T–**M**(O) phase diagram (not shown). The observed
839 peak assemblage (Grt-Sil-Ms-Bt-Pl-Qz-Ilm) is stable

840 with $O = 0.02\text{--}0.2$ mol.%. A nearly average value
841 ($O = 0.10$ mol.%) was chosen for the **P–T** phase diagram
842 construction, and such value is also comparable with a
843 generally reduced character of the metapelite protoliths.
844 With this value, the calculated **P–T** phase diagram is
845 characterized by sillimanite stability field at high temperature
846 conditions, staurolite stability field at medium
847 temperature conditions and garnet stability starting at
848 low temperature conditions (Figure 9a). Garnet rim composition
849 ($\text{alm}_{0.62\text{--}0.63}\text{sp}_{0.20\text{--}0.22}\text{gr}_{0.03}$) broadly matches
850 the modelled isopleths at 5.6–6.0 kbar and 660–670 °C in
851 the stability field of Sil-Grt-Bt-Ms-Pl-Qz-H₂O, as
852 indicated by a circle in Figure 9b, suggesting equilibration
853 of the peak assemblage under typical sillimanite
854 zone conditions. The $X_{\text{Fe}} = 0.77$ in staurolite interiors is
855 comparable with the modelled isopleth in the stability
856 fields of St-Grt-Bt-Ms-Pl-Qz-H₂O and St-Grt-Bt-Ms-Chl-
857 Qz-H₂O. Because the staurolite interior in the Grt-St-Ky-
858 And micaschist sample 15CA34 has a similar X_{Fe} that
859 constrains initial staurolite growth to the stability field of
860 St-Grt-Bt-Ms-Chl-Pl-Qz-H₂O, we assume that staurolite
861 growth in sample 15CA32 could also start at the
862 conditions within the chlorite-bearing stability field
863 ('circle' in Figure 9c). However, to reveal the prograde
864 **P–T** evolution is not straightforward. A weakly zoned
865 spessartine resembles typical prograde garnet zoning in
866 metapelites and therefore is an indicator of originally
867 prograde growth of the garnet, even if affected by significant
868 diffusional homogenization (e.g. Atherton, 1968;
869 Hollister, 1966). However, the overall flat compositional
870 pattern of the garnet (Figure 7b), commonly interpreted
871 as a result of re-equilibration by diffusion at high-grade
872 metamorphic conditions ($T_{\text{max}} > 650$ °C, e.g. Anderson &
873 Olimpio, 1977), prevents a deeper investigation of the
874 prograde **P–T** path. Nevertheless, microstructural
875 observations indicate that staurolite was partly replaced
876 by sillimanite, and, hence, an evolution from the staurolite
877 stability field to the sillimanite stability field ('dashed
878 arrow' in Figure 9d–f) is established, up to the **P–T**

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

884 **5.2 | P–T estimation for the Buchan-type**

885 **M3 assemblage**

886 In samples 15CA34 and 15CA32, growth of andalusite
887 porphyroblasts is very heterogeneous and probably associated
888 with hydrothermal activity. In this regard, phase
889 equilibrium modelling is not straightforward because the
890 estimation of the equilibrium bulk composition is difficult.
891 An alternative way to estimate **P–T** conditions is the
892 use of conventional geothermobarometers, because they
893 do not require an estimation of the composition of the
894 equilibrium volume (Wu, 2017). The very rim portions of
895 matrix garnets in both samples show significant reverse
896 trends in spessartine and pyrope contents (Figure 7a,b),
897 which suggests re-equilibration of the garnet outermost
898 rim. The biotite in contact with the garnet in both
899 samples commonly has X_{Fe} values higher compared with
900 biotite farther from garnet, which likely reflects re-equilibration
901 with the garnet rim. These compositional
902 features can be most easily explained as a result of M3
903 Buchan-type metamorphism, because the rocks were
904 thereafter not affected by any further metamorphic overprinting.
905 Therefore, these features are used as the basis
906 for **P–T** estimations of the M3 event.
907 Among the many geothermometers that can be
908 applied to metapelitic rocks, the garnet–biotite
909 (GB) geothermometer of Holdaway (2000) has been
910 suggested as one of the most robust (see summary in
911 Wu & Cheng, 2006). A recent calibration of garnet–biotite–
912 Al_2SiO_5 –quartz (GBAQ) geobarometer (Wu, 2017)
913 was proved useful for estimating pressures in particular
914 in CaO-deficient metapelitic rocks. Using the above GB–
915 GBAQ geothermobarometers, the outermost garnet rim
916 and biotite in contact with garnet were combined and

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

925 **5.3 | Phase equilibria modelling for Grt-** 926 **Crd-Chl schist, sample 15CA27**

927 The sample is characterized by a well-developed S3 fabric
928 with negligible relics of the S1 fabric and contains the
929 assemblage garnet, cordierite, biotite, plagioclase and
930 quartz (Figure S4). Small chlorite flakes are present as
931 inclusions in cordierite, whereas large post-tectonic chlorite
932 laths locally occur in the matrix. An initially inferred
933 bulk rock composition from thin section showed unusually
934 low X_{Fe} and SiO_2 contents compared with samples
935 15CA34 and 15CA32. Therefore, we performed also an
936 XRF whole rock composition of the sample (Table S1),
937 which showed similarly low X_{Fe} and SiO_2 contents, and
938 this new composition is used for the **P–T** phase equilibria
939 calculation. The effect of ferric iron content was
940 evaluated in a **T–M** (O) phase diagram, in which the
941 observed assemblage (Grt-Crd-Chl-Bt-Pl-Qz-Ilm) is stable
942 with O = 0.01–0.2 mol.% (not shown). An average value
943 (O = 0.10 mol.%) was chosen for the **P–T** phase diagram
944 construction, which is also consistent with a generally
945 reduced character of the metapelite protoliths and would
946 allow the discussion of stability of the observed assemblage
947 in the sample in general.

948 The resulting **P–T** phase diagram (Figure 10a)
949 shows a wide stability field of chlorite, which extends
950 to abnormally high temperature conditions of about
951 630 °C. Because a possible reason for such extended
952 chlorite stability is the high magnesium content of the
953 rock (e.g. Thompson, 1976), we evaluated its effect on
954 the stability of chlorite in a **T–M** (Mg) phase diagram

955 (Figure 10b). From the diagram, it is apparent
956 that increasing magnesium content in the system
957 increases significantly the upper temperature stability
958 of chlorite.
959 The observed assemblage can be correlated with the
960 stability field of Grt-Crd-Bt-Chl-Pl-Qz-H₂O in the calculated
961 **P–T** phase diagram (Figure 10a). Coarsening of
962 inclusions from cordierite core to rim suggests increasing
963 temperature during cordierite growth. Occurrence of
964 chlorite crystals in the core of the cordierite and their
965 absence in the external S3 foliation suggest the rocks
966 reached the chlorite-free Grt-Crd-Bt-Pl-Qz-H₂O stability
967 field (Figure 10c). This is consistent with cordierite compositional
968 zoning (Figure 7h), suggesting a prograde **P–T**
969 path starting from modelled $X_{\text{Fe}} = 0.18$ in the
970 Chl-present stability field of Grt-Crd-Bt-Chl-Pl-Qz-H₂O,
971 to $X_{\text{Fe}} = 0.15–0.16$ in the Chl-absent stability field of
972 Grt-Crd-Bt-Pl-Qz-H₂O, as indicated by the arrow in
973 Figure 10c. Garnet core (alm_{0.56} sps_{0.20} grs_{0.07–0.06}) and garnet
974 rim (alm_{0.60} sps_{0.14} grs_{0.06}) compositions broadly match
975 the modelled isopleths at relatively lower pressure and
976 temperature and higher pressure and temperature,
977 respectively (Figure 10d). Accordingly, a prograde **P–T**
978 path beginning at approximately 3.9– 4.2 kbar and
979 590– 610_ C and ending at approximately 4.4– 5.0 kbar
980 and 630– 650_ C can be established (the arrow in
981 Figure 10c– e). Along such prograde **P–T** path, molar proportions
982 of garnet are increasing (Figure 10e), supporting
983 prograde garnet growth.

984 **6 | U–PB MONAZITE AND LU–HF**

985 **GARNET GEOCHRONOLOGY**

986 **6.1 | Methods and sample preparation**

987 Y, La, Th and U compositional mapping of monazite
988 was performed on a JEOL JXA-8230 electron microprobe
989 at the Key Laboratory of Mineralogy and Metallogeny,
990 GIG-CAS. The operation conditions of an accelerate voltage
991 of 20 kV, a probe current of 700 nA and a beam size
992 of 1 μm were adopted for mapping. Elemental $K\alpha$ line

993 was chosen for all elements during analyses. The dwell
994 time was set to 100 ms for each point. Typically, 4– 6 h
995 was required to accomplish mapping for one grain. In
996 situ monazite U– Pb and simultaneous trace-element
997 analysis were acquired at the State Key Laboratory of
998 Geological Processes and Mineral Resources, China University
999 of Geosciences, Wuhan, using a ArF excimer
1000 GeoLas 2005 193 μ m connected to an Agilent 7500a ICPMS
1001 instrument. The isotopic results were corrected and
1002 calculated using ICPMSDataCal (Version 7.0, Liu
1003 et al., 2008). The age calculations and concordia plots
1004 were done using Isoplot 3.75 (Ludwig, 2003). The dating
1005 results and trace elements are given in Table S2. All ages
1006 are quoted at the 2s level of uncertainty. Among the
1007 above studied samples, only sample 15CA27 contains
1008 inclusion-poor clear garnet crystals, which are ideally
1009 suited for Lu– Hf garnet geochronology. The Lu– Hf
1010 isotopic analyses on this sample were conducted at the
1011 Krakow Research Center, Institute of Geological
1012 Sciences, Polish Academy of Sciences. Sample digestions,
1013 columns chemistry and mass spectrometry procedures
1014 are given in Thirlwall and Anczkiewicz (2004) and
1015 Anczkiewicz et al. (2004). The dating results are
1016 summarized in Table 2. Analytical procedures for U– Pb
1017 monazite and Lu-Hf garnet geochronology are available
1018 in Appendix S1.

1019 **6.2 | U–Pb ages and REE patterns of** 1020 **monazite from the D3 low-strain domain**

1021 Samples 15CA34 and 15CA32 from the D3 low-strain
1022 domain were selected for in situ monazite U– Pb dating
1023 and trace element analysis. These samples show comparable
1024 structures that are characterized by well-preserved
1025 S1 fabric that was folded but not transposed by D3. The
1026 analysis was applied only to the monazite grains included
1027 in either staurolite or andalusite porphyroblasts. Staurolite
1028 porphyroblasts in these samples show syn-D1 growth,
1029 whereas andalusite experienced syn-D3 growth. Monazite
1030 grains included in both staurolite and andalusite were

1031 analysed in sample 15CA34. In sample 15CA32, only
1032 monazites included in staurolite were large enough for
1033 analysis, whereas those in andalusite were too small.
1034 Instead, we studied a texturally and compositionally similar
1035 sample 16CA52 collected from the same metamorphic
1036 zone as sample 15CA32 (Figure 2a), as it preserves large
1037 monazite in andalusite.

1038 **6.2.1 | Monazite in syn-D1 staurolite:**

1039 ***Samples 15CA34 and 15CA32***

1040 Numerous monazite grains were found in samples
1041 15CA34 and 15CA32, which are distributed in the matrix
1042 as well as in staurolite and kyanite porphyroblasts; no
1043 monazite was found in garnet, and those in kyanite were
1044 not selected for analysis due to small grain size. Seven
1045 and four grains included in staurolite porphyroblasts of
1046 samples 15CA34 and 15CA32, respectively, were studied.
1047 Monazite grains in these two samples exhibit similar
1048 morphological features characterized by either subhedral
1049 or irregular shape. In x-ray images, most grains do not
1050 show significant zoning of Y, except for a few grains that
1051 exhibit high Y, U and Th patches (Figure 11a).
1052 Thirty-nine analyses were performed on monazite in
1053 sample 15CA34. The majority of data yield concordant
1054 ages ranging from 280 to 260 Ma (Figure 11a) with
1055 a weighted mean $^{206}\text{Pb}/^{238}\text{U}$ age of 270 ± 3 Ma
1056 (MSWD = 3.8). Several analyses on the high-Y domains
1057 yield relatively older ages spreading from 350 to 300 Ma
1058 (Figure 11a). All analyses show right-dipping REE distribution
1059 patterns with negative Eu anomalies, irrespective
1060 of their ages (Figure 11a).
1061 For sample 15CA32, 13 analyses were performed. The
1062 resulting data show a similar age pattern as in sample
1063 15CA34. Ten analyses yield similar U–Pb ages falling in a
1064 range of 270–260 Ma, corresponding to a weighted mean
1065 $^{206}\text{Pb}/^{238}\text{U}$ age of 262 ± 3 Ma (MSWD = 0.9; Figure 11b).
1066 Two remaining analyses on the high-Y portions give
1067 $^{206}\text{Pb}/^{238}\text{U}$ ages of 309 and 294 Ma, respectively
1068 (Figure 11b). Chondrite-normalized REE patterns exhibit

1069 also decrease towards the HREE. However, when compared
1070 with sample 15CA34, they show weaker negative
1071 Eu anomalies and less depleted heavy REE patterns
1072 (Figure 11b).

1073 **6.2.2 | Monazite in syn-D3 andalusite:**

1074 **Samples 15CA34 and 16CA52**

1075 Six and eight analyses were performed on monazite
1076 grains included in syn-D3 andalusite from samples
1077 15CA34 and 16CA52, respectively. These grains either sit
1078 within the micro-folded S1 fabric or along the S3 fabric
1079 (e.g. Figure 11b). Locally, monazite grains occur in
1080 clusters of multiple tiny grains (Figure 11a), morphologically
1081 similar to hydrothermal monazite or monazite
1082 altered by metamorphic fluids (e.g. Schandl &
1083 Gorton, 2004; Williams et al., 2011). Compared with
1084 monazite included in staurolite, monazite crystals from
1085 andalusite porphyroblasts have irregular shape and lack
1086 Y zoning (Figure 11a,b).

1087 Six analyses in sample 15CA34 yield consistent
1088 $^{206}\text{Pb}/^{238}\text{U}$ ages ranging from 290 to 270 Ma. These data
1089 form a tight cluster of concordant points, giving a
1090 weighted mean $^{206}\text{Pb}/^{238}\text{U}$ age of 279 ± 7 (MSWD = 1.6;
1091 Figure 11a). Eight analyses in sample 16CA52 give
1092 $^{206}\text{Pb}/^{238}\text{U}$ ages ranging between 280 and 260 Ma, with
1093 a weighted mean $^{206}\text{Pb}/^{238}\text{U}$ age of 269 ± 5 Ma
1094 (MSWD = 1.5; Figure 11b). These analyses have higher
1095 heavy REE contents and show relatively weaker Eu negative
1096 anomalies compared with analyses from monazite in
1097 staurolite, despite their similar ages.

1098 **6.3 | Monazite in the D3 high-strain**

1099 **zone**

1100 Monazite grains from the garnet– cordierite– chlorite
1101 schist, sample 15CA27, were analysed. Monazite grains
1102 occur both in the matrix and in cordierite porphyroblasts.
1103 Most grains are too small for analysis, but three relatively
1104 large grains from cordierite porphyroblasts have been
1105 measured. These monazite crystals are all xenomorphic
1106 and do not show notable Y zoning but have patchy

1107 zoning in Th and U (Figure 11c). Four analyses give
1108 nearly equivalent $^{206}\text{Pb}/^{238}\text{U}$ ages in a range of 270–
1109 265 Ma (Figure 11c), forming a single age population
1110 with a weighted mean $^{206}\text{Pb}/^{238}\text{U}$ age of 268 ± 5 Ma
1111 (MSWD = 0.26).

1112 **6.4 | Lu–Hf garnet geochronology**

1113 Lu–Hf isotope dilution analyses were conducted on one
1114 whole rock and three garnet fractions derived from
1115 garnet–cordierite–chlorite schist (sample 15CA27). The
1116 analyses yielded very similar $^{176}\text{Lu}/^{177}\text{Hf}$ ratios for the
1117 garnet fractions, which range between 0.749 and 0.763.
1118 This suggests well-mixed aliquots or homogeneous
1119 distribution of Lu and Hf in garnet. Concentration of Hf
1120 is higher than that in typical metamorphic garnet
1121 (e.g. Anczkiewicz et al., 2007), which indicates the
1122 presence of some Hf-rich inclusion(s) in garnet fractions.
1123 Likely, ilmenite inclusions are enriched in Hf relative to
1124 garnet and therefore could contribute to the Hf concentration
1125 budget, thus lowering $^{176}\text{Lu}/^{177}\text{Hf}$ ratios of the
1126 garnet dissolutions. Importantly, ilmenite is distributed
1127 evenly throughout the garnet and the rest of the rock,
1128 forming part of the paragenesis of the sample (Figure 6c).
1129 As documented in Scherer et al. (2000), when Hf-rich
1130 accessory phases are present in both garnet and the
1131 matrix, their effects on Lu–Hf garnet dating accuracy are
1132 insignificant. Lu–Hf isochron defined by three garnet
1133 fractions and a whole rock yields an age of 261.9 ± 3.4
1134 (MSWD = 3.5; Figure 12). This age is interpreted as
1135 reflecting the time of garnet formation.

1136 **7 | DISCUSSION**

1137 **7.1 | Interpretations of geochronological**

1138 **results**

1139 The Permian ages obtained mostly ranging from 280 to
1140 260 Ma are the most prominent age population in the
1141 study area. A large number of concordant 279–268 Ma
1142 U–Pb ages were obtained from the preferentially
1143 aligned S3-parallel monazite grains in the andalusite
1144 porphyroblasts, from monazite grains in the micro-folded

1145 S1 fabric in andalusite porphyroblasts and from
1146 monazite grains in the syn-D3 cordierite porphyroblasts
1147 (Figure 11). These data are interpreted as the timing of
1148 M3 Buchan-type metamorphism. This interpretation is
1149 supported by the fact that growth of the typical Buchantype
1150 index minerals (andalusite and cordierite) was connected
1151 with the D3 event that was previously constrained
1152 at 280– 273 Ma (Jiang et al., 2019).

1153 A large number of monazite inclusions parallel to S1
1154 fabric in the staurolite porphyroblasts also gave predominant
1155 270– 262 Ma ages and minor older ages scattering at
1156 350, 330, 300 and 290 Ma (samples 15CA34 and 15CA32;
1157 Figure 11a,b). These ages are much younger than supposed
1158 timing of the S1 fabric formation, which should be
1159 at least older than 400 Ma (Jiang et al., 2019), implying
1160 later equilibration or recrystallization of the monazite.

1161 Irrespective of their ages, REE patterns of all these monazite
1162 spots exhibit similar distributions with moderate to
1163 significant Eu troughs in the chondrite-normalized
1164 distribution patterns (Figure 11a,b). These features are
1165 different from those of monazite spots in the andalusite
1166 porphyroblasts, which are marked by insignificant
1167 negative Eu anomalies and systematically higher heavy
1168 REE contents (Figure 11a,b). This apparent difference
1169 implies that monazite grains included in the M1
1170 staurolite are originally dissimilar from those included in
1171 the M3 andalusite. In addition, the analyses giving
1172 >290 Ma ages were mainly obtained from the high-Y
1173 monazite domains, whereas the analyses with uniform
1174 Permian ages were obtained from the low-Y domains
1175 (Figure 11a,b).

1176 Although monazite included in porphyroblasts is
1177 expected to be protected from later equilibration or
1178 recrystallization (e.g. Foster et al., 2004; Martin
1179 et al., 2007), it has been documented that monazite inclusions
1180 can undergo partial or complete recrystallization
1181 when they are connected to the matrix via microcracks
1182 (e.g. Hoisch et al., 2008; Martin et al., 2007; Montel
1183 et al., 2000). When a fluid is present, such recrystallization,

1184 for example, by dissolution– precipitation processes,
1185 can occur, even at temperatures far below the diffusional
1186 closure temperature of the U– Th– Pb system in the
1187 monazite (Kelly et al., 2012; Taylor et al., 2014; Williams
1188 et al., 2011). In this study, the staurolite texture
1189 is poikiloblastic and always contains microcracks
1190 (e.g. Figure 4b,c). Such textural features indicate that
1191 monazite inclusions were not sufficiently isolated from
1192 the matrix and fluids associated with the D3– M3 event
1193 could cause recrystallization of monazite. Altogether, the
1194 Permian monazite ages in the staurolite porphyroblasts
1195 are interpreted to reflect almost complete re-equilibration
1196 of U– Th– Pb system during the Buchan-type metamorphic
1197 overprint. Likewise, the few >290 Ma dates scattering
1198 along the concordia are probably attributed to partial
1199 re-equilibration of the U– Th– Pb system in the monazite,
1200 and these ages are therefore geologically meaningless.
1201 Lu– Hf garnet– whole rock isotopic data of sample
1202 15CA27 formed a regression line with an isochron age of
1203 261.9 ± 3.4 Ma. This age is identical within the analytical
1204 uncertainties to that of monazite included in the cordierite
1205 (268 ± 5 Ma; Figure 12) from the same sample. These
1206 ages are nearly overlapping with the age populations
1207 (270 ± 3, 279 ± 7 Ma, 269 ± 5 and 262 ± 3 Ma;
1208 Figure 11a,b) reported from the majority of monazite
1209 grains from the D3 low-strain domains. Such an age
1210 range is comparable with the zircon ages from the pegmatite
1211 dykes in the region, that is, 277– 273 Ma (Jiang
1212 et al., 2019). The close spatial relation between arrays of
1213 pegmatite dykes and development of Buchan-type
1214 andalusite and cordierite domains (Figure 2d,e) attests to
1215 close genetic links. Jiang et al. (2019) outlined that the
1216 pegmatite dykes emplaced in tensional fractures were
1217 considered to form parallel to principal compressive
1218 stress (Mode I fractures) of the D3. Emplacement of
1219 pegmatite dyke is always thought to connected with
1220 propagating hydrofractures (e.g. Mériaux et al., 1999;
1221 Weinberg, 1999). The intrusion of magma would enhance
1222 wedging-apart of the fracture walls and cause overpressure

1223 effects at the fracture tips, thus leading to further
1224 fracturing (Clemens & Mawer, 1992; Rubin, 1993). This
1225 process would also heat, soften and enhance the matrix
1226 permeability of the wall rocks in the presence of volatilerich
1227 fluids (Connolly et al., 1997). In the study area, we
1228 suggest that such a process would account for advecting
1229 heat flux to the hosting rocks via emplacement of pegmatite
1230 dykes, resulting in widespread growth of andalusite
1231 domains (Figure 2d) and growth of new monazite
1232 included by the andalusite and cordierite porphyroblasts,
1233 and also to nearly complete recrystallization of monazite
1234 in staurolite porphyroblasts. The geochronological
1235 relationships presented here can be interpreted as a result
1236 of progressive deformation and heterogeneous heat and
1237 fluid transfer related to emplacement of pegmatites
1238 (Thompson & Connolly, 1992). In this regard, the current
1239 geochronological work provides constraints on duration
1240 and progression of Permian deformation and metamorphism
1241 in the region, that is, starting from at ca. 280 Ma
1242 and ending at ca. 260 Ma.

1243 **7.2 | P–T–D constraints of**

1244 **tectonometamorphic evolution of the**

1245 **Chinese Altai**

1246 Microstructural data combined with phase equilibria
1247 modelling allow deciphering the **P–T–D** histories of
1248 Barrovian- and Buchan-type metamorphism in the
1249 region. Data from this study together with available
1250 regional data are summarized in Figure 13, and their
1251 significance for regional tectonometamorphic evolution
1252 is portrayed in Figure 14.

1253 **7.2.1 | P–T path of D1–M1**

1254 It has been shown above that growth of the index
1255 minerals garnet (except for cordierite-bearing schists),
1256 staurolite, kyanite and sillimanite is associated with the
1257 D1–M1 event. Phase equilibria modelling constrained
1258 simultaneous increase of pressure and temperature
1259 associated with a Barrovian-type **MP /MT** thermal
1260 gradient of approximately 23_ C/km for the kyanitebearing

1261 assemblage in sample 15CA34, in agreement
1262 with the **P–T** evolution modelled previously for kyanite
1263 zone in the region (see details in Wei et al., 2007). For the
1264 staurolite-bearing assemblage in sample 15CA32, its prograde
1265 **P–T** evolution could not be currently constrained;
1266 however, a **P–T** path beginning at approximately 6– 7 kbar
1267 and 560_ C/and ending at approximately 7.3 kbar and
1268 593_ C (with uncertainties of _ 10_ C and kbar) was
1269 reported for the staurolite zone in the study area (Wei
1270 et al., 2007). If this value is taken into account, a
1271 prograde **P–T** path associated with a thermal gradient of
1272 approximately 23– 24_ C/km for the staurolite-bearing
1273 assemblage can be constructed. In these regards, it can be
1274 concluded that the D1– M1 event accounting for the
1275 generation of staurolite- and kyanite-bearing is marked
1276 by Barrovian-type **MP /MT** prograde **P–T** evolution with a
1277 maximum pressure of 7.3 _ 1.0 kbar or higher (arrow
1278 ‘D1B’ in Figure 13a). Such a prograde **P–T** evolution is
1279 very similar to that deduced from the adjacent staurolite–
1280 kyanite-bearing migmatitic complex; the latter reaches a
1281 maximum pressure of 8.0 _ 1.0 kbar or higher (arrow
1282 ‘D1B’ in Figure 13b; Jiang et al., 2019). These features
1283 indicate that both the orogenic middle crust (e.g. the
1284 medium-grade metamorphosed Habahe Group) and the
1285 orogenic lower crust (e.g. the migmatite-granite complex)
1286 were affected by a Barrovian metamorphism and hence
1287 progressive burial during the course of D1B.
1288 The replacement of staurolite by sillimanite-biotite
1289 aggregates in sample 15CA32 (Figure 5d) indicates reequilibration
1290 of the peak assemblage from the staurolite
1291 stability field to a lower pressure and higher temperature
1292 sillimanite stability field. Development of sillimanitebearing
1293 assemblages would therefore reflect a **P–T** evolution
1294 beginning at approximately 7.3 kbar and 593_ C and
1295 ending at approximately 5.6– 6.0 kbar and 660– 670_ C, the
1296 latter corresponding to a **LP /HT** thermal gradient of
1297 approximately 34_ C/km. The switch from **MP /MT** to **LP /**
1298 **HT** thermal gradients is similar to that deduced from
1299 structurally deeper migmatites from the adjacent

1300 regions (Figure 13b; Jiang et al., 2015). There, the relics
1301 of staurolite- and/or kyanite-bearing assemblages
1302 equilibrated at approximately 8 kbar and 650_ C were
1303 overprinted by garnet– K-feldspar– sillimanite-bearing
1304 assemblage, corresponding to a re-equilibration at
1305 approximately 7.0– 8.0 kbar and 770– 800_ C during D1M
1306 (Jiang et al., 2015; see also the arrow ‘ D1M’ in Figure
1307 13b). In addition, the fabric in the deepest orogenic
1308 lower crust shows extensional structures (Broussolle
1309 et al., 2018; Jiang et al., 2019), which, together with
1310 simultaneously formed volcano-sedimentary basins in
1311 the orogenic upper crust (Wan et al., 2011), indicate a
1312 generalized extensional regime during D1M (Figure 14a).
1313 It is therefore most likely that the growth of S1-parallel
1314 sillimanite reflects the effects of D1M extension on the
1315 orogenic middle crust leading to partial LP /HT reequilibration
1316 of the former Barrovian-type assemblage
1317 (the arrow ‘ D1M’ in Figure 13a).

1318 **7.2.2 | P–T path of D2–M2**

1319 The P–T evolution related to the D2– M2
1320 tectonometamorphic event is not constrained because no
1321 M2 metamorphic assemblages were observed in the investigated
1322 samples. However, garnet, staurolite, kyanite and
1323 sillimanite schists are sharply juxtaposed with the neighbouring
1324 biotite schists of the Devonian sequence prior to
1325 the F3 folding (Figure 2a,b), suggesting that the orogenic
1326 middle crust was exhumed to shallower levels already
1327 during the D2 upright folding (Figure 14a; see also Jiang
1328 et al., 2019). Such large-scale upright folding described in
1329 other parts of the Altai was responsible for the extrusion
1330 of deep-seated orogenic lower crust in the cores of gneiss
1331 domes (Broussolle et al., 2015; Jiang et al., 2015; Lehmann
1332 et al., 2017; Zhang et al., 2015). The biotite schist has a
1333 peak P–T condition of approximately 4 _ 1 kbar and
1334 500 _ 10_ C (Wei et al., 2007). It is hence most likely that
1335 the kyanite and sillimanite schists might probably at least
1336 pass to 4 kbar and 500_ C and shared exhumation with the
1337 biotite schist prior to their juxtaposition. In other words,

1338 the studied kyanite and sillimanite schists were probably
1339 exhumed from approximately 20 km depth ($P = 6\text{--}5$ kbar)
1340 to at least 10 km depth ($P = 4\text{--}2$ kbar) during the D2
1341 event (the arrow ‘D2’ in Figure 13a). The preservation of
1342 kyanite- and sillimanite-bearing assemblages during the
1343 D2 exhumation was probably due to passive elevation of
1344 deep crustal rocks without deformation and in the
1345 absence of aqueous fluids (Wei et al., 2007).
1346 The $P\text{--}T$ evolution related to the D2– M2 event has
1347 been constrained for the orogenic lower crust
1348 where K-feldspar– sillimanite-bearing assemblage was reequilibrated
1349 in the cordierite stability field, as exemplified
1350 by the preservation of this mineral in the F2 planar
1351 leucosomes (e.g. Jiang et al., 2015; Wei et al., 2007; see
1352 also Figure 13b). $P\text{--}T$ phase equilibria modelling revealed
1353 that this evolution was associated with an exhumation
1354 path from 8 to 9 kbar to at least 4– 5 kbar (arrow ‘D2’ in
1355 Figure 13b; Jiang et al., 2015; Wei et al., 2007).
1356 Altogether, the D2– M2 event is characterized by a
1357 significant decompression from different pressure peaks
1358 to 2– 4 kbar. This is in agreement with the model of
1359 growth of a hot migmatite-granite dome responsible for
1360 vertical transposition, elevation and thinning of originally
1361 subhorizontal metamorphic isograds (Figure 14a).

1362 **7.2.3 | $P\text{--}T$ path of D3–M3**

1363 The D3 folding is responsible for the development of
1364 biotite- and muscovite-bearing S3 foliation zones,
1365 suggesting overall biotite-zone regional metamorphic
1366 conditions. However, the S3 was locally associated with
1367 formation of low-pressure andalusite- and cordieritebearing
1368 assemblages in the metamorphosed Habahe
1369 Group. As it has been suggested above, growth of andalusite
1370 in the D3 low-strain domains was developed at
1371 approximately 3– 4 kbar and 530– 600_ C, corresponding to
1372 a thermal gradients of 48_ C/km or more (e.g. samples
1373 15CA34 and 15CA32, arrow ‘D3’ in Figure 13c), and
1374 growth of cordierite in the D3 high-strain zones was associated
1375 with a prograde $P\text{--}T$ path ending at approximately

1376 4.3– 5.0 kbar and 630– 650_ C, corresponding to a thermal
1377 gradient of approximately 41_ C/km (e.g. sample 15CA27,
1378 the arrow ‘ D3’ in Figure 13d). The prominent feature of
1379 the M3 **P–T** evolution is associated with abnormally high
1380 thermal gradients, implying local additional input of
1381 heat. The **P–T** path developed in pressure conditions of
1382 approximately 4.0– 5.0 kbar, within the uncertainties of
1383 the **P–T** estimations, suggesting M3 might have been
1384 nearly isobaric heating or even was associated with slight
1385 burial or exhumation during D3 shortening.
1386 Localized heat supply in the region is indicated by
1387 close spatial and temporal relationships between lowpressure
1388 assemblages and intrusions of pegmatite dykes.
1389 This is best exemplified by the deformation– crystallization
1390 relationships and geochronology of sample 15CA27
1391 in a D3 high-strain zone, where growth of cordierite and
1392 garnet porphyroblasts and emplacement of pegmatite
1393 dykes are dated at ca. 268– 262 Ma (Figures 11 and 12)
1394 and 277– 273 Ma (Jiang et al., 2019), respectively. The
1395 first static overgrowth of the S1 by the cordierite
1396 porphyroblasts (e.g. Figure 6a) may reflect input of heat
1397 advected from incipient intrusion of pegmatite dykes
1398 (‘ Stage 1’ in Figure 13d). Ongoing intrusions of pegmatite
1399 dykes provided further heat, enhancing weakening and
1400 folding of S1 related with the sequential growth of the
1401 cordierite (e.g. Figure 6e,f) and garnet porphyroblasts
1402 during the D3 transposition (‘ Stage 2’ in Figure 13d).

1403 **7.3 | Barrovian- and Buchan-type** 1404 **metamorphism: Proxies for tectonic** 1405 **switching**

1406 In the Chinese Altai, the earliest Barrovian metamorphism
1407 of presumably Late Silurian to Early Devonian age
1408 is characterized by regularly spaced metamorphic zones
1409 and increasing metamorphic grade with depth (e.g. Jiang
1410 et al., 2015; Wei et al., 2007; see also Figure 14a). This
1411 metamorphic phase results in homogeneous but moderate
1412 crustal thickening to approximately 30– 35 km along
1413 **MP /HT** thermal gradients of 20– 25_ C/km (Figure 14c).

1414 Similar Barrovian metamorphism was also reported from
1415 the neighbouring Mongolian Altai (e.g. Burenjargal
1416 et al., 2014; Zorigtkhuu et al., 2011), suggesting that the
1417 whole Altai wedge was first affected by such burial.
1418 Replacement of the Barrovian-type assemblages by LP /
1419 HT ones during D1M is the hallmark of a significant
1420 switch of thermal and tectonic regimes affecting the
1421 whole Altai wedge crust during Middle Devonian. At
1422 depth, the bottom of the thickened orogenic lower crust
1423 was partially molten at low pressure (Figure 14a; Hanžl
1424 et al., 2016; Jiang et al., 2015; Zhang et al., 2015), leaving
1425 behind a garnet– orthopyroxene granulite residuum
1426 (Jiang et al., 2016; Kozakov et al., 2002). The orogenic
1427 middle crust was decompressed during this event as
1428 shown by partial re-equilibration of the Barrovian-type
1429 assemblages in the sillimanite stability field (sample
1430 15CA32; Figures 13a and 14a). At the same time, extrusion
1431 of bimodal volcanic rocks occurs in Devonian extensional
1432 basins (e.g. Soejono et al., 2018; Wan et al., 2011).
1433 This major thermal event is typical for large heat input
1434 from the mantle associated with thinning and horizontal
1435 stretching of the whole lithosphere (Jiang et al., 2016).
1436 The second episode of deformation (D2) is a decompression
1437 process that is related to renewed shortening,
1438 leading to growth of large migmatite-granite domes during
1439 Middle– Late Devonian. During this event, the hot
1440 orogenic lower and middle crust were exhumed to shallow
1441 crustal levels, forming the domes (Figure 14a). In the
1442 Chinese Altai, this deformation episode is characterized
1443 by formation of LP /HT cordierite-bearing migmatites in
1444 the core of domes (e.g. Jiang et al., 2015, see also
1445 Figure 14c), whereas, in the neighbouring Mongolian
1446 Altai, it led to a Buchan metamorphism marked by
1447 growth of andalusite and cordierite in the metamorphic
1448 envelopes of the domes (Broussolle et al., 2015; Lehmann
1449 et al., 2017).
1450 Jiang et al. (2019) showed that the succession of
1451 Devonian tectonometamorphic events affecting the Chinese
1452 Altai can be explained by the tectonic switching

1453 model of Collins (2002), which is characterized by alternation
1454 between shortening and extensional deformation.
1455 In theory, such tectonic switches would lead to formation
1456 of multiple short-lived burial– exhumation cycles in
1457 subduction orogens and result in multiple burial–
1458 exhumation **P–T** cycles (Beltrando et al., 2007). We argue
1459 that the succession of short-lived Silurian– Devonian **MP /**
1460 **MT** and Middle Devonian **LP /HT** metamorphic cycles in
1461 the Chinese Altai may define the tectonic switching
1462 characterized by shortening followed by extension and
1463 renewed shortening (Figure 14a). These metamorphic
1464 and tectonic cycles can be considered as proxies of alternating
1465 retreats and advances of the subduction slab of the
1466 Paleo-Pacific Ocean (Jiang et al., 2017), which is also typical
1467 for accretionary orogenic systems (Collins, 2002).
1468 The most spectacular Buchan-type metamorphism in
1469 the region is related to the Permian D3 heterogeneous
1470 reworking of the whole southern Chinese Altai
1471 (Broussolle et al., 2019). This event is constrained here to
1472 have lasted mainly from 280 to 260 Ma. This highly
1473 heterogeneous metamorphic event is related to nearly
1474 isobaric heating. Locally, ultra-high-temperature, lowpressure
1475 granulite facies metamorphism develops in narrow
1476 NE-trending tabular extrusion zones (e.g. Broussolle
1477 et al., 2018; Liu et al., 2020; Tong et al., 2014; Wang
1478 et al., 2009, see also Figure 15b,c). This important
1479 Buchan-type metamorphic cycle is probably connected
1480 with the coeval collision between the Chinese Altai with
1481 the southerly Junggar arc domain (Guy et al., 2020; Jiang
1482 et al., 2019). The collisional event is considered as the
1483 consequence of an interplay of large-scale oroclinal bending,
1484 exemplified by the Mongolian Orocline (e.g. Guy
1485 et al., 2020), formation of narrow E– W trending alkaline
1486 magmatic provinces that are axial planar to the orocline
1487 (e.g. Kovalenko et al., 2004) and closure of oceanic basins
1488 further south (e.g. Xiao et al., 2015). The heat source of
1489 this Buchan-type metamorphism is most likely linked
1490 with narrow linear thermal perturbations of the mantle
1491 related to the above-mentioned processes. This is

1492 consistent with recent findings that a short-lived Early
1493 Permian extension existed during the collision cited
1494 above (Li et al., 2015, 2017), which might provide heat
1495 source for the high-temperature Buchan-type metamorphism
1496 in the southern Chinese Altai.

1497 **8 | CONCLUSIONS**

1498 We undertook a multidisciplinary investigation that
1499 allowed the tectonometamorphic evolution of spatially
1500 overlapping Barrovian- and Buchan-type metamorphic
1501 cycles in the Chinese Altai to be defined, summarized as
1502 follows:

1503 1. Barrovian-type metamorphic series including garnet,
1504 staurolite and kyanite zones developed synchronously
1505 with the regional S_{1B} foliation, followed by partial reequilibration
1506 of the staurolite zone, resulting in
1507 formation of a sillimanite-bearing S_{1M} foliation, still
1508 in parallelism with S_{1B}. The S_{1B-M} foliations were subsequently
1509 folded by the D₂ upright folds and finally
1510 refolded orthogonally by the D₃ folds. In contrast, the
1511 Buchan-type metamorphic series are represented by
1512 localized overgrowth of andalusite- and cordieritebearing
1513 assemblages associated with D₃ deformation.

1514 2. Barrovian-type staurolite– kyanite sequence shows a
1515 prograde P–T evolution along MP /MT thermal
1516 gradients. This event is followed by high-temperature
1517 partial re-equilibration in the sillimanite stability field
1518 under LP /HT thermal regime and further important
1519 decompression without apparent re-equilibration.

1520 This sequence of metamorphism corresponds to syn-
1521 D_{1B} burial and thickening, syn-D_{1M} decompression
1522 and heating, followed by D₂ compressive doming and
1523 associated exhumation. The Buchan-type metamorphism
1524 is related to syn-D₃ heating associated with a
1525 renewed LP /HT thermal regime.

1526 3. U– Th– Pb system in monazite preserved as inclusions
1527 in the Barrovian index minerals was nearly
1528 completely re-equilibrated, and monazites in the main
1529 Buchan-type assemblage yield ages of 280– 260 Ma.

1530 These data together with Lu– Hf garnet– whole rock
1531 age of ca. 262 Ma constrain the duration of Permian
1532 Buchan-type metamorphism.
1533 4. The sequence of tectonometamorphic events in the
1534 Chinese Altai reflects the Devonian D1– D2 coaxial
1535 suprasubduction sequence of tectonic switching,
1536 whereas the D3 shortening and main Permian
1537 Buchan-type metamorphism reflects complex deformation
1538 and heat transfer process related to Early
1539 Permian collision between the Chinese Altai and its
1540 southerly Junggar arc domain.

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1559 **DATA AVAILABILITY STATEMENT**

1560 The data that supports the findings of this study are available
1561 in the supplementary material of this article

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