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1 **Early Paleozoic intracontinental orogeny in the Yunkai Domain,**
2 **(South China Block): new insights from field observations, zircon U–**
3 **Pb geochronological and geochemical investigations**

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10 **ABSTRACT**

11 Debate on whether the Early Paleozoic tectono–magmatic event in South China is
12 related to a subduction–collision or an intracontinental orogen has been lasted for
13 decades within the geoscience community. This study deals with LA–ICP–MS zircon
14 U–Pb ages, whole–rock chemistry, rare earth elements, trace elements and Hf isotopes
15 from granitoid samples collected in the Yunkai domain in order to better constrain the
16 Early Paleozoic tectonic evolution of the South China Block. The weighted mean
17 $^{206}\text{Pb}/^{238}\text{U}$ ages for eight samples range from 426 Ma to 443 Ma, representing the
18 crystallization ages of the magma. Fourteen samples were analyzed for geochemistry,
19 all of which are characterized by a peraluminous signature with A/CNK values greater
20 than 1.0. The REE geochemistry reveals enrichment in light rare earth element.
21 LREE/HREE values range from 2.81 to 30.36 and $(\text{La}/\text{Yb})_{\text{N}}$ vary from 1.23 to 55.14
22 (mean of 14 analyses is 14.69). All the samples exhibit distinct negative Ba, Sr and Nb
23 anomalies and enrichment in Rb, Th, U and Pb. Hf isotopic analyses indicate negative
24 $\epsilon_{\text{Hf}}(t)$ values mainly ranging from -3 to -12, corresponding to two model age
25 distributing from 1637 Ma to 2208 Ma. The geochemical analyses indicate that the
26 Silurian granitic magmas in the Yunkai domain were derived from partial melting of
27 crustal materials with little or no input of mantle source. These new data support the

28 intracontinental subduction model already proposed to account for the Early Paleozoic
29 tectonic, metamorphic and magmatic event of South China.

30

31 **Keywords:** Early Paleozoic Granitoids; U–Pb geochronology; Geochemistry; Hf
32 isotope; Yunkai domain; South China

33

34 **1. Introduction**

35 The Cathaysia and Yangtze blocks amalgamated in the Neoproterozoic (Fig. 1),
36 forming the South China Block (SCB), and its subsequent Phanerozoic tectonic
37 framework is an important constituent of tectonic evolution of Asia (Cawood et al.,
38 2013; Li et al., 2008a, b, 2009; Rong et al., 2010; Song et al., 2015; Zhang and Zheng,
39 2013; Zhang et al., 2011; Zhao and Cawood, 2012). Furthermore, as recognized since
40 1920's (Grabau, 1924), the Middle Devonian terrigenous rocks unconformably
41 covering Early Paleozoic folded rocks and granitoids, document an Early Paleozoic
42 tectono–magmatic event widespread in the SE part of the South China Block (Fig. 1).
43 This pre–Devonian orogeny has been improperly referred to as the "Caledonian
44 orogeny", however as precise time constraints are now available (Song et al., 2015;
45 Wang et al., 2007), and because this belt is unrelated to the true Caledonian belt of
46 Norway and N America, this term should be abandoned. The Early Paleozoic orogeny
47 of SE China is characterized by: i) the regional absence of Silurian strata, ii) the
48 unconformity between middle Devonian coarse clastic sequence and Ordovician
49 marine flysch sequence, iii) a greenschist to amphibolite facies metamorphism coeval
50 with a ductile deformation, and iv) the occurrence of numerous S–type granitic plutons
51 (BGMRFJ, 1985; BGMRGX, 1985; BGMRHN, 1988; BGMRJX, 1984; BGMRZJ,
52 1989; Charvet, 2013; Charvet et al., 2010; Faure et al., 2009; Li et al., 2010; Shu, 2012;
53 Shu et al., 2008b, 2015; Yao et al., 2013, 2014). The geodynamic significance of this
54 belt, whether as a collisional orogen or as an intracontinental one, has been debated
55 since a long time. Guo et al. (1989) argued for the presence of Early Paleozoic
56 ophiolites in the Yunkai and Wuyi massifs of the SE part of the South China Block.
57 However, in the past decades, several investigations indicated that the previously

58 proposed oceanic subduction and collision models do not properly account for the Early
59 Paleozoic lithological and tectonic features of this part of the SCB (Charvet et al., 2010;
60 Faure et al., 2009; Shu et al., 2014, 2015; Song et al., 2015; Wang et al., 2007, 2011,
61 2013b; Zhang et al., 2011). The main facts arguing against a collisional model are the
62 lack of ophiolites, accretionary complexes and magmatic arc. Indeed, ophiolitic
63 gabbros, once considered as Early Paleozoic, are in fact Neoproterozoic in age, ca. 850–
64 800 Ma (Li et al., 2005; Shu et al., 2006, 2011). Moreover, most of the "basaltic rocks"
65 interlayered in the Early Paleozoic strata have been reassessed as meta–greywacke (Shu
66 et al., 2008a, 2014). Granitoid is an efficient rock to understand the tectonic evolution
67 of the continental crust (e.g., Pearce et al., 1984; Pitcher, 1983). Some studies have dealt
68 with the Early Paleozoic granitoids of the South China Block. However, precise
69 chronological and geochemical data related to their petrogenesis are rare (e.g., Deng et
70 al., 2012; Li et al., 2010; Liu et al., 2010; Shu et al., 2008a; Wang et al., 2010, 2011;
71 Xia et al., 2014; Xu et al., 2011; Yang et al., 2010). This study presents zircon U–Pb
72 geochronology, bulk geochemistry and Hf isotope analysis of granitoids from the
73 Yunkai domain, providing new insights into the crustal evolution of the SE part of the
74 SCB in Early Paleozoic.

75

76 **2. Geological setting**

77 ***2.1 The general framework of SCB***

78 The southern part of the South China Block (Fig. 1) also referred to as “the South
79 China Fold Belt” is separated by the Neoproterozoic Jiangnan Orogen from the Yangtze
80 Block (Grabau, 1924; Guo et al., 1989; Li, 1997; Li et al., 2009, 2012; Shu and Charvet,
81 1996; Shu et al., 2006, 2014; Wang et al., 2007). This NE–SW trending belt is a
82 complex area that experienced several successive tectonic, metamorphic and magmatic
83 events, namely: i) a Neoproterozoic (Tonian–Cryogenian, ca. 850 Ma) collision; ii) a
84 Late Neoproterozoic (Cryogenian–Ediacarian, ca. 820–690 Ma) rifting event marked
85 by volcanoclastic sedimentation and bimodal magmatism; iii) an Early Paleozoic
86 orogeny marked by a Devonian angular unconformity, and iv) a Middle Triassic
87 orogeny marked by a Late Triassic unconformity (BGMRGX, 1985; BGMRJX, 1984;

88 [Faure et al., 2009](#); [Lin et al., 2008](#); [Shu, 2012](#)).

89 The Neoproterozoic Jiangnan orogenic belt in the southeastern margin of the
90 Yangtze Block is a NW–ward subduction–collision belt developed during the
91 amalgamation of the Cathaysia and Yangtze blocks ([Shu, 2012](#); [Yao et al., 2014](#)).
92 Ophiolite, I–type granite, rhyolite, basalt and gabbro, dated at ca. 1000–855 Ma, crop
93 out along the Shaoxing–Jiangshan–Guilin fault zone ([Li et al., 2009](#); [Shu, 2012](#); [Yao et
94 al., 2014](#)). The amalgamation resulted in the collision of island arcs with the Yangtze
95 Block and the closure of a back–arc basin. After the Middle Neoproterozoic orogeny,
96 from ca. 820 to 690 Ma, the South China block experienced a rifting event coeval with
97 a bimodal magmatism ([Li et al., 2003, 2005](#); [Shu, 2006](#); [Shu et al., 2011](#); [Wang and Li,
98 2003](#); [Wang et al., 2006](#)). In Cambrian times, the southern part of the rift was a littoral–
99 neritic depositional environment whereas during the early–middle Ordovician period,
100 this area was dominated by a neritic–bathyal setting ([Rong et al., 2010](#); [Shu et al., 2014](#)).
101 In the late Ordovician, the SE part of the South China Block underwent a
102 sedimentological change to a littoral silico–clastic environment, along with the
103 initiation of uplift processes ([BGMRFJ, 1985](#); [BGMRJX, 1984](#); [Shu, 2012](#); [Shu et al.,
104 2014](#)).

105 During the Silurian, extensive folding, thrusting, metamorphism and anatexis
106 developed. The emplacement of numerous granitic plutons represents the end of the
107 orogeny. These features are well recorded in the Wuyi, Jinggang, Nanling and Yunkai
108 areas ([Figs. 1 and 2](#)). The maximum shortening can reach up to 67% in the Jinggang
109 and Wuyi belts ([Charvet et al., 2010](#); [Shu, 2012](#); [Shu et al., 2008a, 2015](#)). Fold axes
110 strike are dominantly E–W. The kinematic analysis in the Jinggang and Wuyi domains
111 shows that the ductile shearing was directed to the S or SE, however, a northwestward
112 vergence may develop in the northwestern part of the belt. Concerning the Early
113 Paleozoic orogeny, two pre–Devonian litho–tectonic units have been identified: 1) A
114 slate unit, composed of the Sinianto Ordovician marine sandy–muddy rocks, which
115 experienced a low greenschist facies metamorphism before the intrusion of Silurian
116 granitoids; 2) A metamorphic unit, comprising Neoproterozoic mica schist, amphibolite,
117 paragneiss and orthogneiss, and locally Paleoproterozoic amphibolites, gneisses and

118 gneissic granites (Faure et al., 2009; Shu, 2012; Shu et al., 2014; Yu et al., 2009).
119 However, this orogeny is not well developed in the northwestern part of the South China
120 Block where the Early Paleozoic strata did not experience any significant
121 metamorphism, and only underwent slight brittle deformation, together with weak
122 magmatism. From a geodynamic point of view, the Early Paleozoic orogeny of the
123 South China Block was interpreted as the consequence of a rift closure from the Late
124 Ordovician to Early Silurian. Therefore, the orogeny corresponds to an intracontinental
125 event accommodated by the continental subduction of the southern part of the rift below
126 its northern one (Faure et al., 2009). It is worth to note that the rift closure area that
127 corresponds to a crustal scar instead of an ophiolitic suture does not coincide with the
128 Neoproterozoic suture but is located within the Cathaysia Block. The upper part of the
129 belt (i.e. the slate unit) is a fold-and-thrust belt limited at its base by a ductile
130 décollement localized in the Sinian system. The lower part of the belt (i.e. the
131 metamorphic unit underlying the basal décollement) was characterized by deep burial
132 giving rise to an amphibolite facies metamorphism (Zhao and Cawood, 1999). During
133 the exhumation of this lower part, those metamorphic rocks experienced retrogression
134 and partial melting at ca. 444–420 Ma represented by migmatite and granitoid (Faure
135 et al., 2009).

136 From the Middle Devonian to Early Carboniferous, quartz sandstone, feldspathic
137 sandstone, conglomerate, and siltstone intercalated with chert, limestone and bioclastic
138 limestone were deposited unconformably on the Early Paleozoic sequences (BGMRFJ,
139 1985; BGMRGX, 1985; BGMRJX, 1984; Shu et al., 2008b, 2015).

140 A Late Triassic regional unconformity implies a Middle Triassic orogeny (e.g.,
141 Chu et al., 2012; Faure et al., 2009; Lin et al., 2008; Shu, 2012; Wang et al., 2013b).
142 The Triassic orogens are widespread around the margins and inside the South China
143 Block. The Early Paleozoic metamorphic rocks were intensely reworked by the Triassic
144 events.

145 ***2.2 Geology of the Yunkai domain***

146 The NE trending Yunkai domain consists of a wide variety of plutonic,
147 sedimentary and metamorphic rocks (Fig. 1; BGMRGX, 1985; Faure et al., 2016; Lin

148 [et al., 2008](#)). In the study area, located in the NW part of the Yunkai domain, several
149 stratigraphic units are exposed: 1) a set of low greenschist facies and folded
150 Neoproterozoic sedimentary rocks, comprising slate, sandy slate and conglomerate; 2)
151 weakly metamorphosed and folded Early Paleozoic sequences mainly composed of
152 metasandstone and metamudstone; 3) Post–Ordovician sedimentary rocks ([Fig. 2](#)).

153 In the Yunkai domain, Early Paleozoic biotite–muscovite granite and
154 monzogranite are well exposed. These plutons are heterogeneously deformed with an
155 unfoliated core to a gneissic, sometimes mylonitic, margin ([BGMRGX, 1985](#)). In the
156 section, a progressive reduction in grain size from medium–coarse to fine grained can
157 be observed from the pluton core to margin. Meanwhile, Early and Middle Triassic
158 peraluminous plutons are also distributed in this area ([Lin et al., 2008](#); [Wang et al.,](#)
159 [2007](#)).

160 Our field observations indicate that the bulk architecture results of NE–SW and
161 E–W trending upright and overturned folds ([Fig. 2](#)) with NW and N vergences,
162 implying two tectonic deformation stages corresponding to Neoproterozoic and Early
163 Paleozoic, respectively ([Faure et al., 2016](#); [Lin et al., 2008](#)).

164 **3. Sample description**

165 Petrographic features and GPS locations of the analyzed samples are listed in
166 [Table 1](#). Sample 1571 is a fine to medium grained–massive two–mica granite composed
167 of ca. 40% K–feldspar, 35% quartz, 10% plagioclase, 7% muscovite, 5% biotite and
168 other accessories([Fig. 3A and 3F](#)) ([Table 1](#)). Sample 1582 of monzogranite is mainly
169 composed of 40% quartz, 35% K–feldspar, 15% albite, 5% muscovite and 5% other
170 accessory minerals ([Table 1](#)). It is located in the boundary of the pluton, displaying a
171 weak mineral fabric. Sample 1586 is a medium–grained massive K–feldspar muscovite
172 granite with 45% K–feldspar, 40% quartz, 8% plagioclase, 5% muscovite and other
173 accessories ([Fig. 3B and 3G](#)) ([Table 1](#)). From the pluton core to margin, a progressive
174 reduction in grain size from medium–coarse to fine–grained can be observed. Sample
175 1601 exhibits a distinct granitic texture, with about 35% K–feldspar, 30% quartz, 20%
176 biotite, 5~8% albite, 5% muscovite and other accessories ([Fig. 3C and 3H](#)) ([Table 1](#)).
177 Sample 1621 of the muscovite granite is mainly composed of 40% quartz, 38% K–

178 feldspar, 12% plagioclase, 7% muscovite and 3% other accessory minerals (Table 1),
179 polysynthetic twin and cross hatched twin are well developed in the plagioclase and
180 microcline, respectively (Fig. 3D). Sample 1628 of the biotite-rich proto-mylonitic
181 granite with feldspar augen shows about 40% K-feldspar, 30% quartz, 15~18% biotite,
182 10% plagioclase and other accessories (Table 1), in which the K-feldspar phenocrysts
183 are commonly deformed (Fig. 3E and 3I). Sample 1630 is a well-foliated gneissic
184 biotite granite composed of 45% microcline, 25% biotite, 15% quartz, 10% plagioclase
185 and 5% other accessory minerals (Table 1). Sample 1633 also shows a gneissosity, but
186 weaker than in Sample 1630. It is mainly composed of 40% microcline, 25% quartz,
187 20% biotite, 10% plagioclase and 5% other accessories. The dome-shaped granitic
188 pluton displays obvious zoning with gneissic granite in the middle, and mylonitic
189 granite around the margin.

190 In this study, five granite samples (1571, 1582, 1586, 1601, 1621), one mylonitic
191 granite (1628) and two gneissose granites (1630, 1633) were collected for zircon U-Pb
192 dating and Hf isotope analysis (see locations in Fig. 2). Fourteen rock samples (1571,
193 1571-1, 1582, 1586, 1586-1, 1601, 1601-1, 1601-2, 1601-3, 1621, 1621-1, 1628,
194 1630, 1633) were used for whole-rock geochemical study.

195

196 **4. Analytical methods**

197 Whole-rock major element contents were analyzed by ARL-9900 X-ray
198 fluorescence spectrometer (XRF) at the Testing Center of Shandong Bureau of China
199 Metallurgical Geology Bureau. The uncertainties reported in this study are 2% for
200 major elements. Trace elements and rare earth elements (REE) were measured by
201 Inductively Coupled Plasma- Mass Spectrometry (ICP-MS) (Finnigan Element II) at
202 the Testing Center of Shandong Bureau of China Metallurgical Geology Bureau and
203 ALS Chemex (Guangzhou) co., Ltd, respectively. International standards were used to
204 define the analytical precision and accuracy throughout the analytical processes for
205 ICP-MS. The uncertainties are 5 percent for trace elements. For the detailed analytical
206 procedure we refer to those documented in Gao et al. (2003).

207 Zircons were separated from the crushed rocks using heavy liquid and magnetic

208 techniques and then handpicked under a binocular microscope. The zircon grains were
209 mounted in epoxy resin and polished to expose the approximate center of grains, and
210 then coated with gold. Cathodoluminescence (CL) images of the zircons were obtained
211 using a JEOL JXA8230 electron probe microanalyzer at the Testing Center of Shandong
212 Bureau of China Metallurgical Geology Bureau.

213 The laser ablation (LA)–ICP–MS analysis of zircon U–Pb isotopic compositions
214 was performed at the State Key Laboratory for Mineral Deposits Research, Nanjing
215 University, using an Agilent 7500a ICP–MS system attached to New Wave 213 nm
216 laser ablation system with an in–house sample cell. Samples were analyzed in runs of
217 ca. 15 analyses including five zircon standards and 10 sample points. The details of the
218 analytical procedure are documented in [Jackson et al. \(2004\)](#). The U–Pb fractionation
219 was corrected using zircon standard GEMOC GJ–1 with a $^{207}\text{Pb}/^{206}\text{Pb}$ age of 601 ± 12
220 Ma and the accuracy was controlled using the zircon standard Mud Tank with an age
221 of 735 ± 12 Ma ([Black and Gulson, 1978](#)). The U–Pb ages were calculated from the raw
222 signal data using the software Glitter (ver.4.4). Because the ^{204}Pb could not be measured
223 owing to low signal and interference from ^{204}Hg in the gas supply, common lead
224 correction was carried out using the EXCEL program common Pb correction ([Andersen,](#)
225 [2002](#)). For zircons older than 1000 Ma, because of large amounts of radiogenic Pb, the
226 $^{207}\text{Pb}/^{206}\text{Pb}$ age is more reliable than $^{206}\text{Pb}/^{238}\text{U}$, whereas for zircons younger than 1000
227 Ma, as a result of the low content of radiogenic Pb and uncertainty of common Pb
228 correction, the $^{206}\text{Pb}/^{238}\text{U}$ age is more reliable ([Griffin et al., 2004](#)).

229 Zircon Hf isotopic composition was analyzed by Neptune MC–ICP–MS, which is
230 a double focusing multi–collector ICP–MS and has the capability of high mass
231 resolution measurements in a multiple collector mode. During laser ablation analyses,
232 the isobaric interference of ^{176}Lu on ^{176}Hf is negligible due to the extremely low
233 $^{176}\text{Lu}/^{177}\text{Hf}$ value in zircon (normally <0.002). However, the interference of ^{176}Yb on
234 ^{176}Hf must be intensively corrected since the contribution of ^{176}Yb to ^{176}Hf . This method
235 can provide an accurate correction of the ^{176}Yb interference on ^{176}Hf ([Kemp et al., 2006](#)).
236 During analysis, an isotopic ratio of $^{176}\text{Yb}/^{172}\text{Yb} = 0.5887$ was applied. Standard zircon
237 91500 was used for the external correction, with a $^{176}\text{Hf}/^{177}\text{Hf}$ value of 0.282300 ± 8

238 (2 σ). The detailed analytical procedure is similar to the description by Yuan et al. (2008).
239 Initial $^{176}\text{Hf}/^{177}\text{Hf}$ values were calculated based on Lu decay constant of $1.865\text{E}-11$
240 (Scherer et al., 2001). Model ages were calculated under the assumption that the
241 $^{176}\text{Lu}/^{177}\text{Hf}$ of average crust is 0.015, and the $^{176}\text{Hf}/^{177}\text{Hf}$ and $^{176}\text{Lu}/^{177}\text{Hf}$ ratios of
242 chondrite and depleted mantle at the present are 0.282772 and 0.0332, 0.28325 and
243 0.0384, respectively (Blichert-Toft and Albarede, 1997). The model ages (TDM)
244 provide only a minimum age for the source material of the magma from which the
245 zircons crystallized.

246

247 **5. Analytical results**

248 **5.1 Major and trace element compositions**

249 The analytical results of major and trace elements of 14 representative samples are
250 given in Table 2. Samples 1571, 1571-1, 1586, 1586-1, 1601-2, 1621, 1621-1, 1628,
251 1630, 1633 exhibit SiO_2 contents ranging from 69.91 to 77.46 wt%, and plot in the
252 granite field in the total alkali-silica diagram, displaying $(\text{Na}_2\text{O}+\text{K}_2\text{O})$ versus SiO_2
253 (TAS) (Fig. 4A). However, Samples 1582, 1601, 1601-1 and 1601-3 plot in the field
254 of quartz diorite, yielding SiO_2 contents between 65.03 and 69.67 wt%. As shown by a
255 plot of $\text{Al}/(\text{Na}+\text{K})$ versus $\text{Al}/(\text{Ca}+\text{Na}+\text{K})$ (Fig. 4B), all samples fall in the peraluminous
256 area with high A/CNK (>1.1), except Samples 1571, 1571-1 and 1586 which indicate
257 slightly lower A/CNK (1.08, 1.08 and 1.05, respectively). According to Figure 4C, most
258 of the samples belong to the high-K calc-alkaline affinity, while only Sample 1582
259 plots into the calc-alkaline field.

260 The most intuitive character of all samples analyzed for geochemistry is that they
261 display similar chondrite-normalized steep rare earth (REE) element patterns revealing
262 obvious enrichment in light rare earth element (LREE) with respect to heavy rare earth
263 element (HREE) (Fig. 5A). LREE/HREE values range from 2.81 to 30.36, and
264 $(\text{La}/\text{Yb})_N$ vary from 1.23 to 55.14 (the mean of 14 analyses is 14.69) (Fig. 5A, Table2).
265 All the samples mark distinct negative Eu anomalies (Eu/Eu^* value of 0.06-0.59, the
266 mean of 14 analyses is 0.4). On the primitive mantle-normalized spider diagrams (Fig.
267 5B), all samples exhibit strongly negative Ba, Sr and Nb anomalies and the enrichment

268 of Rb, Th, U and Pb. Nb/Ta values range from 3.76 to 23.22 (average 9.50), consistent
269 with the geochemical features of crustal derived granite (Corfu et al., 2003; Hoskin and
270 Schaltegger, 2003; Pearce, 1996; Pearce et al., 1984).

271 **5.2 Zircon U–Pb ages**

272 Typical Cathodoluminescence (CL) images of zircons are presented in Figure 7.
273 U–Pb results are listed in Table 3 and graphically illustrated in Figure 8. Zircons
274 analyzed in this study range in length from 60 to 160 μm , with length/width ratio ranges
275 from 2:1 to 4:1. Most of zircon grains display oscillatory zoning (Fig. 7) and high Th/U
276 values (average 0.80; Table 3), indicating their magmatic origin (Corfu et al., 2003;
277 Hoskin and Schaltegger, 2003).

278 Two hundred and twenty seven zircon U–Pb ages are obtained, which fall in the
279 range of 409–1306 Ma. However, most of the data indicate ages between 409 to 460
280 Ma. These data plotted on concordia diagrams yield eight groups of weighted mean
281 $^{206}\text{Pb}/^{238}\text{U}$ ages as $426 \pm 3\text{Ma}$ (Sample 1571, 19 values out of total 25 data); 439 ± 3
282 Ma (Sample 1582, 21 values out of total 32 data); $436 \pm 3\text{Ma}$ (Sample 1586, 18 values
283 out of total 25 data); $443 \pm 3\text{Ma}$ (Sample 1601, 18 values out of total 26 data); $430 \pm$
284 3Ma (Sample 1621, 23 values out of total 30 data); $435 \pm 3\text{Ma}$ (Sample 1628, 23
285 values out of total 33 data); $431 \pm 3\text{Ma}$ (Sample 1630, 19 values out of total 29 data)
286 and $429 \pm 3\text{Ma}$ (Sample 1633, 18 values out of total 27 data) (Fig. 8). The above results
287 indicate a Silurian age for these granitoids.

288 **5.3 In situ Hf isotopes**

289 More than half of the U–Pb dated zircons were chosen for in–situ Hf isotopic
290 analysis. The Hf analyses were executed near the fields used for U–Pb dating spots. For
291 purpose of discussing the Hf isotopic evolution history, the initial $^{176}\text{Hf}/^{177}\text{Hf}$ values
292 and $\epsilon\text{Hf}(t)$ were calculated using the zircon $^{206}\text{Pb}/^{238}\text{U}$ ages. The results of the Hf
293 isotopic analyses are presented in Table 4, and the $\epsilon\text{Hf}(t)$ versus U–Pb age diagram is
294 illustrated in Figure 9.

295 Lu–Hf isotopic results for the eight granitoid samples are similar. All samples
296 show negative $\epsilon\text{Hf}(t)$ values, ranging from -0.80 to -38.96. For the individual plutons,
297 the ranges are -4.44 to -17.51 (average of -7.60 for Sample 1571), -1.58 to -4.56

298 (average of -3.00 for Sample 1582), -3.00 to -38.96 (average of -15.42 for Sample 1586),
299 -0.80 to -25.17 (average of -11.06 for Sample 1601), -4.48 to -11.55 (average of -7.25
300 for Sample 1621), -1.70 to -15.20 (average of -7.92 for Sample 1628), -4.35 to -12.02
301 (average of -7.67 for Sample 1630), -5.06 to -10.68 (average of -6.99 for Sample 1633),
302 respectively. Correspondingly, on the $\epsilon\text{Hf}(t)$ versus U–Pb age plot, the two model ages
303 ($T_{\text{DM}2}$) mainly concentrate on 1481–2208 Ma, while sparsely distribute in 2208–3843
304 Ma (Fig. 9). These results indicate that all of the analyzed granitic rocks are derived
305 from the partial melting of Paleoproterozoic continental basement rocks. 14 groups of
306 two-model-age data greater than 2.3 Ga may indicate that the magma has been
307 contaminated by ancient crustal material to a certain extent. In addition, all the isotopic
308 data are plotted under the CHUR line, suggesting that the involvement of a mantle
309 component in the granitic magma was negligible.

310

311 **6. Discussion**

312 **6.1 Crystallization age of the granitoids**

313 The 227 zircons from all eight granitoid samples from the Yunkai domain yield
314 similar weighted mean U–Pb ages between 426 and 443 Ma (Fig. 8, Table 3), indicative
315 of a Silurian crystallization age. Many works have been undertaken to investigate the
316 formation age of the Early Paleozoic granites in the South China Block (Li et al., 2010;
317 Shu et al., 1999, 2008a, 2008b; Song et al., 2015; Wang et al., 2007, 2013b; Zhang et
318 al., 1998; Zhao et al., 2013; Zhong et al., 2013). For example, the granitoids for the
319 Xuefeng, Guidong, Zhuguang, Wugong and Wuyi Mountains display zircon U–Pb ages
320 of 410–430 Ma, 422–427 Ma, 414–434 Ma, 428–462 Ma, 431–441 Ma, respectively
321 (Chu et al., 2012; Li, 1990, 1994; Lou et al., 2006; Xu et al., 2011) (Fig. 1). Also,
322 $^{40}\text{Ar}/^{39}\text{Ar}$ age can be utilized to trace the evolution process of the plutons (Kay et al.,
323 2011; Maksiyev et al., 2004; McDougall and Harrison, 1999; Reynolds et al., 1981). Xu
324 et al. (1992) reported two synkinematic phengites from the mylonites that yield
325 $^{40}\text{Ar}/^{39}\text{Ar}$ plateau age of 428 Ma from the Dexing–Shexian shear zone in the northern
326 margin of the Yangtze Block. Wang et al. (2007) showed that the synkinematic phengite
327 of the mylonitic rocks from the Nanfeng–Yingtian shear zone at the northern margin of

328 the northern Wuyi domain yields $^{40}\text{Ar}/^{39}\text{Ar}$ plateau at 423–426 Ma. In the Jiuling Mts,
329 biotite and muscovite from mylonites yield $^{40}\text{Ar}/^{39}\text{Ar}$ weighted plateau ages ranging
330 from 379 Ma to 468 Ma (Chu and Lin, 2014). Lastly, Shu et al. (2015) indicated that
331 the $^{40}\text{Ar}/^{39}\text{Ar}$ data on newly grown biotites from deformed K–granites and biotite schist
332 occurring along the Shaoxing–Pingxiang fault zone show pseudoplateau ages of 428–
333 433 Ma. These $^{40}\text{Ar}/^{39}\text{Ar}$ ages reflect the regional deformation time that can be
334 interpreted as the contemporary period of the granitic magmatism. In addition, the
335 timing of the magmatic activities can be constrained by geological occurrence. To the
336 east of Hezhou, Cambrian strata are intruded by granite (section A–B in Fig. 2).
337 Moreover, the Silurian and lower Devonian sequences are absent in this area, along
338 with the pre–Devonian strata unconformably covered by the middle Devonian sequence.
339 According to the geological literatures (BGMRGX, 1985; Faure et al., 2009; Shu et al.,
340 2008a; Wang et al., 2007), it was probably related to a doming induced by thermal event.
341 Therefore, the field observations support the view that the Early Paleozoic magmatic
342 event in the Yunkai domain occurred during the Silurian.

343

344 ***6.2 Source and petrogenesis of the Early Paleozoic granitoids***

345 A number of studies have addressed the origin and petrogenesis of the Early
346 Paleozoic granites in the South China block (e.g., Shu et al., 1999, 2015; Wang et al.,
347 2007, 2010, 2011; Zhang et al., 2011; Zhao et al., 2013). Song et al. (2015) showed that
348 Early Paleozoic granites collected from the Jiangnan Orogen and the eastern part of the
349 SCB display $\varepsilon_{\text{Hf}}(t)$ values of -1.44 to -36.84 with two stage model ages of Hf isotope
350 range from 1.3 Ga to 3.7 Ga, which indicates that the Early Paleozoic granites were
351 derived from crustal source without or with a small input of mantle materials.
352 Additionally, the Early Paleozoic granites from Xuefeng, Jinggang and Yunkai domains
353 show Sr–Nd isotopic compositions with initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratios of 0.70924–0.72935,
354 negative $\varepsilon_{\text{Nd}}(t)$ values of -4.7 to -11.5 and Proterozoic Nd model ages (Wang et al.,
355 2013b, and references therein). The results of this study indicate that the samples
356 collected from the Yunkai area also support crustal sources for the genesis of the
357 granitic magmas. Geochemically, the granite samples are K–feldspar megacryst rich,

358 and display high total alkali and alumina contents, which all plot in the peraluminous
359 field (Fig. 4B). Geochemically, in the tectonic discrimination diagrams (Fig. 6), most
360 of the granitoid samples plot in the post–collision field. The geochemical signatures
361 indicate that the Early Paleozoic granitic magmas represent typical anatectic products
362 of continental crust. They were generated during the Silurian post–orogenic collapse
363 event, developed after the intracontinental thickening event. Isotopically, all 125
364 isotopic data are under the CHUR line (Fig. 9). This feature agrees with the melting of
365 a continental basement with little or no input of juvenile crust (Song et al., 2015). Two–
366 stage continental crust model ages of the analyzed zircon grains show a concentration
367 of 1481–2208 Ma ages (Fig. 9), which reveals that some Paleoproterozoic rocks
368 probably exist beneath the South China Block. Moreover, 13 groups of $\epsilon\text{Hf}(t)$ values
369 less than -15 reveal that Archean material may exist beneath the South China Block or
370 not far away.

371 Recently, some mafic rocks have been dated at 409–434 Ma. The Taoyuan
372 hornblende gabbro complex, located to the north of the Wugongshan domain, was
373 formed by magma mixing and mingling, whose mafic member originated from a
374 metasomatized lithospheric mantle during the Silurian (Zhong et al., 2013). These
375 authors interpreted this rock to be generated from the post–orogenic collapse stage of
376 an intracontinental tectonic regime. Moreover, several gabbroic plutons have been
377 identified in the Yunkai area (Wang et al., 2013a). It is proposed that the generation of
378 the source of the mafic magmas was located within Mesoproterozoic to Early
379 Neoproterozoic asthenospheric mantle, metasomatized by paleosubduction–modified
380 wedge derived fluids during asthenospheric upwelling in the Silurian, resulting in the
381 melting of the continental lithospheric mantle. Similar mechanism is well inspired by
382 de Jong et al. (2015) and references therein. According to the above lines of evidence,
383 it seems that the granitic magmas in this study area were generated by partial melting
384 of continental crust without contribution of juvenile crustal components.

385 Generally, the granites can be used to explore petrogenesis and tectonic settings
386 (Abdallah et al., 2007; Castro, 2014; Foden et al., 2015; John and Blundy, 1993; Shu et
387 al., 2008a; Song et al., 2015; Vigneresse, 2014; Wang et al., 2007). Field observation

388 shows that the granite structures were gradually changed from the center to the margin
389 of the plutons. Foliation characterized by the preferred orientation of quartz, feldspar
390 and mica is common at the pluton margin (Fig. 3). Regional-scale folding during the
391 Early Paleozoic was widespread together with extensive thrusting and ductile shearing.
392 Various fold types, such as overturned, chevron, recumbent, and sheath, thrust sheets,
393 and mylonitic shear zones are widely developed in the pre-Devonian units (Charvet et
394 al., 2010; Li et al., 2010; Shu et al., 1997; Xu et al., 2011; Yao et al., 2011). The intense
395 folding and shortening accommodated by intracontinental subduction might raise the
396 temperature of parent rocks (Shu, 2006; Song et al., 2015; Zhang et al., 2011).
397 Furthermore, additional heat would be provided by radioactive crustal elements
398 concentrated during the ductile deformation. Thus the heat budget would trigger partial
399 melting of the continental crust able to generate the Early Paleozoic granitic magma.
400 Also, dehydration melting of the hydrous minerals, such as micas, probably plays
401 important roles in the partial melting process. After its burial, the thickened crust was
402 quickly eroded and recovered its normal thickness. The metamorphosed continental
403 rocks were gradually exhumed and decompressed under an extensional tectonic setting,
404 resulting in the rise of the granitic magma and its emplacement in the upper crust (Faure
405 et al., 1996).

406

407 ***6.3 Tectonic setting of the South China Block in the Early Paleozoic***

408 A large-scale tectono-magmatic event took place in the southeastern South China
409 Block during the Silurian (Charvet et al., 2010; Faure et al., 2009; Shu et al., 2014;
410 Wang et al., 2007). But the tectonic setting of the Early Paleozoic magmatism is still
411 controversial. Guo et al. (1989) proposed that a subduction zone was developed along
412 the Zhenghe-Dapu section in the Sinian-Late Ordovician period, resulting in the arc-
413 continent collision and the formation of a large-scale fold belt. According to the
414 geochemical features of the granites in the Yunkai belt, Peng et al. (2006, 2016a, 2016b)
415 suggested an ocean-continent subduction-collision and a post-collisional extension-
416 delamination-underplating model for the South China Block in the Early Paleozoic.
417 Similarly, a subduction-related mechanism in the northern margin of the Yunkai

418 domain during the Early Paleozoic was assumed by [Qin et al. \(2013\)](#) owing to the arc
419 magmatic geochemical characters of the Hudong plutonic complex.

420 Nevertheless, the field geological evidences are always the first hand and most
421 important information concerning the tectonic setting. The Ordovician orogenic event
422 (ca. 470–460 Ma), referred to as the Grampian Phase in the British Isles or Taconian
423 Phase in western New England ([McKerrow et al., 2000](#)), is generally accepted as being
424 due to collision of the rifted Laurentian margin with a continent facing arc. The model
425 was put forward on the basis of relatively clear field evidence of deformation, ophiolite
426 obduction and stratigraphy in the orogen ([Dalziel, 1997](#); [Dewey and Shackleton, 1984](#);
427 [Dewey and Ryan, 1990](#); [Van Staal et al., 1998](#)). Comparing to the classical collisional
428 belts, however, evidence for oceanic basin, ophiolite suite and subduction complex
429 associated with an Early Paleozoic subduction is absent in the South China Block.
430 Contemporaneous arc volcanics and high pressure metamorphic rocks have so far never
431 been documented in South China. Turbidites with Bouma sequences are also lacking in
432 spite of widely developed thick flysch sedimentary successions ([BGMRFJ, 1985](#);
433 [BGMRGX, 1985](#); [BGMRJX, 1984](#); [Wang et al., 2010](#); [Shu, 2012](#); [Shu et al., 2014](#)).
434 Consequently, it is unreasonable to interpret the Early Paleozoic tectonics of South
435 China as related to an oceanic subduction and a subsequent collision. On the contrary,
436 an intracontinental orogeny accommodated by a continental subduction is likely (e.g.,
437 [Charvet et al., 2010](#); [Chen et al., 2010](#); [Faure et al., 2008](#); [Shu et al., 2008a](#); [Song et al.,](#)
438 [2015](#)).

439 In the present study, the lithofacies paleogeography of South China provides
440 constraints on the source direction of thick siliciclastic sediments in the Southeast basin
441 that were mainly derived from the southeast, such northwestward sedimentary transport
442 is also supported by several palaeocurrent indicators preserved in the Cambrian-
443 Ordovician strata, indicating that a paleoplate probably existed to the southeast of SE
444 SCB ([Ren, 1964](#); [Ren et al., 1990](#); [Rong et al., 2010](#); [Shu, 2012](#); [Shu et al., 2008a](#); [Wu](#)
445 [et al., 2010](#)). Regional-scale folding during the Early Paleozoic is widespread in the
446 South China Fold Belt ([BGMRFJ, 1985](#); [BGMRGX, 1985](#); [BGMRJX, 1984](#); [Jahn et](#)
447 [al., 1990](#); [Shu et al., 1991](#); [Song et al., 2015](#)). Moreover, kinematic studies in the South

448 China Block show a fan-shaped thrust pattern, with top-to-the-northwest in the
449 northwestern part of SE South China Block and top-to-the-southeast in the
450 southeastern part of SE South China Block (Faure et al., 2009; Lin et al., 2008, 2011;
451 Shu et al., 2014). Such double-vergent thrusting may develop in intraplate settings,
452 associating with syntectonic magmatism, like Petermann and Alice Springs belts in
453 central Australia (Haines et al., 2001; Raimondo et al., 2009, 2010, 2014; Wade et al.,
454 2005) or the Middle Triassic Xuefengshan belt in the central part of South China (Chu
455 et al., 2012).

456 It is widely accepted that the SE SCB has experienced crustal thickening and
457 anatexis in the Early Paleozoic (Wang et al., 2010; Zeng et al., 2008). Continent-
458 continent collision may cause thickening of the continental crust over a large wide zone
459 (1,000 km), with far-field continental shortening, and last for 50 million years. Such
460 zones are characterized by a wide orogenic plateau, with surrounding and internal
461 basins bordered by thrust belts (Dewey, 2005).

462 According to the available geological investigations, therefore, we propose that
463 the Early Paleozoic tectono-magmatic event in SE SCB is probably related to the
464 continent-continent collision. A Suspected East China Sea Block subducted
465 northwestward beneath the SE SCB together with the NW SCB subducted
466 southeastward beneath the SE SCB (Fig. 10D), resulting in the crustal thickening and
467 intracontinental deformation which may be caused by the stress generated at plate
468 boundaries and stress transmission with the lithosphere acting as an effective stress
469 guide (Aitken et al., 2013; Gorczyk and Vogt, 2015; Raimondo et al., 2014).
470 Subsequently, regional extension occurred in response to crustal thickening at an earlier
471 stage of the orogeny (Dewey, 1988; Strachan, 1994). The clockwise metamorphic P-T
472 paths in Chencai, Wuyi and Yunkai regions indicate that the SE SCB experienced
473 isothermal decompression (Li et al., 2010; Wang et al., 2012; Zhao and Cawood, 2012),
474 representing the progressive release of stress and rapid crustal denudation after the
475 crustal thickening and uplift. Coincidentally, the Sr-Nd isotopic compositions of Early
476 Paleozoic granites from the South China Block are similar to those of Caledonian,
477 Hercynian tectonic belts and other classical collisional belts (Jahn, 2004; Jahn et al.,

478 2014), which probably represent the similar dynamic mechanism during the
479 decompression and remelting of the crustal sources.

480

481 *6.4 Geodynamic evolution of the South China Block from the Neoproterozoic to* 482 *Early Paleozoic*

483 A possible spatial and temporal geodynamic evolution model accounting for the
484 Neoproterozoic to Early Paleozoic geological events is proposed in the following.

485 The Jiangnan orogen is a collisional belt developed during the assembly of the
486 South China block (Charvet et al., 2010; Li et al., 2002; Shu et al., 2011; Yan et al.,
487 2015; Yao et al., 2012; Zhao and Cawood, 2012). Previous studies have shown that the
488 Cathaysia and Yangtze blocks collided at ca. 860 Ma along the Shaoxing–Pingxiang–
489 Guilin fault zone (Shu et al., 2015; Yao et al., 2014) (Figs. 1 and 10A). During the Late
490 Neoproterozoic, a regional–scale rifting event affected the South China Block, resulting
491 in the development of fault–bounded basins. The bimodal volcanic rocks
492 contemporaneous to the rifting were dated at 820–690 Ma (Wang and Li, 2003) (Fig.
493 10B). During the Cryogenian–Ediacarian (Sinian) to Early Paleozoic, from 690 to 460
494 Ma, the southeastern part of the South China Block was in a depositional stage.
495 Stratigraphic sequences in this period are characterized by graptolite bearing sandy–
496 muddy rock associations, overlying the rift deposits. Proximal deposits, coarse grain
497 sandstone, conglomerate and ripple–mark structures suggest a littoral–shelf–slope
498 depositional environment (Shu, 2012; Shu et al., 2014) (Fig. 10C).

499 During the Late Ordovician and Early Silurian, the tectonic framework of South
500 China was mainly controlled by a regional compression characterized by km–scale
501 folds and thrusts, the development of southeast–directed décollement shearing,
502 amphibolite facies metamorphism. During the late to post–thickening evolution, crustal
503 melting characterized by migmatites and granitic plutons took place (Charvet et al.,
504 2010; Faure et al., 2009; Shu, 2006, 2012; Shu et al., 2008a, 2015; Song et al., 2015;
505 Wang et al., 2013b; Zhang et al., 2011). These structural, metamorphic and magmatic
506 features were related to the northwest–ward continental subduction of the southeastern
507 part of the Neoproterozoic rift. This continental area can be considered as a part of the

508 inferred East China Sea Block proposed since a long time (Ren, 1964; Ren et al., 1990)
509 (Fig. 10D). Subsequently, during the Devonian, the South China Block was in a stable
510 sedimentary environment.

511

512 **7. Conclusions**

513 According to the zircon geochronology, and geochemical study of granitoids in
514 the NW part of the Yunkai domain, we reach the following conclusions:

515 (1) The analyzed plutons mostly crystallized during 426–443 Ma, suggesting an
516 Early Paleozoic magmatic event in the Yunkai domain;

517 (2) All granites are peraluminous and represent typical anatectic products of
518 continental crust, mostly plot in the post–collision field in tectonic
519 discrimination diagrams. Hf isotopic data indicate that parent magma was
520 derived from Paleoproterozoic–Mesoproterozoic crustal components with
521 little or no input of mantle sources;

522 (3) The intracontinental orogeny already proposed to account for the Early
523 Paleozoic plutonic event in the South China Block is well supported by our
524 new investigations.

525

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536

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938 **Figure captions**

939 **Figure 1.** Tectonic outline of China and the distribution map of Early Paleozoic granites
940 in the South China Block (Shu et al., 2015; Song et al., 2015), the numbers 1–5 are
941 corresponding to the Xuefeng, Guidong, Zhuguang, Wugong and Wuyi, respectively.

942 **Figure 2.** Geological sketch map of the Yunkai domain with cross sections A–B and C–
943 D (BGMRGX, 1985).

944 **Figure 3.** Representative field photos and photomicrographs. (A) Photomicrograph of
945 Sample 1571 (two–mica granite) in Yonghe of Lianshan county; (B) Photomicrograph
946 of Sample 1586 (muscovite granite) in Xindi of Wuzhou city.; (C) Photomicrograph of
947 Sample 1601 (two–mica granite) in Songguangling of Cenxi city; (D) Photomicrograph
948 of Sample 1621 (muscovite granite) in Beijie of Xinyi city; (E) Photomicrograph of
949 Sample 1628 (biotite mylonitic granite) in Lingshan of Rongxian county; (F) Field
950 photograph of Sample 1571; (G) Field photograph of Sample 1586; (H) Field
951 photograph of Sample 1601; (I) Field photograph of Sample 1628. Mineral
952 abbreviations: Kfs, K–feldspar; Mc, microcline; Pl, plagioclase; Ms, muscovite; Bt,
953 biotite; Qtz, quartz.

954 **Figure 4.** Geochemical features of the granitoids from the Yunkai domain. (A)
955 (Na₂O+K₂O) versus SiO₂ diagram (Cox et al., 1979); (B) Al/(Na+K) versus
956 Al/(Ca+Na+K) diagram (Maniar and Piccoli, 1989); (C) SiO₂ versus K₂O diagram
957 (Rollison, 1993).

958 **Figure 5.** Distribution of (A) rare earth elements and (B) trace elements for the samples
959 derived from the Yunkai domain. The normalization values for (A) and (B) are cited
960 from Sun and McDonough (1989) and McDonough and Sun (1995), respectively.

961 **Figure 6.** Tectonic discrimination diagrams for the Early Paleozoic granites from the
962 Yunkai domain. (A) Rb versus (Y+Nb) diagram (after Pearce, 1996; Pearce et al., 1984);
963 (B) Rb versus (Yb+Ta) diagram (after Pearce, 1996; Pearce et al., 1984).

964 **Figure 7.** Cathodo-luminescence images of zircons from the granitoids of the Yunkai
965 domain, attached with analyzed locations and U–Pb ages.

966 **Figure 8.** U–Pb concordia plots for the zircon grains from eight granitoid samples.

967 **Figure 9.** Plot of Epsilon Hf(t) versus U–Pb age of zircons from eight granitoid samples.

968 **Figure 10.** Geodynamic evolution models for the South China Block from
969 Neoproterozoic to Early Paleozoic. (A) Collision of Yangtze and Cathaysia blocks (Shu,
970 2012; Yao et al., 2014); (B) Rifting stage of the Yangtze–Cathaysia Block (Shu, 2012;
971 Yao et al., 2014); (C) Stable depositional stage of the SE South China Block (Ren, 1964;
972 Shu et al., 2014, 2015); (D) Intraplate Orogeny stage of the SE South China Block (Shu,
973 2012; Shu et al., 2014, 2015). SCB: South China Block. Depth is not in proportion.

974

975 **Tables**

976 **Table 1** GPS locations of the samples and petrological descriptions.

977 **Table 2** Major and trace element data for representative samples from the Yunkai
978 domain.

979 **Table 3** U–Pb data of the zircons from eight granitoid samples from the Yunkai domain.

980 **Table 4** Hf isotope analyses of the zircons from eight granitoid samples from the Yunkai
981 domain.