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1 **Early Cretaceous continental arc magmatism in the**
2 **Wakhan Corridor, South Pamir: mantle evolution**
3 **and geodynamic processes during flat subduction of**
4 **the Neo-Tethys oceanic slab**

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28 **ABSTRACT**

29 The petrogenesis of continental arc magmas provide critical insights into thermal
30 evolution and geodynamics of the continental lithosphere and crust-mantle interaction
31 and deep dynamic processes. In this study, we report new zircon U-Pb ages along with
32 isotopic and elemental whole-rock geochemistry, mineral chemistry and Hf-O isotope
33 data, for Kalaqigu diorites and monzogranites in the Chinese Wakhan Corridor, South
34 Pamir. Zircon U-Pb dating indicates that the Kalaqigu pluton was emplaced in the
35 Early Cretaceous (ca. 108–106 Ma). The diorites are geochemically characterized by
36 low SiO₂ (51.9–54.5 wt.%) and CaO (7.7–9.4 wt.%) contents, but high MgO (5.3–8.3
37 wt.%), Al₂O₃ (12.8–16.8 wt.%) and TiO₂ (0.6–1.1 wt.%) contents as well as high Mg[#]
38 (56–65) values, and so are similar to high-Mg diorites. They are enriched in large ion
39 lithophile elements (LILEs, e.g., K, Sr and Ba) and light rare earth elements (LREEs),
40 while depleted in high field strength elements (HFSEs, i.e., Nb, Ta, Zr and Hf).
41 Combined with negative $\epsilon_{Nd}(t)$ (-6.9 to -14.0), $\epsilon_{Hf}(t)$ (-9.9 to -12.2) and high
42 $(^{87}Sr/^{86}Sr)_i$ (0.7075–0.7086) ratios, these observations indicate they originated from an
43 enriched lithospheric mantle source. High $\delta^{18}O_{zm}$ (7.49–9.01‰) values in conjunction
44 with relatively high $^{207}Pb/^{206}Pb$ and $^{208}Pb/^{206}Pb$ ratios suggest that the source was
45 modified by subducted sediment-derived melts. Variable Cr contents (54–117 ppm)
46 are likely controlled by minor fractionation of olivine and orthopyroxene. The
47 monzogranites show high SiO₂ contents (69.2–72.0 wt.%), low Rb/Sr (0.4–0.6),
48 $(K_2O+Na_2O)/CaO$ (2.6–4.8) and FeO^T/MgO ratios (2.6–3.2). They contain diagnostic
49 cordierite and show strongly-peraluminous characteristics ($A/CNK > 1.1$) with high
50 $\delta^{18}O_{zm}$ (7.82–8.85‰) values, compatible with typical S-type granites. Their abundant
51 inherited zircons, with age populations similar to those of detrital zircons from
52 regional Early Paleozoic metasedimentary rocks, indicate they were derived from
53 partial melting of ancient metasedimentary rocks. Phase equilibrium modelling is
54 consistent with biotite-dehydration melting of metagreywacke, probably at ~750 °C
55 and ~6.0 kbar indicated by the biotite chemistry. A south-to-north magmatic migration
56 based on regional geochronology suggests that northward flat-slab subduction of the
57 Neo-Tethys oceanic slab played an important role in the generation of these

58 widespread Early Cretaceous continental arc magmatic rocks. However, the granitoids
59 were generated earlier than the mantle-derived mafic rocks, which suggests that
60 crustal melting occurred during the early stage of subduction. The subsequent flat
61 subduction resulted in continuous metasomatism by subducted sediments.
62 Contemporaneous regional compression primarily occurred far north of the
63 subduction zone (i.e., North and Central Pamir), inducing deformation as well as
64 crustal shortening. With the flare-up of continental arc magmatism in South Pamir,
65 crustal shortening moved southward. These processes, combined with addition of
66 voluminous mantle-derived magmas, played an important role in crustal thickening in
67 the Pamir during the Early Cretaceous.

68

69 **INTRODUCTION**

70 Andean-type continental arcs, where the oceanic crust subducts beneath an active
71 continental margin, serve as ideal natural laboratories for studying destructive plate
72 margin processes and the geodynamic evolution of the Earth's crust and mantle (e.g.,
73 Rudnick and Gao, 2003; Davidson and Arculus, 2006; Jagoutz and Schmidt, 2012).
74 Consequently, the formation and evolution of Andean-type continental arcs is a topic
75 of considerable interest in earth sciences (e.g., Ducea et al., 2015; Suo et al., 2019;
76 Qin et al., 2022). As a result of the subduction of oceanic slabs, continental arcs
77 generally have undergone intense mantle metasomatism and crust-mantle interaction,
78 resulting in arc-related igneous suites ranging in compositions from mafic to felsic
79 (e.g., Dhuime et al., 2012; Ducea et al., 2015; Cashman et al., 2017; Tang et al., 2019;
80 Xiao et al., 2022). A major focus is the spatiotemporal and petrogenetic connections
81 between the high-Mg basic to intermediate rocks generated from a modified mantle
82 and the coeval felsic granitoids (Castro, 2019). For instance, I-type granites within
83 "Cordilleran" batholiths can represent fractionated liquids from intermediate magma
84 systems of broadly high-Mg andesitic composition (e.g., Castro et al., 2010), while
85 sanukitoid magmas can act as water donors that trigger extensive melting of the lower
86 crust to also generate granitoids, for example from "Cimmerian" batholiths of Iberia
87 (e.g., Pereira et al., 2015). Subduction processes within continental arc systems are

88 invariably complicated, and may include flat-slab subduction (e.g., Zhang et al., 2022),
89 slab rollback (e.g., Ma et al., 2013; Lei et al., 2023) or oceanic ridge subduction (e.g.,
90 Zhu et al., 2019; Ma et al., 2022). The geochemical characteristics of
91 contemporaneous subduction-related magmatism (Rudnick and Gao, 2003; Zhu et al.,
92 2019; Qi et al., 2023) reflect this range of processes (e.g., Gianni and Luján, 2021; Ma
93 et al., 2023). Therefore, ancient subduction-related arc magmatism contains key
94 information for understanding the characteristics of subduction.

95 The South Pamir experienced northward subduction of the Neo-Tethys Ocean
96 from the Shyok suture (e.g., Schwab et al., 2004; Faisal et al., 2016; Chapman et al.,
97 2018a), and is regarded as a preeminent natural laboratory for studying oceanic
98 subduction geodynamics and Andean-style orogenesis. Subduction-related rocks in
99 South Pamir are important components (e.g., Faisal et al., 2016; Aminov et al., 2017;
100 Zhang et al., 2022), and led to the formation of a typical continental arc (e.g., Schwab
101 et al., 2004). The Wakhan Corridor (located within Karakorum Batholith, South Pamir)
102 preserves complete continental arc-related magmatic sequence, including mafic
103 igneous rocks (e.g., basalt, andesite, diorite) with granitoids (e.g., granodiorite and
104 monzogranite) (Jiang et al., 2014; Li et al., 2016; Liu et al., 2019; Zhang et al., 2022).
105 Their field relations, age distribution, petrology and geochemistry provided valuable
106 information for understanding Neo-Tethys subduction. Based on the distance (ca. 200
107 km) from the Shyok suture to the South Pamir batholith, as well as the Early
108 Cretaceous active shortening of the Pamir (Robinson, 2015), Cretaceous low-angle
109 flat subduction of the Neo-Tethys oceanic slab has been proposed (e.g., Chapman et
110 al., 2018a; Zhang et al., 2022). Although many studies have been devoted to the
111 magmatic rocks of the Wakhan Corridor, their origin is still controversial. In particular,
112 the processes of mantle metasomatism derived from Neo-Tethys slab subduction are
113 still unclear, as is the petrogenesis and evolution of related crust-derived magmas.

114 In this paper, we present new U-Pb geochronological, geochemical, and
115 Sr-Nd-Pb-Hf-O isotope data from the Cretaceous diorites and monzogranites of the
116 Kalaqigu pluton, east of Wakhan Corridor. Our primary focus is to investigate their
117 petrogenesis and elucidate their tectonic setting. By integrating the new results with

118 existing data from the Wakhan Corridor, we aim to further constrain the geodynamic
119 processes of Neo-Tethys subduction during the Cretaceous and contribute to the
120 global understanding of continental arc magmatism.

121

122 **GEOLOGICAL SETTING AND SAMPLING**

123 **Geological Setting**

124 The Pamir, which borders the Tianshan, Tarim Basin and Tibetan Plateau, has
125 undergone prolonged processes of accretion, collision, and suturing during the Early
126 Paleozoic-Mesozoic (Fig. 1a; e.g., Tapponnier et al., 1981; Robinson et al., 2012;
127 Robinson, 2015). The Cenozoic collision between the Indian and Asian continents
128 resulted in further deformation, structural overprinting, and tectonic uplift of Pamir,
129 forming its remarkable present-day configuration (e.g., Yin and Harrison, 2000; Liu et
130 al., 2017; Rutte et al., 2017). Geologically, the Pamir region can be divided into three
131 tectonic units, from north to south: North Pamir, Central Pamir and South Pamir,
132 separated by Paleozoic and Mesozoic sutures (Fig. 1b; Burtman and Molnar, 1993;
133 Angiolini et al., 2013; Robinson, 2015). The North Pamir represents a Paleozoic
134 accretionary complex (Schwab et al., 2004; Robinson et al., 2012), while the Central
135 Pamir and South Pamir are believed to have rifted from the northern margin of
136 Gondwana during the Late Carboniferous to Early Permian (Burtman and Molnar,
137 1993; Angiolini et al., 2013; Robinson, 2015).

138 The North Pamir contains abundant Paleozoic and Permo-Triassic magmatic
139 rocks (Fig. 1b; Chapman et al., 2018a), while Central Pamir hosts mainly Cenozoic
140 plutons, with a few of Cretaceous age (Fig. 1b; Robinson, 2015; Chapman et al.,
141 2018a; Ma et al., 2023). The South Pamir can be further divided into southwest Pamir
142 and Karakoram terrane, separated by a system of Cenozoic extensional detachments
143 (Fig. 1b; Schwab et al., 2004; Schmidt et al., 2011; Angiolini et al., 2013). The
144 southwest Pamir, which continues into the Hindu Kush mountain range, consists of
145 Precambrian basement domes and records a complex history of later magmatism
146 spanning from the Early Paleozoic to the present day (Schmidt et al., 2011; Faisal et
147 al., 2016; Soret et al., 2019). In the southeast Pamir, Permian to Cretaceous

148 sedimentary sequences dominate, accompanied by highly deformed and
149 metamorphosed Precambrian basement (Zanchetta et al., 2018; Zhang et al., 2018;
150 Imrecke et al., 2019). In addition to this, widespread Early Cretaceous magmatic
151 rocks are observed in the South Pamir. This magmatic suite is dominated by S-type
152 and I-type granitoids in addition to a small amount of intermediate-silicic volcanic
153 rocks (e.g., Faisal et al., 2016; Aminov et al., 2017; Ma et al., 2023), with a notable
154 high-flux magmatic event occurring between 110 and 105 Ma (Fig. 1b; Schwab et al.,
155 2004; Heuberger et al., 2007; Chapman et al., 2018a). These Cretaceous magmatic
156 rocks provide important insights into Mesozoic subduction of the Neo-Tethys Ocean
157 and the crust-mantle interaction (Ravikant et al., 2009; Zhang et al., 2022). To the
158 south, the Kohistan-Ladakh Arc, situated between the Shyok Suture (SSZ) and the
159 Indus-Tsangpo Suture Zone (ITSZ), developed during the Mesozoic above a
160 north-dipping subduction zone within the Neo-Tethys Ocean (Fig. 1b; Schwab et al.,
161 2004; Ravikant et al., 2009; Jagoutz and Schmidt, 2012; Chapman et al., 2018a).

162 The diorites and monzogranites investigated in this study are exposed in the
163 Kalaqigu pluton located in the Chinese Wakhan Corridor, which is part of the South
164 Pamir (Figs. 1b–1c). The basement of the Wakhan Corridor comprises the Archean (ca.
165 2.5 Ga) Mazar complex. This was intruded by ca. 840 Ma granites and 500–490 Ma
166 mafic rocks as well as Early Cretaceous mafic-to-acidic intrusions (Ji et al., 2011; Li
167 et al., 2016; Zhang et al., 2018; Zhang et al., 2022). As a result of Neo-Tethys oceanic
168 subduction, these Cretaceous magmatic rocks have formed complete volcanic
169 sequences related to arc activity, creating an east-west trending magmatic belt along
170 the Wakhan Corridor (Li et al., 2016; Liu et al., 2020; Zhang et al., 2022). Jiang et al.
171 (2014) first reported the granodiorite and monzogranites of the Kalaqigu pluton and
172 suggested that they were generated by partial melting of metasedimentary basement.
173 They also reported that a mafic basalt-andesite sequence with enriched Nd isotope
174 compositions (-5.9 to -9.6) (Zhang et al., 2022) and dioritic enclaves with $\epsilon_{Nd}(t)$ of ca.
175 -4.74 (Liu et al., 2020) erupted at 104–98 Ma in the west of the studied area. These
176 originated from metasomatized sub-continental lithospheric mantle and underwent
177 variable assimilation fractional crystallization. I-type granitoids generated by

178 crust-mantle interaction are also reported from the Wakhan Corridor (Liu et al., 2020).
179 Additionally, in the northwest of the Wakhan Corridor, the contemporaneous (110–92
180 Ma) Teshiktash-Beik volcanic basin consists of grey and reddish lavas, tuffaceous
181 lavas, and dacitic tuff breccias (Fig. 1b; Aminov et al., 2017). To the north, S-type
182 granites were emplaced in the Taxkorgan pluton, in the northeastern part of the
183 continental arc (Fig. 1b; Jiang et al., 2014; Ma et al., 2023).

184 In the field, the Kalaqigu pluton intrudes Paleozoic and Jurassic strata and is
185 adjacent to the Hongqilapu pluton (Fig. 1c). Granitoids derived from crust-mantle
186 mixing as well as mantle-derived dioritic dikes have also been identified in the
187 Hongqilapu pluton (Jiang et al., 2014; Li et al., 2016).

188

189

190 **Sampling and Description**

191 Diorite and monzogranite were collected from the Kalaqigu pluton in the
192 Chinese part of the Wakhan Corridor, Southern Pamir, as indicated in Figure 1c. The
193 diorite exhibits a medium-fine grained subhedral granular structure and is primarily
194 composed of plagioclase (60 vol.%), amphibole (25 vol.%), biotite (10 vol.%), and
195 quartz (1–5 vol.%), with accessory minerals such as apatite and zircon (Figs. 2a–2b).
196 The biotite shows partial replacement by chlorite (Fig. 2b). The monzogranite exhibits
197 a fine-grained or porphyritic texture and is composed of K-feldspar (35 vol.%),
198 plagioclase (30 vol.%), quartz (25 vol.%), biotite (1–5 vol.%), cordierite (1 vol.%)
199 and sillimanite (< 1 vol.%), with minor amounts of apatite and zircon within the
200 matrix (Figs. 2c–2d). Partially enclosed K-feldspar grains contain small biotite
201 inclusions (Figs. 2c–2d). Metasomatic worm-like structures can be observed at the
202 contacts between plagioclase and K-feldspar and sillimanite appear as small needles
203 predominantly embedded in the K-feldspar. Cordierite is pseudomorphed by serpentine
204 (Fig. 2d).

205

206 **ANALYTICAL METHODS**

207 Detailed analytical methods are given in Supplementary File 1 and a short

208 summary is given here. Zircon U-Pb dating using LA-ICP-MS and in-situ zircon
209 Lu-Hf isotopes were analyzed at the Institute of Geology and Geophysics, Chinese
210 Academy of Sciences (IGG CAS). Zircon oxygen isotopes were determined by
211 secondary ion mass spectrometer (SIMS) at the at State Key Laboratory of Isotope
212 Geochemistry (SKLaBIG), Guangzhou Institute of Geochemistry, Chinese Academy
213 of Sciences (GIG CAS). Major elements analyses of amphibole, plagioclase and
214 biotite were carried out using JEOL JXA-8100 Electron Probe Micro Analyzer
215 (EPMA) at IGG CAS. Major element contents were determined by X-ray
216 fluorescence spectrometer (XRF) on fused glass beads at SKLaBIG, GIG CAS. Trace
217 element analyses were determined using a Perkin-Elmer ELAN-DRC-e inductively
218 coupled plasma mass spectrometer at the State Key Laboratory of Ore Deposit
219 Geochemistry (SKLOG). Whole-rock Sr-Nd-Pb isotopes were determined using a
220 MC-ICP-MS at SKLaBIG, GIG CAS.

221

222 **RESULTS**

223 Zircon was separated from two samples for LA-ICP-MS zircon U-Pb dating.
224 Whole-rock major and trace elements and Sr-Nd-Pb isotopes, zircon U-Pb-Hf-O
225 isotopic compositions as well as analysis of mineral compositions are provided in the
226 Supplementary Table 1-5.

227

228 **Zircon U-Pb Ages**

229 All zircon crystals from Kalaqigu diorite and monzogranite are generally
230 transparent and colorless, with length/width ratios ranging from 4:1 to 1:1 (Fig. 3).
231 They exhibit well-developed concentric oscillatory zoning in cathodoluminescence
232 (CL) images (Fig. 3) and high Th/U ratios (typically > 0.2; Supplementary Table 1),
233 indicating a magmatic origin (Belousova et al., 2002). A few zircon grains from the
234 monzogranite display a clear core-rim structure in CL images (Fig. 3d). Eighteen
235 analytical spots from the diorite yielded concordant $^{206}\text{Pb}/^{238}\text{U}$ ages ranging from 103
236 to 109 Ma, with a weighted mean age of 105.9 ± 0.3 Ma (MSWD = 0.93) (Fig. 3a).
237 Dating of monzogranite was conducted on the zircon rims and cores (Figs. 3b–3d).

238 The zircon rims yield $^{206}\text{Pb}/^{238}\text{U}$ ages ranging from 106 to 120 Ma, with weighted
239 mean ages of 108.4 ± 0.4 Ma ($n = 7$, MSWD = 0.59) (Fig. 3b). These ages indicate
240 that Kalaqigu pluton was emplaced during the Early Cretaceous (108–105 Ma). Ages
241 for the inherited zircon cores from the monzogranite exhibit a wide range, from 1032
242 to 334 Ma (Supplementary Table 1; Figs. 3c–3d).

243

244 **Mineral Chemistry**

245 Amphibole is the dominant mafic mineral in the Kalaqigu diorites but is absent
246 from the monzogranite. Based on the classification by Leake et al. (1997), these
247 belong to the calcic subgroup ($\text{Ca}_B = 1.83\text{--}1.92$ and $(\text{Na} + \text{K})_A = 0.33\text{--}0.59$) with high
248 $\text{Mg}^\#$ values (52–62). Amphibole phenocrysts from WK1616 that yield $(\text{Na} + \text{K})_A <$
249 0.5 based on amphibole classification diagrams classify as magnesio-hornblende (Fig.
250 4a, Leake et al., 1997; Hawthorne and Oberti, 2007). The amphibole phenocrysts with
251 higher content of alkaline elements ($(\text{Na} + \text{K})_A \geq 0.5$) classify as magnesio-hastingsite
252 ($^{\text{VI}}\text{Al} < \text{Fe}^{3+}$) and pargasite ($^{\text{VI}}\text{Al} > \text{Fe}^{3+}$) (Fig. 4b). Only one amphibole has a high
253 content of Si (> 6.5) and classifies as edenite (Fig. 4b). To estimate the temperature
254 and pressure of amphibole crystallization, we used the formulas from Ridolfi et al.
255 (2010) and the empirical geobarometer from Krawczynski et al. (2012). Calculated
256 temperatures and pressures for these amphibole grains range from 771–907 °C and
257 3.26–7.43 kbar with average of 849 °C and 5.16 kbar, respectively (Supplementary
258 Table 2). We also calculated the $\text{H}_2\text{O}_{\text{melt}}$ contents (5.19–6.86 wt.%) and oxygen
259 fugacity ($\Delta\text{FMQ} = 0.45\text{--}1.89$) using the formulas from Ridolfi et al. (2010).

260 Biotite from the diorites is mainly Mg-biotite, with relatively high concentrations
261 of SiO_2 (37.3–37.4 wt.%), MgO (12.2–12.5 wt.%) and $\text{Mg}^\#$ (55), as well as low Al_2O_3
262 (14.1–14.5 wt.%), FeO^{T} (17.8–18.0 wt.%) and TiO_2 (3.73–3.85 wt.%) (Fig. 4c,
263 Supplementary Table 2). In the monzogranite, biotite is Fe-biotite with lower SiO_2
264 (34.4–35.4 wt.%), MgO (6.66–7.37 wt.%) and $\text{Mg}^\#$ (36–40), as well as high Al_2O_3
265 (18.5–19.2 wt.%), FeO^{T} (20.0–21.2 wt.%) and TiO_2 (3.65–4.02 wt.%) (Fig. 4c,
266 Supplementary Table 2) than the biotite in the diorites. Calculated temperature and
267 pressure values of biotite from diorites show a range of 770–783 °C and 4.56–5.32

268 kbar, whereas those of biotite from monzogranites have a range of 746–772 °C and
269 5.50–6.42 kbar (Li and Zhang, 2022; Supplementary Table 2), representing the
270 crystallization T and P of biotites.

271 Representative plagioclase grains were analyzed from core to rim (Figs. 4b–4c).
272 For diorite (WK1616), plagioclase grains typically show a decrease in modal anorthite
273 (An) from core to rim (Fig. 4d; Supplementary Table 2). These grains generally have a
274 core of $An_{80-85}Ab_{15-20}Or_{0-1}$ and a rim of $An_{45-50}Ab_{45-55}Or_{1-2}$. This indicates that
275 plagioclase crystallized from bytownite, through labradorite to andesine from early to
276 later stages (Fig. 4d). Plagioclase compositions in monzogranite (WK1617) vary
277 between An_{22} – An_{35} and classify as oligoclase-andesine (Fig. 4d).

278

279 **Major and Trace Element Geochemistry**

280 ***Diorites***

281 The diorites are characterized by low SiO_2 (51.9–54.5 wt.%) (Fig. 5a), high MgO
282 (5.4–8.3 wt.%), and $Fe_2O_3^T$ (8.1–9.5 wt.%) contents with high $Mg^\#$ (56–65) values.
283 They have relatively high alkalis ($K_2O + Na_2O = 4.31$ – 5.35 wt.%) and K_2O/Na_2O
284 (0.9–1.3) ratios, showing high-K calc-alkaline to shoshonite (Fig. 5b) and potassic
285 features (Fig. 5c). In addition, they exhibit high Al_2O_3 (12.8–16.8 wt.%) and CaO
286 (7.73–9.35 wt.%) contents, plotting in metaluminous field ($A/CNK = 0.6$ – 0.8).

287 On chondrite-normalized rare earth element (REE) diagrams (Fig. 6a), the
288 diorites show sub-parallel light-REE enriched patterns with relatively flat heavy REEs
289 ($(La/Yb)_N = 6.4$ – 12.1 ; $(Gd/Yb)_N = 1.7$ – 2.3), with weak negative Eu anomalies
290 [$Eu/Eu^* (Eu_N/\sqrt{Sm_N \times Gd_N}) = 0.77$ – 0.88]. On a primitive mantle normalized
291 diagram, they are enriched in LILEs (K, Sr and Ba) but depleted in HFSEs and show
292 negative Nb, Ta, P, Zr, Hf, and Ti anomalies (Fig. 6b). The diorite samples are
293 characterized by high Sr (365–607 ppm) and Y (20.9–25.8 ppm) contents with low
294 Sr/Y (17–27) ratios. In addition, they have high Cr (54.6–117.2 ppm), Co (18.6–33.7
295 ppm), and Ni (7.84–34.6 ppm) contents.

296

297 **Monzogranites**

298 The monzogranites have relatively silicic compositions ($\text{SiO}_2 = 69.2\text{--}72.0$ wt.%)
299 (Fig. 5a). They are characterized by low $\text{Fe}_2\text{O}_3^{\text{T}}$ (1.8–2.9 wt.%), MgO (0.6–1.0 wt.%;
300 $\text{Mg}^{\#} = 36\text{--}40$), CaO (1.6–2.6 wt.%), as well as high Al_2O_3 (14.4–16.2 wt.%) and
301 alkalis ($\text{K}_2\text{O} + \text{Na}_2\text{O} = 6.8\text{--}7.7$ wt.%) contents. They show calc-alkaline to high-K
302 calc-alkaline (Fig. 5b), potassic (Fig. 5c) and peraluminous ($\text{A/CNK} = 1.15\text{--}1.23$)
303 features (Fig. 5d).

304 The monzogranites show steeply fractionated REE patterns (Fig. 6c), with
305 marked enrichment in LREEs and steep HREEs ($(\text{La/Yb})_{\text{N}} = 41.4\text{--}60.7$; $(\text{Gd/Yb})_{\text{N}} =$
306 $6.0\text{--}7.2$), plus moderate negative Eu anomalies ($\text{Eu/Eu}^* = 0.67\text{--}0.83$). On a primitive
307 mantle normalized diagram, they also show enrichment in LILEs (such as Th, Rb, U,
308 and K) relative to HFSEs and LREEs, with negative Ba, Nb, Ta, Sr, P, and Ti
309 anomalies (Fig. 6d). The monzogranite samples have relatively low Sr (205–293 ppm)
310 and Y contents (9.1–10.2 ppm) with Sr/Y ratios of 24 to 30. In addition, they exhibit
311 low Cr (2.17–2.91 ppm), Co (1.68–2.87 ppm), and Ni (0.83–1.47 ppm) contents.

312

313 **Whole-rock Sr-Nd-Pb Isotopes**

314 Initial Sr-Nd-Pb isotopic values were calculated at the relevant crystallization
315 age (106 or 108 Ma). The Kalaqigu diorites have initial $^{87}\text{Sr}/^{86}\text{Sr}$ isotopic ratios of
316 0.7075 to 0.7086 and $\varepsilon_{\text{Nd}}(t)$ values of -5.97 to -7.18 with old Nd model ages ($T_{\text{DM}} = \text{ca.}$
317 $1491\text{--}1592$ Ga) (Fig. 7a). The studied monzogranites exhibit $\varepsilon_{\text{Nd}}(t)$ values (-12.3 to
318 -12.5) with two-stage model ages of 1909 to 1921 Ma and initial $^{87}\text{Sr}/^{86}\text{Sr}$ isotopic
319 ratios (0.7154–0.7158), but these differ significantly from those of the diorites (Fig.
320 7a). The diorites and monzogranites have similar $(^{206}\text{Pb}/^{204}\text{Pb})_i$ ratios (18.64–18.68 in
321 diorites and 18.66–18.67 in monzogranites) but different $(^{207}\text{Pb}/^{204}\text{Pb})_i$ (15.70–15.71
322 and 15.757–15.758, respectively) and $(^{206}\text{Pb}/^{204}\text{Pb})_i$ ratios (39.02–39.12 and 39.14–
323 39.16, respectively) relative to the Northern Hemisphere Reference Line (NHRL)
324 (Figs. 7b–7c).

325

326 **Zircon Hf-O Isotopic Compositions**

327 The zircons from the Kalaqigu diorites (WK1616) have negative $\epsilon_{\text{Hf}}(t)$ values
328 ranging from -9.9 to -12.2 (-10.9 on average) and old T_{DM2} (1.79–1.93 Ga and 1.85
329 Ga on average) ages (Supplementary Table 4). The Kalaqigu diorites (WK1616) and
330 monzogranites (WK1617) have similar $\delta^{18}\text{O}$ values (i.e., 7.49–9.01‰ and 7.82–
331 8.85‰, respectively; Supplementary Table 5), which are higher than those ($5.3 \pm$
332 0.6‰, 2SD) of igneous-origin zircons in equilibrium with mantle-derived magmas
333 (Fig. 8; Valley, 2003).

334

335 **DISCUSSION**

336 **Geochemical Affinities**

337 The Kalaqigu diorites exhibit high MgO, Al₂O₃ and TiO₂ contents, but low
338 FeO^T/MgO ratios (1.0–1.4) and CaO contents, thus showing geochemical affinities to
339 high-Mg andesites (HMAs; Tatsumi, 2001). In addition, they have low Sr (365–607
340 ppm), Y (20.9–25.8 ppm) and Yb contents (1.9–2.5 ppm) with low Sr/Y (17–27) and
341 (La/Yb)_N ratios (11–15), which are compositionally analogous to those of sanukite
342 from the Setouchi Volcanic Belt (Figs. 9a–9b; Yogodzinski et al., 1994; Shimoda et al.,
343 1998; Tatsumi, 2001). Their high Mg[#] values and Cr contents also suggest that they
344 are sanukitic HMAs (Kamei et al., 2004; Martin et al., 2005).

345 Granites are often categorized as S-, I-, or A-types (Chappell and White, 1974;
346 Loiselle and Wones, 1979; Whalen et al., 1987). The monzogranites from the
347 Kalaqigu pluton are hornblende-free and characterized by high SiO₂, low Rb/Sr,
348 (K₂O+Na₂O)/CaO (2.6–4.8) and FeO^T/MgO ratios (2.6–3.2), implying that they are
349 unfractionated granites (Chappell and White, 1974, 1992). They have low 10000 ×
350 Ga/Al ratios (2.1–2.6) and (Zr + Nb + Ce + Y) contents (199–257) and low zircon
351 saturation temperatures (761–824°C), which are distinct from those of A-type granites
352 (Figs. 9c–9d; Whalen et al., 1987). Their A/CNK values are higher than 1.1 (Fig. 5d),
353 and they contain diagnostic peraluminous minerals such as cordierite (Fig. 2d;
354 Barbarin, 1999). The U-Pb ages of inherited zircon cores from the monzogranite show
355 large variations (Fig. 3c), consistent with typical S-type granitoids (Gao et al., 2016).
356 In addition, they have initial ⁸⁷Sr/⁸⁶Sr ratios of 0.7154 to 0.7158 and low $\epsilon_{\text{Nd}}(t)$ values

357 of -12.3 to -12.4 (Figs. 7a–7b). These geochemical features strongly indicate that
358 these monzogranites are unfractionated, high-K, S-type granites.

359

360 **Petrogenesis and Magma Sources**

361 *Diorites*

362 *AFC effect*

363 As mentioned above, the Kalaqigu diorites show affinities with HMAs. These
364 high-Mg diorites have low SiO₂ contents and high Mg[#] values (Fig. 10a) with variable
365 Cr and Ni contents, indicating they are derived from a mantle source. For
366 mantle-derived melts, wall-rock assimilation and shallow-level fractional
367 crystallization are inevitable during transport and emplacement (DePaolo, 1981).
368 Crustal assimilation is likely to cause an increase in (⁸⁷Sr/⁸⁶Sr)_i and a decrease in ε_{Nd}(*t*)
369 as well as Nb/La and Nb/Th ratios. This is because the continental crust is typically
370 characterized by low ε_{Nd}(*t*) values, low Nb/La and Nb/Th ratios, but high (⁸⁷Sr/⁸⁶Sr)_i
371 ratios relative to those of the mantle (Rudnick and Fountain, 1995). However, these
372 high-Mg diorites show virtually identical Sr-Nd isotopic compositions (Fig. 7a) and
373 limited variations in Nb/La (0.31–0.42) and Nb/Th (0.64–2.51) ratios as well as zircon
374 ε_{Hf}(*t*) values (-12.2 ~ -9.9). These observations suggest that they have undergone
375 limited crustal assimilation.

376 These high-Mg diorites however, have generally low but variable compatible
377 element contents, such as V, Cr and Ni, indicating some fractionation of mafic
378 minerals. The positive correlation between MgO and Fe₂O₃^T and CaO as well as
379 negative correlation between MgO and Al₂O₃ suggests fractionation of
380 olivine/pyroxene/spinel in a lower magma chamber (Figs. 10b-10d). The negative
381 correlation of MgO and TiO₂ suggests there is no fractionation of Fe-Ti oxides (Fig.
382 10f). Absence of Eu, Sr and Ba anomalies also suggests that feldspar was probably not
383 a fractioning phase (Figs. 6a–6b). These fractional crystallization trends are consistent
384 with models derived from Rhyolite-MELTS using a pressure of 5.16 kbar and water
385 content of 5.84 wt.% (Supplementary Table 2), as well as a range of oxygen fugacity

386 (Δ QFM +1, +2 and +3) (Figs. 10b–10f). The models support minor fractionation of
387 olivine + orthopyroxene +/- clinopyroxene/spinel during magma evolution.

388

389 *Magma Evolution*

390 Several petrogenetic models have been proposed for the formation of HMAs,
391 including (1) partial melting of a subducted oceanic crust with assimilation of mantle
392 peridotites (Yogodzinski et al., 1994; Kelemen, 1995); (2) partial melting of
393 delaminated mafic crust at mantle depths (Chen et al., 2013); (3) direct partial melting
394 of hydrous mantle (Hirose, 1997; Wood and Turner, 2009; Mitchell and Grove, 2015);
395 (4) interaction of fluids/melts derived from subducted slabs and/or sediments with the
396 overlying mantle wedge (Shimoda et al., 1998; Tatsumi, 2001, 2006).

397 In general, HMAs generated by models one and two are similar to adakitic and
398 bajaitic HMAs because their trace element characteristics, such as high Sr/Y ratios,
399 absence of Eu anomalies and depletion in HREEs (Kelemen, 1995; Kelemen et al.,
400 2004; Chen et al., 2013), reflect the melting residue of oceanic slab or lower crust.
401 However, the Kalaqigu high-Mg diorites have low Sr/Y ratios (Fig. 9b) with only
402 slight negative Eu anomalies (Fig. 6a) and they are enriched in LREEs with limited
403 depletion in HREEs (Fig. 6b). Coupled with their high Al₂O₃ (12.8–16.8 wt.%), Sc
404 contents (24.5–30.0 ppm) and low Y (20.9–25.8 ppm) and Yb (1.9–2.5 ppm), this
405 geochemical signature suggests a garnet-free source region (Defant and Drummond,
406 1990; Hoskin and Schaltegger, 2003; Macpherson et al., 2006). Further, it is thought
407 that the crust of South Pamir did not significantly delaminate during the Early
408 Cretaceous (Soret et al., 2019), therefore this also rules out the second model.
409 Accordingly, and given the following observations, we consider that these high-Mg
410 diorites to represent the products of interaction between subducted sedimentary melts
411 and hydrous mantle:

412 (1) The Kalaqigu high-Mg diorites have low $\epsilon_{\text{Hf}}(t)$ values (-12.2 to -9.91) that are
413 lower than those of HMA formed by direct melting of hydrous mantle (Wood and
414 Turner, 2009). Their enriched Sr-Nd-Pb isotopic compositions are also inconsistent
415 with simple partial melting of a mantle source (Fig. 7).

416 (2) They have significant depletions in Nb ($\text{Nb/La} = 0.3\text{--}0.4$), Ta, Ti, Zr and Hf
417 (Fig. 6b), the quintessential signature of subduction-related rocks (e.g., Defant and
418 Drummond, 1990; Hawkesworth et al. 1997a). Associated high LILEs (e.g., Rb, Sr
419 and Ba; Supplementary Table 3), and high LREEs (Fig. 6a) require enrichment of the
420 mantle source before partial melting. Both these elemental characteristics are likely to
421 be related to subduction of Neo-Tethys oceanic lithosphere (e.g., Ravikant et al., 2009;
422 Liu et al., 2020; Zhang et al., 2022).

423 (3) The moderate Rb/Ba and Rb/Sr ratios of these high-Mg diorites plot along
424 the mixing trend between basalt and a calculated pelite-derived melt (Fig. 11a),
425 indicating that their magma source was most likely a mixture between basalt and
426 sedimentary rocks. The samples also display a positive correlation between Th/La and
427 Th/Sm (Fig. 11b), consistent with simple binary mixing (i.e., between DMM and
428 GLOSS).

429 (4) Subduction fluids generally carry LILEs (e.g., Rb, Sr and Ba) and other
430 fluid-mobile trace elements (e.g., U and Pb) into the mantle wedge (Hawkesworth et
431 al., 1997a, b). However, the Kalaqigu high-Mg diorites show low Ba/La (17–32),
432 Ba/Th (51–167), Sr/Th (39–120) and U/Th (0.2–0.4) ratios, indicating that their
433 source was not significantly metasomatized solely by a fluid component (Figs. 11c–
434 11d; Hawkesworth et al., 1997a; Turner et al., 1997). In contrast, addition of sediment
435 is likely to increase La/Sm and Th/Yb (Hanyu et al., 2006; Tatsumi, 2006; Labanieh et
436 al., 2012). All samples yield $(\text{La/Sm})_N$ ratios of 2.4–3.4 and Th/Yb ratios of 1.6–6.5,
437 consistent with the addition of subducted sediments (Figs. 11c–11d), but their partial
438 melting is indicated by higher Th/La and Th/Sm ratios than those in the GLOSS (Fig.
439 11b). The indicative model curve suggests that the degree of partial melting of
440 GLOSS is ~4%, and the proportion of GLOSS melt in the high-Mg diorites is ~10%
441 (Fig. 11b).

442 (5) The Sr-Nd-Pb isotopic compositions of the diorites plot within or close to the
443 GLOSS field, also consistent with a subduction-related enrichment (Fig. 7). The
444 diorites also have zircon $\delta^{18}\text{O}$ values which are markedly higher than that of the
445 depleted mantle (Fig. 8a), likely inherited from their sedimentary source components

446 (Valley, 2003; Workman et al., 2005).

447 To summarize, we infer that the Kalaqigu high-Mg diorites were generated by
448 partial melting of enriched mantle modified by subducted sediment-derived melts.

449

450 *Mantle Source and Magma Evolution*

451 As mentioned above, the Kalaqigu high-Mg diorites originated from a
452 metasomatized mantle modified by subducted sediments. When high-pressure silicate
453 liquids from the subducted crust are out of equilibrium with the overlying mantle
454 rocks, they will interact with the mantle, resulting in hybrid silicate-carbonate melt
455 compositions (Sekine and Wyllie, 1982; Wyllie and Sekine, 1982; Ionov et al., 1997).
456 These hybrid melts may be consumed by reaction with the overlying mantle rocks,
457 creating metasomatized domains (e.g., Sekine and Wyllie, 1982; Wyllie and Sekine,
458 1982). The reactions of carbonate-rich melts with overlying peridotite would produce
459 lherzolites and harzburgites (e.g., Lambart et al., 2012). In general, these lherzolites
460 and harzburgites would be in equilibrium with an aluminous mineral which changes
461 from plagioclase to spinel to garnet with increasing pressure (Wyllie, 1979; Müntener
462 and Ulmer, 2018).

463 These diorites are characterized by only small negative Eu anomalies (Fig. 6a)
464 and moderate fractionation of HREE (Fig. 6b), supporting the absence of plagioclase
465 and garnet as residual phases. Instead, the patterns are consistent with the presence of
466 spinel in the mantle source (Guo et al., 2006). As shown in Figure 12a, Dy/Yb and
467 La/Yb systematics also indicate that the diorites need an enriched mantle source for
468 spinel-harzburgite mantle partial melting. In addition, their high K₂O contents (2.1–
469 3.0 wt.%; Fig. 5b) imply that K-rich phases including phlogopite and/or K-rich
470 amphibole as residual phases occur in the mantle source. However, these diorites have
471 low Rb/Sr (0.1–0.3) and Ba/Rb (4.4–11.0) ratios that suggests the presence of residual
472 phlogopite (Fig. 12b; Furman and Graham, 1999). Phlogopite is a common
473 metasomatic volatile-bearing K-rich phase (e.g., Sekine and Wyllie, 1982; Wyllie and
474 Sekine, 1982) and its consumption usually results in high K₂O contents. Therefore,
475 we propose that the Kalaqigu high-Mg diorites were generated from a

476 phlogopite-bearing spinel-harzburgite mantle and have undergone fractionation of
477 olivine, orthopyroxene and spinel within a lower magma chamber, which is similar to
478 the petrogenesis of sanukitic HMAs (Tatsumi, 2001; Wang et al., 2008).

479 Petrography demonstrates that amphibole occurs as inclusions in feldspar (Fig.
480 2a), indicating amphibole was an early crystallizing phase. Biotite and quartz occur as
481 late interstitial crystallizing phases. Thus, the inferred sequence of crystallization is
482 amphibole → plagioclase → biotite + quartz. A previous experimental study has
483 shown that high H₂O contents (> 3%) would suppress plagioclase and lead to earlier
484 crystallization of amphibole (Müntener et al., 2001). Plagioclase crystallization later
485 than amphibole indicates that the primitive parental melts may have had high water
486 contents. This is also supported by the frequent presence of hydrous minerals
487 (amphibole and biotite), high H₂O_{melt} and oxygen fugacity values calculated by
488 amphibole compositions (Supplementary Table 2). Mantle hybridization by influx of
489 sediments and/or partial melts derived from them can directly form pargasitic
490 amphibole (Mandler and Grove 2016). Such amphibole can be stable over a wide
491 range of mantle pressures up to 4 GPa and temperatures of about 1000–1100 °C
492 (Mandler and Grove 2016). In subduction zones, the downgoing slab commonly
493 undergoes multistage dehydration and melting, which continuously releases fluids and
494 melts to metasomatize the mantle wedge (Sekine and Wyllie, 1982; Wyllie and Sekine,
495 1982). The initial fluids generated by dehydration progressively evolve and are
496 accommodated in the fugitive aqueous phases during heating and solidification of the
497 magma (Wyllie and Sekine, 1982). Thus, the mantle source region may have been
498 metasomatized by fluids prior to the generation of these diorites. This is consistent
499 with the proposal that sanukitic magmas may contain important amounts of dissolved
500 water (Castro, 2020).

501 As mentioned above, Early Cretaceous magmatic rocks are widely distributed in
502 Chinese Wakhan Corridor (Jiang et al., 2014; Li et al., 2016; Zhang et al., 2022).
503 Among them, mantle-derived magmatic rocks, including basalt-andesite (ca. 100–98
504 Ma; Zhang et al., 2022) and diorite (ca. 104–100 Ma; Li et al., 2016; Liu et al., 2020),
505 crystallized marginally later than high-Mg diorites in this paper (ca. 106 Ma). Despite

506 this, they exhibit similar Sr-Nd isotopic compositions (Fig. 7a) and are also thought to
507 have been generated from an enriched hydrous mantle source (Liu et al., 2020; Zhang
508 et al., 2022). However, as shown on Fig. 11, the mantle source region of these
509 younger rocks requires more sediment input (i.e., subducted sediments and/or
510 sediments from overlying crust; Li et al., 2016; Liu et al., 2020). Given the similar
511 mantle sources (i.e., phlogopite-bearing spinel-harzburgite mantle; Figs. 12a–12b)
512 over ca. 10 Ma, the metasomatism most likely occurred at a stable depth (<100 km;
513 Wyllie, 1979; Klemme and O'Neill, 2000). During this period (ca. 106–98 Ma), the
514 interaction between partial melts of subducted sediments as well as crustal materials
515 and mantle wedge persisted below the Chinese Wakhan Corridor and led to a
516 relatively uniform, stable enriched mantle source.

517

518 *Monzogranites*

519 *Origin of the inherited zircons*

520 The Kalaqigu S-type monzogranites contain inherited zircon cores with ages of
521 334–1032 Ma (Figs. 3c–3d). Several alternative mechanisms can explain how these
522 inherited zircons became incorporated into the Early Cretaceous monzogranites. They
523 could potentially be captured from the crustal country rocks during emplacement of
524 granitic magmas. However, there is no obvious evidence of crustal assimilation
525 because no xenoliths of the country rocks have been observed in the studied S-type
526 monzogranites. In addition, whole-rock Sr-Nd isotopes of these monzogranites show
527 relatively limited variations (Fig. 7) and do not change with increasing SiO₂ contents.
528 It is thus unlikely that these inherited zircons were captured from the local continental
529 crust, rather were inherited from crustal sources.

530 The monzogranites have high $\delta^{18}\text{O}$ values of 7.82 to 8.85‰ (Fig. 8b), which are
531 significantly higher than those of igneous zircons from lower crust-derived magmas
532 (5‰–7.5‰; Valley et al., 2005), but close to those of sedimentary rocks ($\delta^{18}\text{O} > 8$ ‰;
533 Valley et al., 2005). In addition, their inherited zircon cores show variable U-Pb ages
534 that are consistent with the ages of detrital zircons in southeastern Pamir (Fig. 3c;
535 Imrecke et al., 2019). Therefore, the geochemical characteristics indicate that the

536 studied S-type monzogranites were most likely generated by melting of Early
537 Paleozoic sedimentary rocks (Imrecke et al., 2019).

538

539 *Melting mechanism and characteristics of the magma source*

540 As mentioned above, the Kalaqigu monzogranites have unfractionated S-type
541 granites affinities, with high SiO₂ contents, low Mg[#] values (Fig. 10a), as well as
542 negative ε_{Nd}(*t*) (-12.3 to -12.5). Combined with their peraluminous geochemical
543 signature (Fig. 5d), these characteristics are consistent with a metasedimentary source.
544 In general, the geochemical characteristics of S-type granites are controlled not only
545 by the properties of metasedimentary rocks in the source area, but also by both the
546 extent and mechanism of melting (Patiño Douce and Harris, 1998; Patiño Douce,
547 1999; Hopkinson et al., 2017). On the one hand, the crustal metasedimentary source
548 can be divided into clay-rich metapelite, and clay-poor metagraywacke (e.g.,
549 Sylvester, 1998). The Kalaqigu S-type samples exhibit higher CaO/Na₂O ratios (0.49–
550 0.63) relative to melts derived from metapelite (CaO/Na₂O ratios < 0.5; Jung and
551 Pfänder, 2007). In addition, they have low Rb/Sr and Rb/Ba ratios, suggesting they
552 were derived mainly from clay-poor metagreywacke-derived melts rather than
553 metapelite-derived melts (Fig. 11a; Whalen et al., 1987; Sylvester, 1998). Their
554 geochemical features are also more similar to experimental melts of metagraywacke
555 than that of metapelite (Figs. 13a–13e). In addition to the Chinese Wakhan Corridor,
556 coeval (ca. 119-108 Ma) S-type granites of the Taxkorgan pluton were intruded north
557 of the studied area (Fig. 1b; Jiang et al., 2014; Li et al., 2019; Ma et al., 2023).
558 However, these are characterized by variable CaO/Na₂O (0.21–0.53), Rb/Sr (0.93–
559 3.55) and Rb/Ba (0.32–2.47) ratios (Supplementary Table 3), indicating different
560 source characteristics (i.e., metagreywacke or metapelite-derived melts; Fig. 11a and
561 Figs. 13a–13e). The zircon ε_{Hf}(*t*) values of the Taxkorgan samples show a variation of
562 ~4.5 epsilon units (-10 in Jiang et al., 2014 and -14.5 in Li et al., 2019), also
563 indicating that their supracrustal sources were heterogeneous. In summary, we
564 interpret the Kalaqigu S-type monzogranites to be the products of partial melting of
565 metagreywacke, while the Taxkorgan S-type rocks were formed by either

566 metagreywacke or metapelite-derived melts, due to the differences between crustal
567 composition of the Central Pamir and South Pamir (Imrecke et al., 2019).

568 In addition, water distribution during the melting process also plays an important
569 role in the generation of S-type granites (e.g., Sylvester, 1998; Clemens and Stevens,
570 2012). Partial melting of metagreywackes can generally be divided into H₂O-fluxed
571 melting and dehydration reactions of hydrous minerals such as muscovite or biotite
572 (Conrad et al., 1988; Montel and Vielzeuf, 1997; Patiño Douce and Harris, 1998).
573 Several observations suggest that the Kalaqigu S-type monzogranites were generated
574 by biotite dehydration melting under water-absent conditions. Firstly, plagioclase
575 dissolves easily in the melt under water-present conditions, which leads to higher Ca
576 and Na contents as well as positive Eu anomalies (Patiño Douce and Harris, 1998;
577 García-Arias et al., 2015). However, these monzogranites have relatively low CaO
578 contents (1.6–2.6 wt.%) and high K₂O/Na₂O ratios (0.6–1.3), distinct from
579 H₂O-fluxed melting of metagreywackes (CaO = 1.4–3.9 wt.%, K₂O/Na₂O = 0.3–1.0,
580 Conrad et al., 1988). Furthermore, they exhibit negative Eu anomalies (Eu/Eu* =
581 0.67–0.83; Fig. 6c), indicating the existence of residual plagioclase. Secondly,
582 negative correlation between Rb/Sr ratios and Sr but positive correlation with Ba
583 contents are consistent with fluid-absent biotite dehydration melting (Figs. 13e–13f;
584 Inger and Harris, 1993). Similarly, low CaO contents (0.8–1.7 wt.%), high K₂O/Na₂O
585 ratios (1.0–1.5), as well as negative Eu anomalies of Taxkorgan S-type granites
586 (Eu/Eu* = 0.4–0.6; Supplementary Table 3) also suggest they were generated by
587 dehydration melting.

588 Accordingly, we used a metagreywacke sample from Vielzeuf and Montel (1994)
589 for phase equilibrium modeling to determine the pressure-temperature (P-T)
590 conditions during anatexis (Fig. 14a). Phase equilibrium modeling used the GeoPS
591 software tool (<http://www.geology.ren/>; Xiang and Connolly, 2022). The contents of
592 biotite and H₂O_{Bt} (H₂O in biotite) decrease with the rising temperature, indicating that
593 anatexis was most likely a result of biotite-dehydration melting reaction (Fig. 14b).
594 This is consistent with peritectic cordierite and K-feldspar with biotite inclusions
595 (Figs. 2c–2d), as cordierite and K-feldspar modes increase at the expense of biotite

596 during prograde melting (Fig. 14c). Therefore, the Kalaqigu S-type monzogranites
597 were probably generated by biotite-dehydration melting from metagreywacke.

598 Estimating P–T conditions is essential to constrain not only the melting
599 mechanism, but also the nature of magma source and process of magma
600 crystallization. According to our phase equilibrium modeling (Figs. 14b–14c), the
601 reaction biotite + plagioclase + quartz = orthopyroxene + garnet (under high
602 pressure)/cordierite (under low pressure) + K-feldspar + melt occurs during the
603 fluid-absent partial melting of metagreywacke, consistent with previous studies
604 (Thompson, 1982; Clemens, 1984; Vielzeuf and Montel, 1994). The formation of
605 garnet and its existence as a residual phase in the source area are also accordance with
606 the geochemical features of these monzogranites (particularly REE fractionation; Fig.
607 6c).

608 As the biotite occurs as late interstitial crystallizing phase, the pressures (5.5–6.4
609 kbar, with an average of 6.0 kbar) and temperatures (746–772 °C, with an average of
610 757 °C) of the biotites represent minimum estimates (Fig. 14a), corresponding to the
611 magma pressure and temperature during late-stage crystallization. The disappearance
612 of biotite represents the completion of biotite-dehydration melting (Vielzeuf and
613 Montel, 1994), which suggests that crystallization began at ~850 °C and >6 kbar. The
614 calculated and estimated pressure-temperature results are also in agreement with the
615 inferred range of pressure conditions based on experimental phase relations (Fig. 14a).
616 Under these P-T conditions, the inferred sequence of crystallization is likely to be
617 K-feldspar → plagioclase → cordierite → biotite + quartz. This interpretation is
618 supported by the petrological observation that K-feldspar occurs as phenocrysts and
619 biotite occurs as inclusions in the K-feldspar and cordierite (Fig. 2c). In brief, we
620 consider that the magma of Kalaqigu S-type monzogranites formed at pressures > 6
621 kbar, corresponding to lower crustal conditions.

622

623 **Tectonic Implications**

624 ***Onset of Andean-type Continental Arc***

625 It has been proposed that there are two arcs in South Pamir, namely the

626 Kohistan-Ladakh intra-oceanic arc and the continental arc subducted northward along
627 the SSZ in the South Pamir-Karakorum (Zanchi et al., 2000; Zanchi and Gaetani,
628 2011; Chapman et al., 2018a). The Early Cretaceous high-Mg diorites, and the
629 mantle-derived volcanic rocks in the Wakhan Corridor, show enrichment in LILEs
630 and depletion in Nb, Ta and HFSEs, suggesting an arc affinity (Figs. 6 and Fig. 15a; Ji
631 et al., 2016; Liu et al., 2020; Zhang et al., 2022). These units have geochemical
632 signature of continental arc magmatic rocks, as determined on Nb/Yb versus Th/Yb
633 and Ti/V versus Th/Nb diagrams (Figs. 15b–15c; Shervais, 2022). Previous studies
634 suggested that the Neo-Tethys oceanic lithosphere subducted northward along the
635 SSZ beneath the Karakoram terrane (Fraser et al., 2001; Bouilhol et al., 2013; Kumar
636 et al., 2017; Chapman et al., 2018a). This resulted in a combination of continental arc
637 magmatism in the South Pamir and the development of the thrust belt in the North
638 Pamir and Central Pamir (Robinson et al., 2004; Imrecke et al., 2019; Li et al., 2022;
639 Villarreal et al., 2023), forming an Andean style orogenic belt, similar to the western
640 North American Cordillera (Dickinson et al., 1978; Gutscher et al., 2000; Axen et al.,
641 2018). Thus, it is reasonable to deduce that the Early Cretaceous intrusive rocks have
642 similar Andean-type continental arc-related fingerprints and have developed along the
643 Neo-Tethys oceanic subduction zone.

644 Our new zircon U–Pb data combined with previously published data show that
645 there were two episodes of Early Cretaceous continental arc magmatism in the
646 Wakhan Corridor (Fig. 15d). In the first stage (ca. 115–102 Ma with ca. 105 Ma as a
647 high-flux event), extensive crustal remelting and crust-mantle interaction formed
648 continental arc-related magmatic rocks (Fig. 15d). Following this episode (ca. 107–98
649 Ma), a relatively small volume of mantle-derived rocks formed in the South
650 Pamir-Karakorum (Figs. 15d and 16a). In the Early Cretaceous, the Neo-Tethys
651 oceanic lithosphere may have undergone low-angle to flat subduction beneath the
652 South Pamir-Karakoram (e.g., Fraser et al., 2001; Bouilhol et al., 2013; Kumar et al.,
653 2017; Chapman et al., 2018a). This is further supported by the magmatic migration
654 from the Wakhan Corridor to the northwest Teshiktash-Beik (Fig. 15d; Aminov et al.,
655 2017; Ma et al., 2023). During this period, the subduction of the oceanic slab

656 produced fluids at sub-arc depths, which induced melting of mantle wedge,
657 interaction with overlying Pamir crust as well as the remelting of the ancient lower
658 crust (Fig. 16a). In this scenario, a series of granitoids at ca. 105 Ma as a high-flux
659 event formed arc-related magmatic rocks (Figs. 15d and 16a). Following this period,
660 mantle-derived basalt-andesite as well as the studied high-Mg diorites in the Wakhan
661 Corridor resulted from interaction between slab melts (i.e., sedimentary melts) and the
662 mantle (Figs. 15d and 16b). We propose that subducted sediments played an important
663 role in the formation of the mantle-derived magmatic rocks in the Wakhan Corridor
664 during the Early Cretaceous (Figs. 7, 8 and 11).

665

666 **Implications for the Correlation of Early Cretaceous S-type Granites**

667 As mentioned above, contemporaneous S-type granites are also exposed in the
668 Taxkorgan pluton, albeit slightly earlier (ca. 118-108 Ma; Jiang et al., 2014; Li et al.,
669 2019; Ma et al., 2023) than that in the Wakhan Corridor (Figs. 15c and 16a). They
670 also exhibit typical subduction-related continental arc geochemistry (Figs. 15b-15c;
671 Jiang et al., 2014; Li et al., 2019) and have the geochemical characteristics of
672 syn-collisional granite (Fig. 15a; Pearce et al., 1984). Based on the temporal-spatial
673 evolution of Cretaceous arc magmatism in the Pamir, it is likely that the Taxkorgan
674 S-type granites were generated in a collisional setting caused by the northward
675 subduction of the Neo-Tethys oceanic lithosphere (Fig. 16a). The north-south
676 difference in crystallization age indicates that the collisional compression caused by
677 subduction occurred slightly earlier than the flare-up of the South Pamir-Karakoram
678 magmas.

679 Furthermore, the crustal source of the Taxkorgan S-type granites is
680 heterogeneous (i.e., metagreywacke and metapelite; Figs. 11a, 13a-d and 16a). Based
681 on the structural relationships and distribution of rock units Imrecke et al. (2019)
682 suggested that southeast Pamir consists of two distinct structural/lithologic domains:
683 the northern schist and gneiss region and the southern metamorphic sedimentary rocks.
684 Previous studies have shown that low-angle or flat-slab subduction is closely related
685 to upper plate shortening and back-arc deformation (Egawa, 2013; Gianni et al., 2018;

686 Schellart, 2020). The Cretaceous Neo-Tethys oceanic slab migrated progressively
687 landward from South Pamir-Karakorum to the north beneath the Pamir crust. This
688 resulted in a maximum arc-trench distance of ca. 400 km in the Early Cretaceous (Ma
689 et al., 2023), and caused obvious retroarc shortening (Robinson et al., 2007, 2012), as
690 well as imbrication in the Taxkorgan due to the development of a regionally extensive
691 thrust nappe in the North Pamir (Imrecke et al., 2019). Thus, we interpret that such
692 imbrication along SW-vergent thrust is the main reason for the heterogeneous crustal
693 source of the S-type granites.

694 Landward arc migration caused by flat-subduction is often accompanied by
695 shortening and thickening events (e.g., Gianni et al., 2018). Research on
696 low-temperature thermochronology, sedimentary petrology and metamorphic
697 petrology show that Early Cretaceous (ca. 140-110 Ma) crustal shortening and
698 thickening appears to be focused along the North Pamir (Robinson et al., 2012;
699 Robinson, 2015; Villarreal et al., 2023). This is manifested in the amphibolite-facies
700 metamorphism at ca. 130-110 Ma (Robinson et al., 2004), broadly coeval exhumation
701 in the hanging wall of thrusts (Robinson et al., 2007; Imrecke et al., 2019; Villarreal et
702 al., 2023), as well as the widespread occurrence of thrust fault movement in the North
703 Pamir (Chapman et al., 2018b; Li et al., 2022; Villarreal et al., 2023). These
704 observations, combined with Taxkorgan syn-collisional granites, suggest the
705 subduction of the Neo-Tethys oceanic slab resulted in compressive deformation
706 primarily occurring far north of the subduction zone during the Early Cretaceous,
707 which also caused a general lack of magmatism in this area (Ma et al., 2023). This
708 compression-dominated environment also resulted in significant crustal thickening in
709 the Pamir during the Early Cretaceous (Li et al., 2022; Ma et al., 2023; Villarreal et al.,
710 2023). Following this period, prograde metamorphism indicates southward migration
711 of crustal shortening and thickening into the Central Pamir and South Pamir have
712 occurred during ca. 110-75 Ma (Chapman et al., 2018b), broadly coeval with the main
713 phase of magmatism in the South Pamir-Karakorum (Fig. 15d).

714 To summarize, we interpret Early Cretaceous crustal shortening to have resulted
715 from the low-angle and flat-subduction of Neo-Tethys oceanic lithosphere (Fig. 16).

716 Crustal thickening could have been caused by regional compression (Fig. 16a) as well
717 as addition of mantle-derived magma (Fig. 16b) (Ma et al., 2023). In the Late
718 Cretaceous (<90 Ma), the Neo-Tethys Oceanic slab may have undergone slab
719 roll-back, which resulted in extension-related magmatism in the Pamir (Fig. 16b;
720 Chapman et al., 2018a).

721

722 **CONCLUSIONS**

723 (1) Zircon LA-ICP-MS U-Pb dating reveals that Kalaqigu high-Mg diorites and
724 S-type monzogranites in the Wakhan Corridor were emplaced in the Early Cretaceous
725 (ca. 108.4–105.9 Ma).

726 (2) The high-Mg diorites formed from an enriched phlogopite-bearing
727 spinel-lherzolite hydrous mantle source modified by subducted sediment-derived
728 melts, and underwent low-degree fractionation of olivine and orthopyroxene.

729 (3) The S-type monzogranites were generated by biotite-dehydration melting from
730 metagreywacke under lower crustal conditions.

731 (4) In the subduction zone beneath the South Pamir, the subducted slab first
732 underwent dehydration and the resultant fluids generated a hydrous mantle source and
733 induced crust-mantle interaction as well as remelting of the lower crust to generate
734 granitoids. The partial melts of subducted sediments then continued to metasomatize
735 the mantle wedge, which generated extensive mantle-derived magmatic rocks.

736 (5) The northward low-angle flat-subduction of Neo-Tethys oceanic lithosphere had
737 subducted and migrated inland underneath the Pamir continent, leading to significant
738 Early Cretaceous continental-arc magmatism, inducing crust-mantle interaction
739 beneath the South Pamir and resulting in crustal shortening and thickening in the
740 Pamir.

741

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755

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1268

1269 **Figure captions**

1270 **Figure 1.** (a) Regional tectonic map, showing the geographic location of the Pamir
1271 (after Burtman and Molnar, 1993; Yin and Harrison, 2000); (b) Simplified tectonic
1272 map of the Pamir, showing distribution of tectonic terranes and plutons (Modified
1273 from Robinson, 2015; Aminov et al., 2017; Zanchetta et al., 2018); (c) Simplified
1274 geological map of Kalaqigu pluton at the Chinese Wakhan Corridor, Southern Pamir.
1275 IYSZ–Indus-Tsangpo Suture Zone; KKF–Karakorum Fault; MPT–Main Pamir Thrust;
1276 RPSZ–Rushan-Pshart Suture Zone; SSZ–Shyok Suture Zone; TSZ–Tanyamas Suture
1277 Zone; WTBZ–Wakhan-Tirich Boundary Zone.

1278

1279 **Figure 2.** Representative thin section photomicrographs of Kalaqigu diorite (a-b) and
1280 monzogranite (c-d). Figures (a) and (c) are under cross-polarized light and Figures (b)
1281 and (d) are under the plane-polarized light. Amp—amphibole; Bi—biotite;
1282 Chl—chlorite; Crd—cordierite; Kfs—K-feldspar; Pl—plagioclase; Q—quartz;
1283 Sil—sillimanite.

1284

1285

1286 **Figure 3.** Concordia diagrams with representative zircon CL images for LA-ICP-MS
1287 zircon analyses of Kalaqigu diorite (a) and monzogranite (b). (c) Histogram of U-Pb
1288 ages for inherited zircon cores from the studied monzogranite; (d) Representative
1289 cathodoluminescence (CL) images of zircon cores from the studied monzogranite.
1290 The data sources for detrital zircons of southeastern Pamir are from Imrecke et al.
1291 (2019).

1292

1293 **Figure 4.** (a-b) Classification diagrams for amphiboles (Leake et al., 1997;
1294 Hawthorne and Oberti, 2007). (c) Mg vs. $(Al^{VI}+Fe^{3+}+Ti)$ vs. $(Fe^{2+}+Mn)$ ternary
1295 diagram for biotite (Foster, 1960). (d) Or–Ab–An classification diagram for
1296 plagioclase (modified after Deer et al. 1992). An = anorthite, And = andesine, Ab =
1297 albite, By = bytownite, La = labradorite, Ol = oligoclase, Or = orthoclase.

1298

1299 **Figure 5.** Geochemical classification and major element geochemical features for the
1300 Kalaqigu diorites and monzogranites. (a) TAS classification diagram (Middlemost,
1301 1994); (b) K_2O versus Na_2O diagram (Rollinson, 1993); (c) K_2O versus Na_2O
1302 diagram (Le Maitre, 1989); (d) A/NK versus A/CNK diagram (Maniar and Piccoli,
1303 1989).

1304

1305 **Figure 6.** Chondrite-normalized REE patterns (a and c) and primitive
1306 mantle-normalized multi-element patterns (b and d) for the Kalaqigu diorites and
1307 monzogranites. Chondrite and primitive mantle values are from Sun and McDonough
1308 (1989).

1309

1310 **Figure 7.** (a) $\epsilon_{Nd}(t)$ versus $(^{87}Sr/^{86}Sr)_i$; $(^{207}Pb/^{204}Pb)_i$ (b) and $(^{208}Pb/^{204}Pb)_i$ (c) versus
1311 $(^{206}Pb/^{204}Pb)_i$ diagrams for the Kalaqigu diorites and monzogranites. The depleted
1312 MORB-source mantle (DMM) shown is from Workman and Hart (2005). The
1313 Archean basement is from Ji et al., (2011). Cretaceous basalt-andesite, diorites and

1314 granitoids from the Chinese Wakhan Corridor are shown for comparison (Jiang et al.,
1315 2014; Li et al., 2016; Liu et al., 2020; Zhang et al., 2022). The Northern Hemisphere
1316 Reference Line (NHRL) is from Hart, (1984). EMI, EMII and GLOSS are from Hart
1317 (1988), Zindler and Hart (1986) and Plank and Langmuir (1998), respectively.

1318

1319 **Figure 8.** Oxygen isotope data for zircon from the Kalaqigu diorite and
1320 monzogranites. The mantle values of zircon oxygen isotope (5.3 ± 0.6 ‰, 2SD) are
1321 from Valley (2003).

1322

1323 **Figure 9.** (a) $\text{MgO}/(\text{MgO} + \text{FeO}^{\text{T}})$ versus TiO_2 , and (b) Y versus Sr/Y discrimination
1324 diagrams (after Kamei et al., 2004). (c) 10,000 Ga/Al versus Nb, and (d) $(\text{Zr} + \text{Nb} +$
1325 $\text{Ce} + \text{Y})$ versus $\text{FeO}^{\text{T}}/\text{MgO}$ (Whalen et al., 1987). FG—fractionated M-, I-, and S-type
1326 granite; OGT—unfractionated M-, I-, and S-type granite.

1327

1328 **Figure 10.** (a) $\text{Mg}^{\#}$ versus SiO_2 (after Rapp et al., 1999; Martin et al., 2005); (b-f)
1329 Fenner diagrams showing selected major elements variations for the Kalaqigu diorites.
1330 The green and blue lines represent crystallization trends defined by major element
1331 modelling using Rhyolite-MELTS, divided into olivine + orthopyroxene and
1332 clinopyroxene + spinel steps. Diorites samples WK1616-2 with lowest contents of
1333 SiO_2 and WK1616-4 with highest MgO have been taken as primary melts
1334 (Supplementary Table 3). The model was run at 5.16 kbar with $\text{H}_2\text{O}_{\text{melt}} = 5.84$ wt.%,
1335 calculated from amphibole compositions (Supplementary Table 2).

1336

1337 **Figure 11.** (a) Rb/Ba versus Rb/Sr, showing geochemical compositions of magma
1338 source for the Kalaqigu diorites and monzogranites. The mixing curve between the
1339 basalt- and pelite-derived melts is from Sylvester (1998); (b) Th/La versus Th/Sm.
1340 N-MORB (normal mid-oceanic-ridge basalt) and OIB (oceanic-island basalt) values
1341 are from Sun and McDonough (1989), and the values for average GLOSS are after
1342 Plank and Langmuir (1998). The curve shows different mixing ratios between partial
1343 melt (4%) of the GLOSS average and N-MORB. The D_{Th} and D_{La} are 0.16 and 1.2,

1344 respectively (Plank, 2005). We use the same D as La for Sm ($D_{Sm} = 1.2$). (c) Ba/Th
1345 versus $(La/Sm)_N$ and (d) Th/Yb versus Ba/La discrimination diagrams for
1346 metasomatic agents added to the mantle wedge. Taxkorgan S-type granites in (a) are
1347 from Jiang et al. (2014) and Li et al. (2019). Cretaceous mantle-derived magmatic
1348 rocks (i.e., basalt-andesite and diorite) with $MgO > 3\%$ from Chinese Wakhan
1349 Corridor are shown for comparison (Li et al., 2016; Liu et al., 2020; Zhang et al.,
1350 2022). Data of sanukitic HMAs in Figures (c) and (d) are from Hanyu et al. (2006)
1351 and Tatsumi (2006).

1352

1353 **Figure 12.** Dy/Yb versus La/Yb (a) and Rb/Sr versus Ba/Rb (b) diagrams. All the
1354 mantle models in (a) are from Xu et al. (2001). Amphibole and phlogopite arrows in
1355 (b) refer to these as residual phases in the source region (Furman and Graham, 1999).
1356 Cretaceous mantle-derived magmatic rocks are those shown in Fig. 11.

1357

1358 **Figure 13.** Plots of $Al_2O_3 + MgO + FeO^T + TiO_2$ versus $Al_2O_3/(MgO + FeO^T + TiO_2)$
1359 (a); $Na_2O + K_2O + FeO^T + MgO + TiO_2$ versus $(Na_2O + K_2O)/(FeO^T + MgO + TiO_2)$
1360 (b); Na_2O/K_2O versus FeO^T (c) and $CaO + FeO^T + MgO + TiO_2$ versus $CaO/(FeO^T +$
1361 $MgO + TiO_2)$ (d) (a–d are after Patiño Douce, 1999). (e–f) Plots of Rb/Sr ratios versus
1362 Sr (ppm) and Ba (ppm), respectively (after Inger and Harris, 1993). Taxkorgan S-type
1363 granites in (a–d) are shown in Fig. 13.

1364

1365 **Figure 14.** (a) Pressure-temperature (P-T) pseudosection calculated for
1366 metagreywacke from Vielzeuf and Montel. (1994). Yellow circle represents P-T
1367 conditions calculated from biotite compositions (Li and Zhang, 2022; Supplementary
1368 Table 2). Bi—biotite; Crd—cordierite; Gt—garnet; Ilm—ilmenite; Kfs—K-feldspar;
1369 Ms—muscovite; Opx—orthopyroxene; Pl—plagioclase; q—quartz; ru—rutile;
1370 sill—sillimanite. (b–c) Isomodes of biotite, plagioclase, H_2O_{Bt} , garnet, cordierite and
1371 K-feldspar in different intervals, indicating growth of garnet, cordierite and
1372 K-feldspar at the expense of biotite and plagioclase during melting as P-T increases.

1373

1374 **Figure 15.** (a) Rb versus (Y + Nb) diagrams (after Pearce et al., 1984); (b) Th/Yb
1375 versus Nb/Yb and (c) Th/Nb versus Ti/V diagrams (after Shervais, 2022); (d)
1376 Histogram of zircon U-Pb ages. VAG—volcanic arc granites;
1377 syn-COLG—syn-collisional granites; WPG—within-plate granites;
1378 ORG—ocean-ridge granites. Cretaceous mantle-derived magmatic rocks and
1379 granitoids are shown in Figs. 11. Cretaceous granitoids from the Chinese Wakhan
1380 Corridor in (b-c) are also shown for comparison (Jiang et al., 2014; Li et al., 2016; Liu
1381 et al., 2020). Taxkorgan S-type granites are from Jiang et al. (2014), Li et al. (2019)
1382 and Ma et al. (2023). Teshiktash-Beik volcanic rocks, northwest of Chinese Wakhan
1383 Corridor, are from Aminov et al. (2017).

1384

1385 **Figure 16.** Schematic diagram showing the Cretaceous multi-stage and multi-source
1386 processes for the architecture of the continental arc of the South Pamir. (a) Northward
1387 low-angle and flat-slab subduction of the Neo-Tethys oceanic lithosphere resulted in
1388 the generation of Cretaceous granitoids. (b) Continuous flat-slab subduction of the
1389 Neo-Tethys oceanic lithosphere prompted sedimentary melts to metasomatize the
1390 mantle, produced mantle-derived magmatic rocks. The figures also show Early
1391 Cretaceous crustal shortening and thickening events in the Central Pamir and South
1392 Pamir. The Neo-Tethys oceanic slab may undergo slab roll-back in the Late
1393 Cretaceous (<90 Ma) (Chapman et al., 2018a).

1394

Figure 1

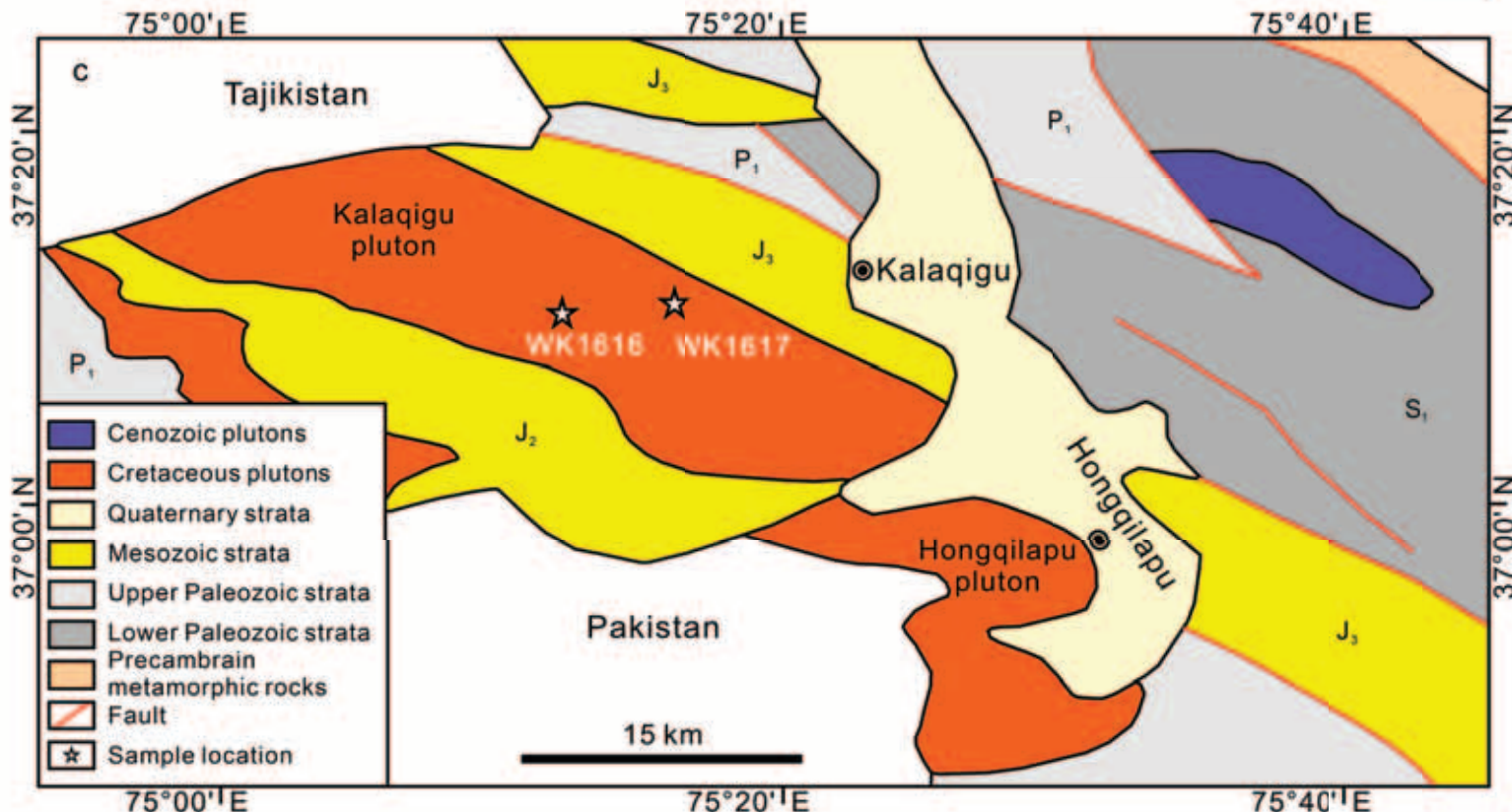
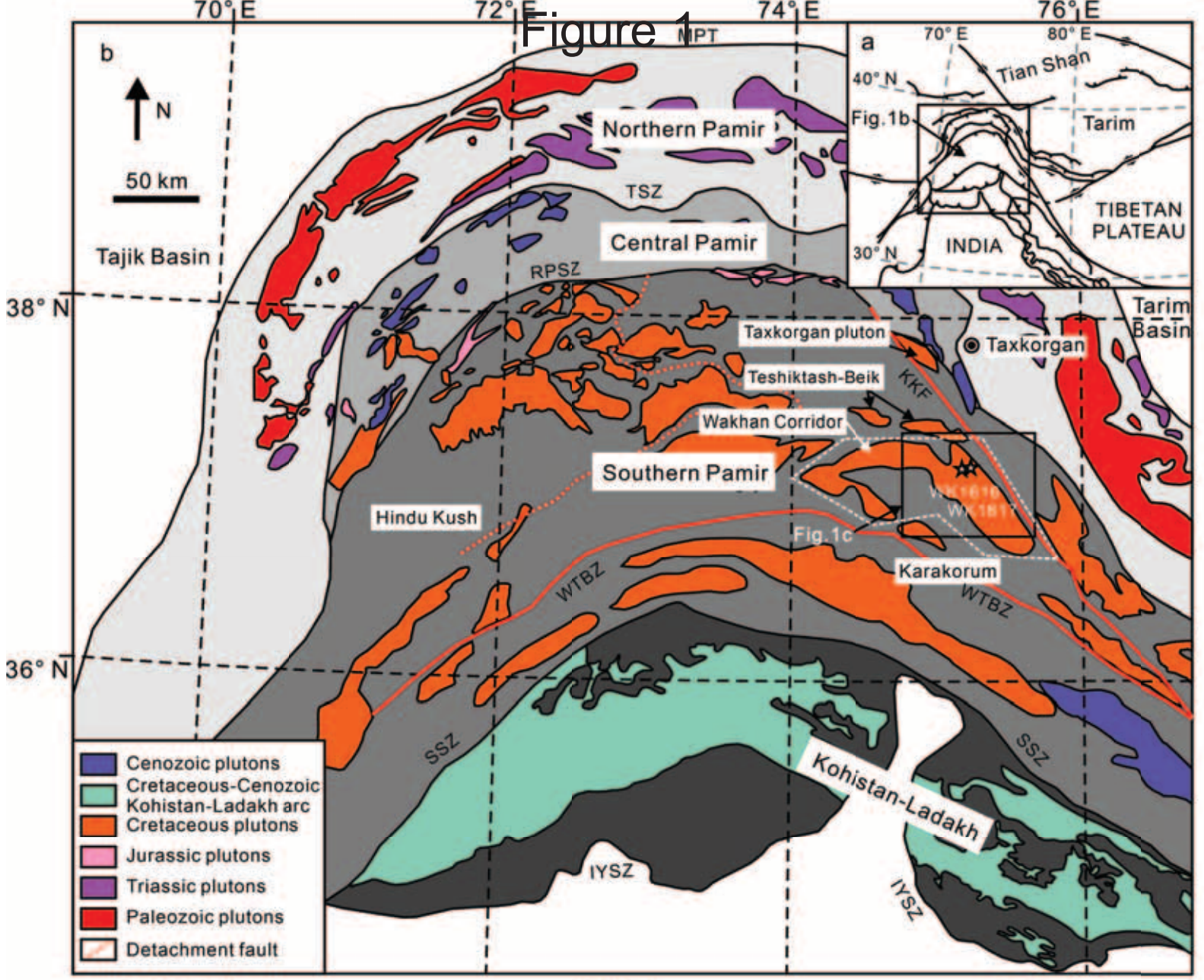


Figure 2

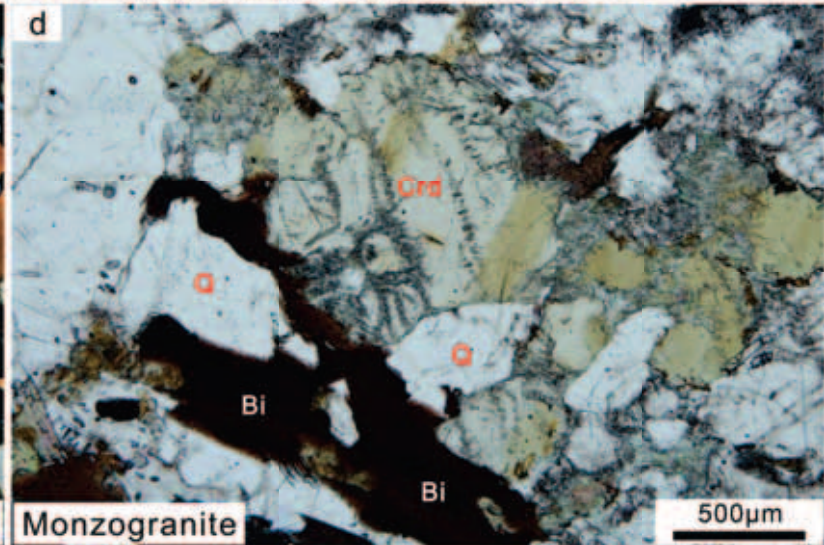
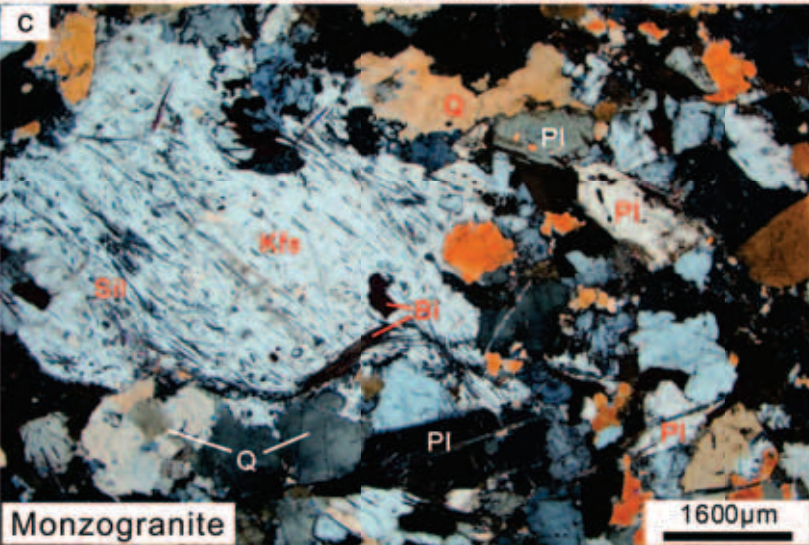
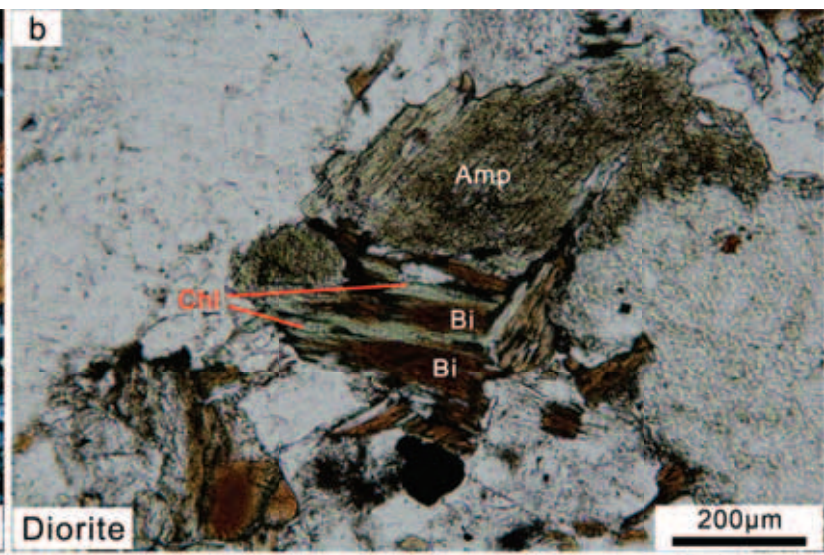
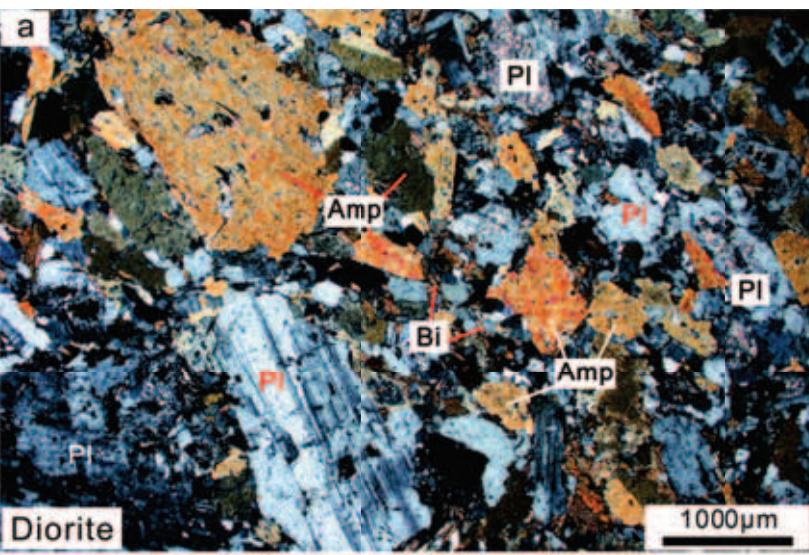


Figure 3

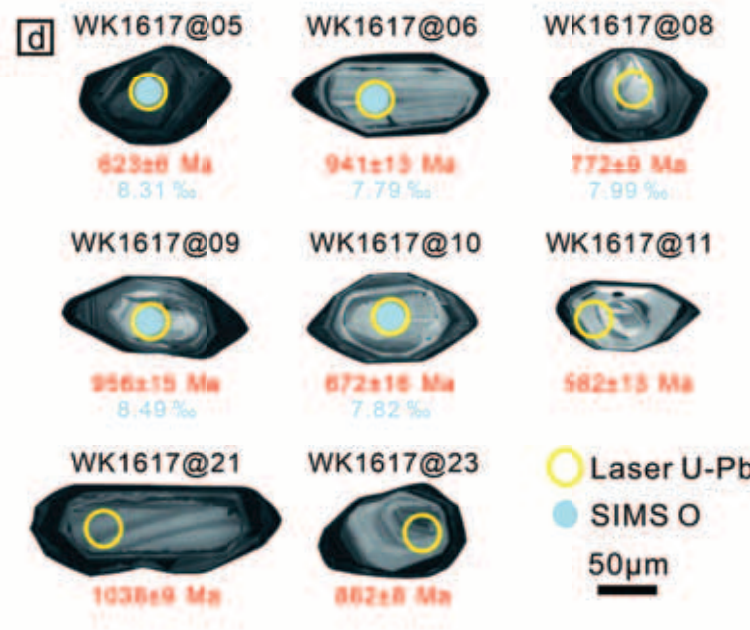
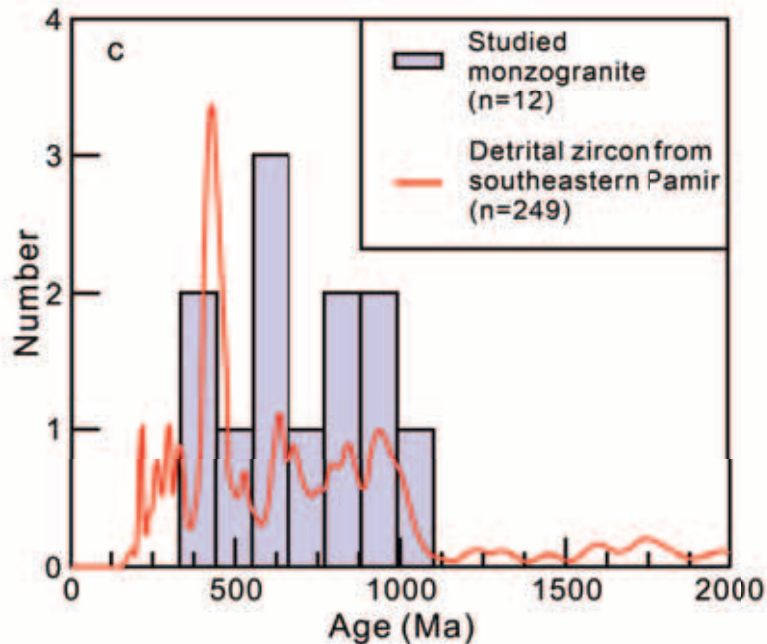
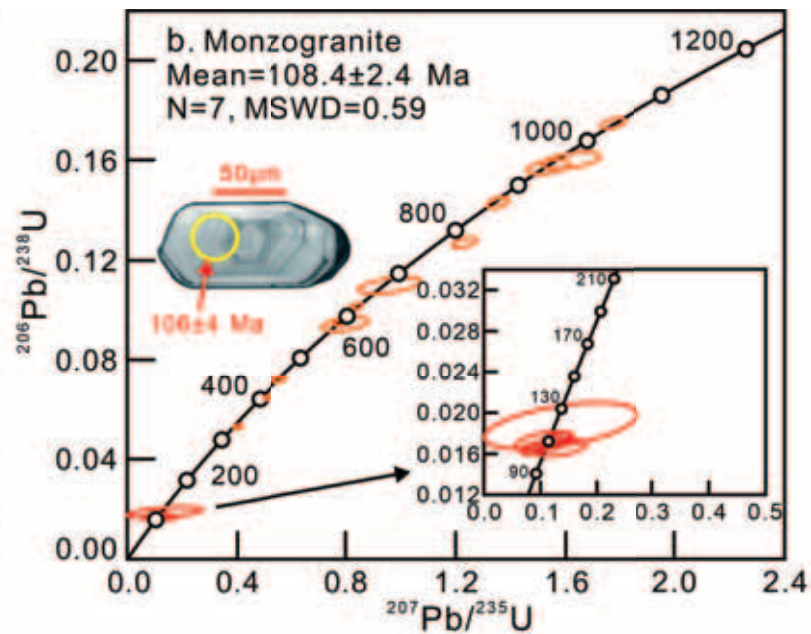
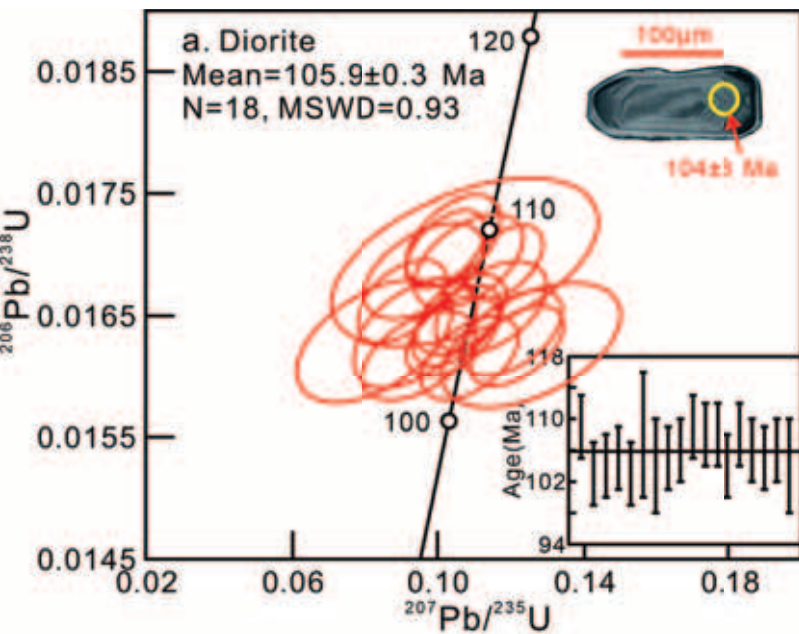


Figure 4

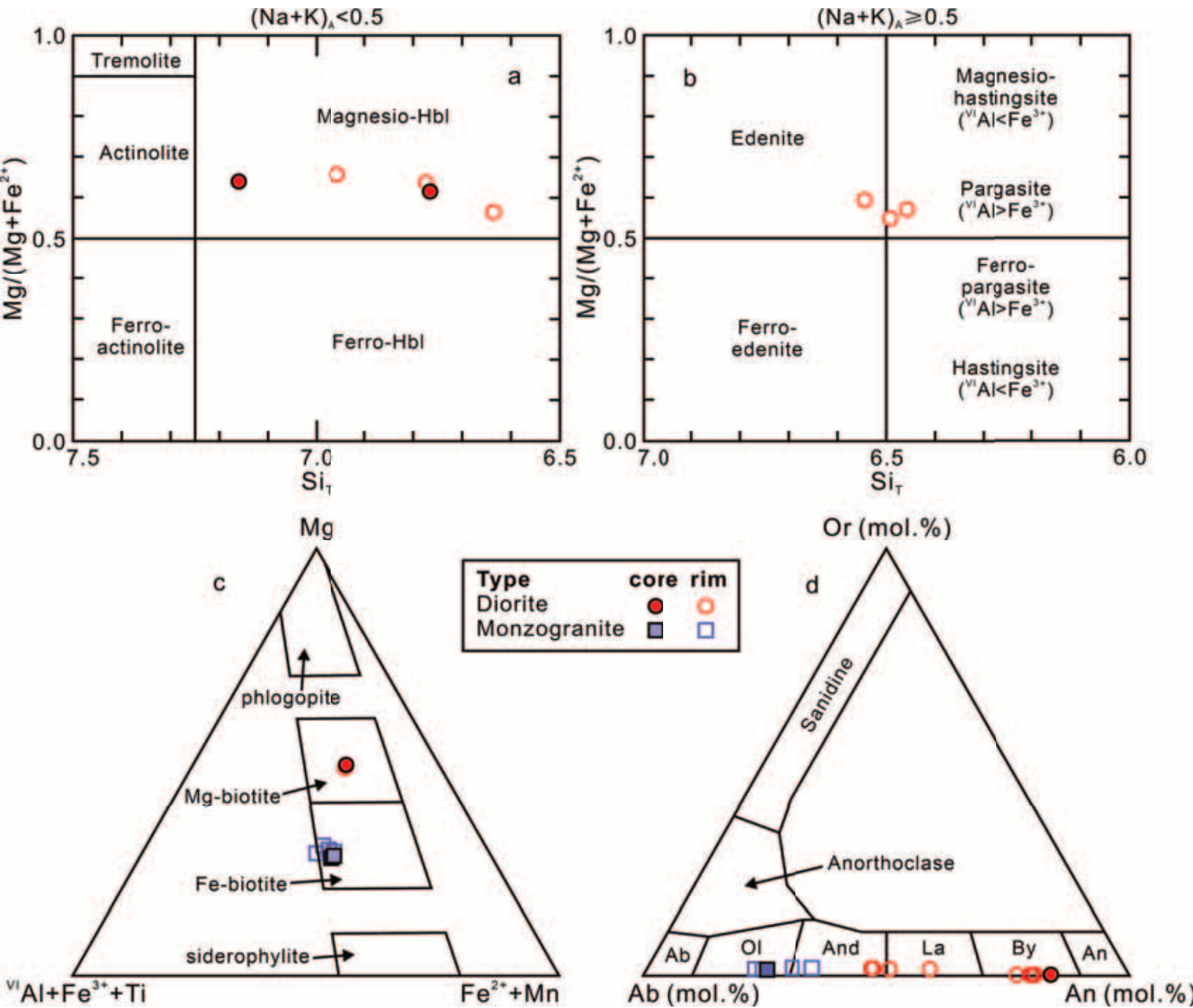


Figure 5

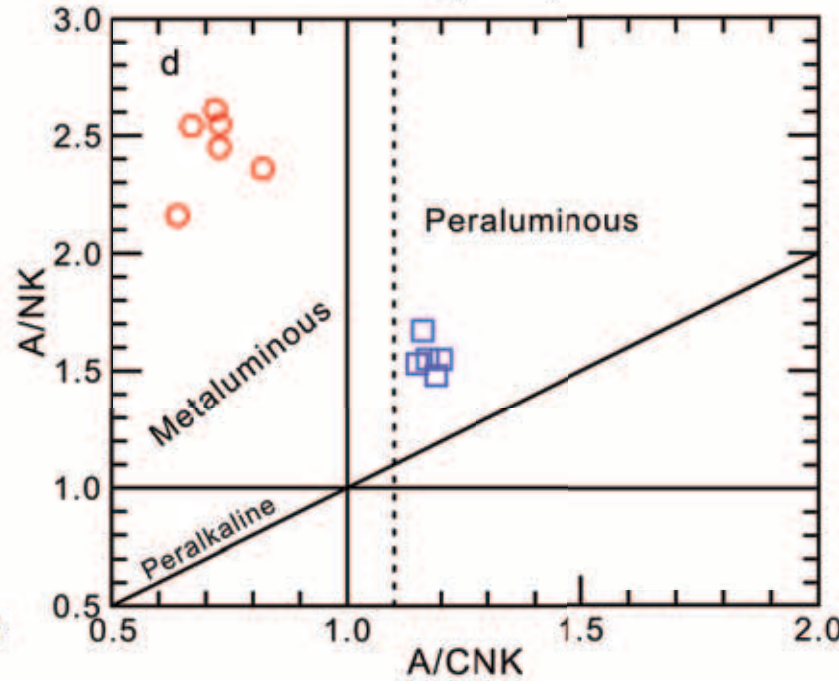
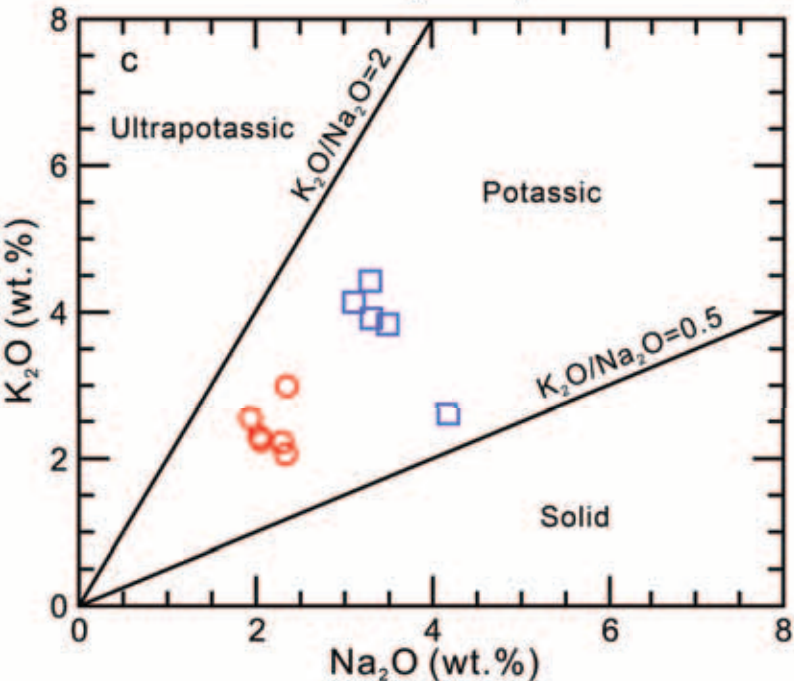
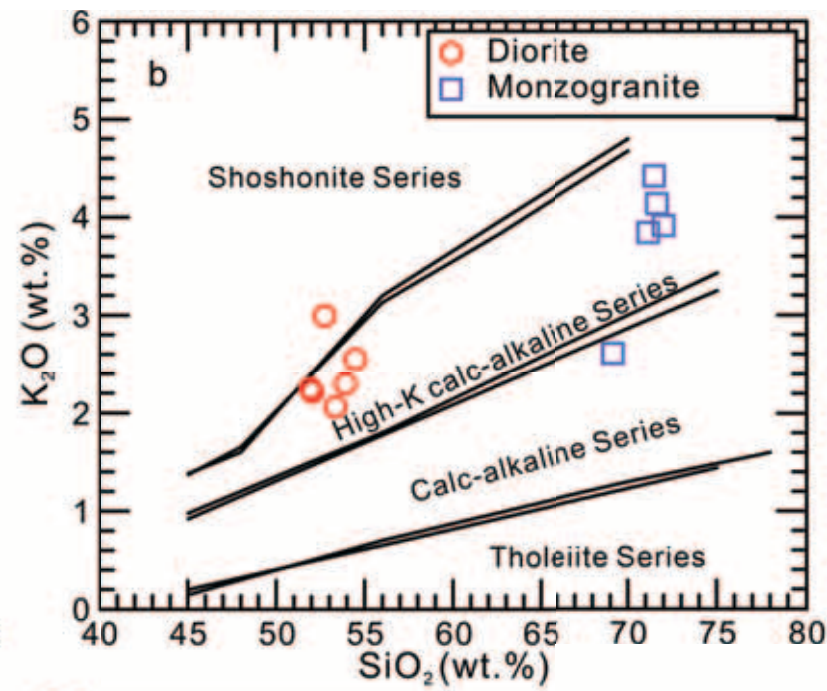
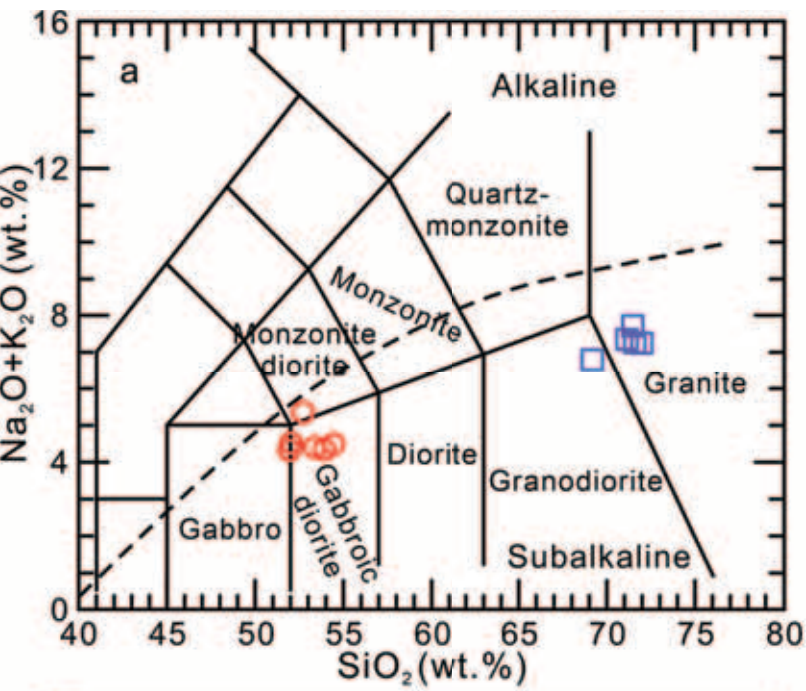


Figure 6

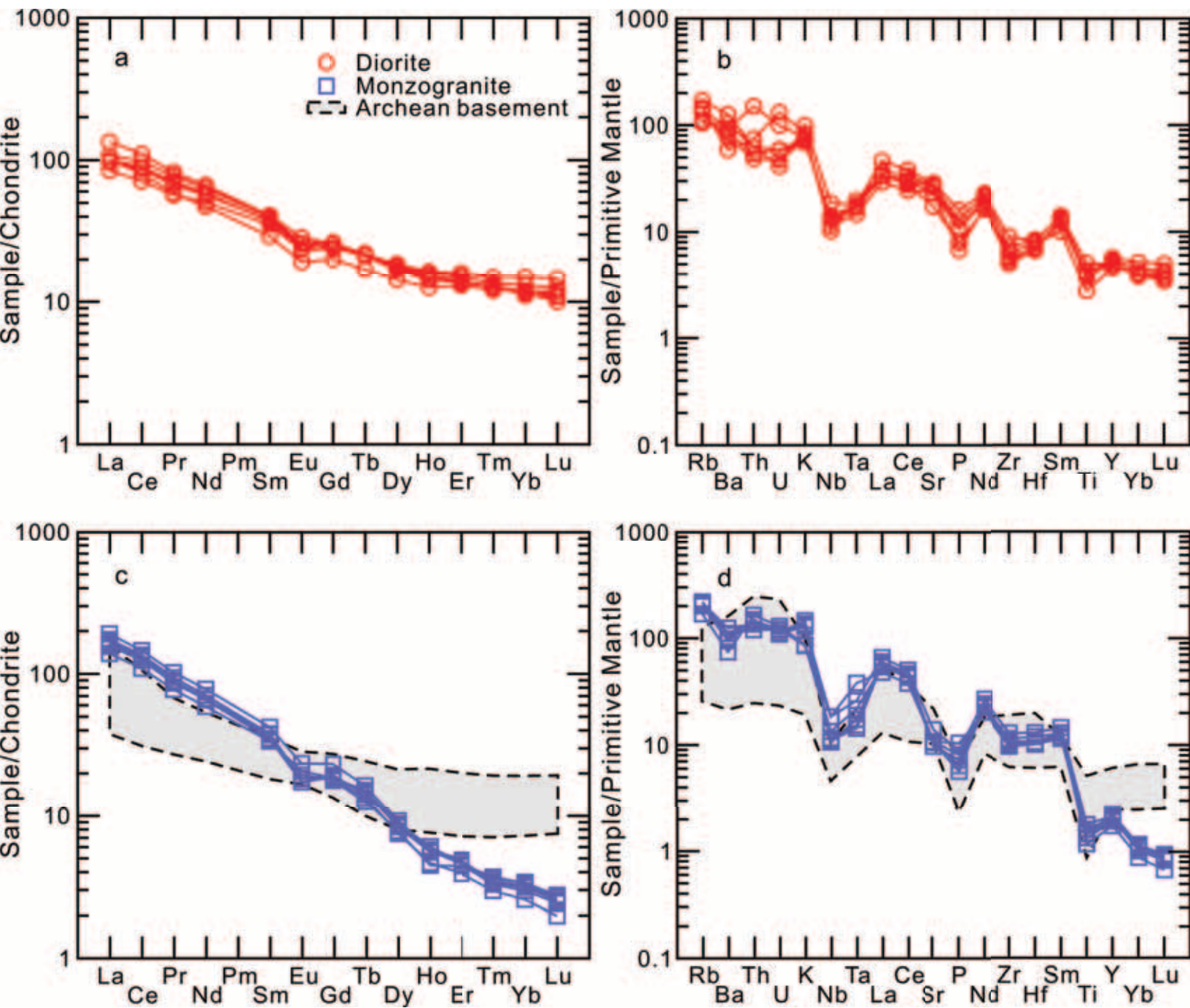


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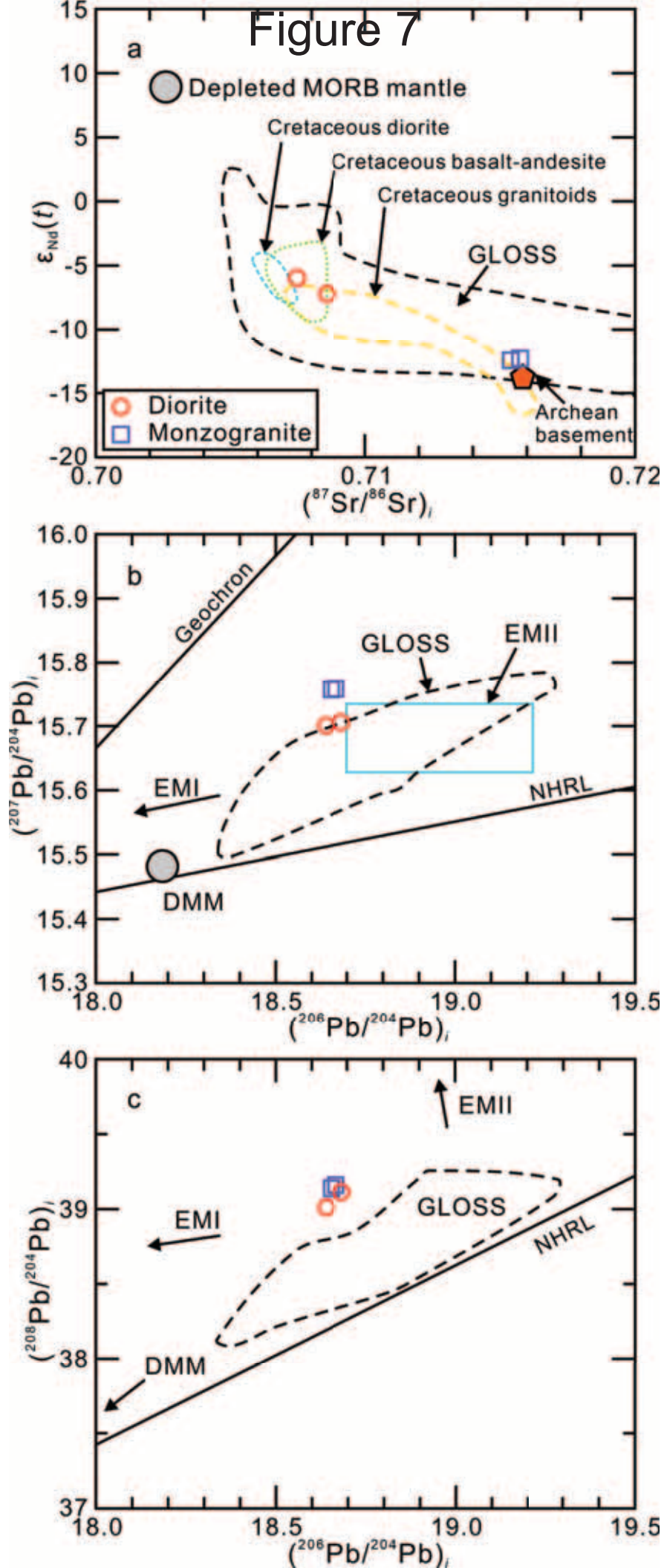


Figure 8

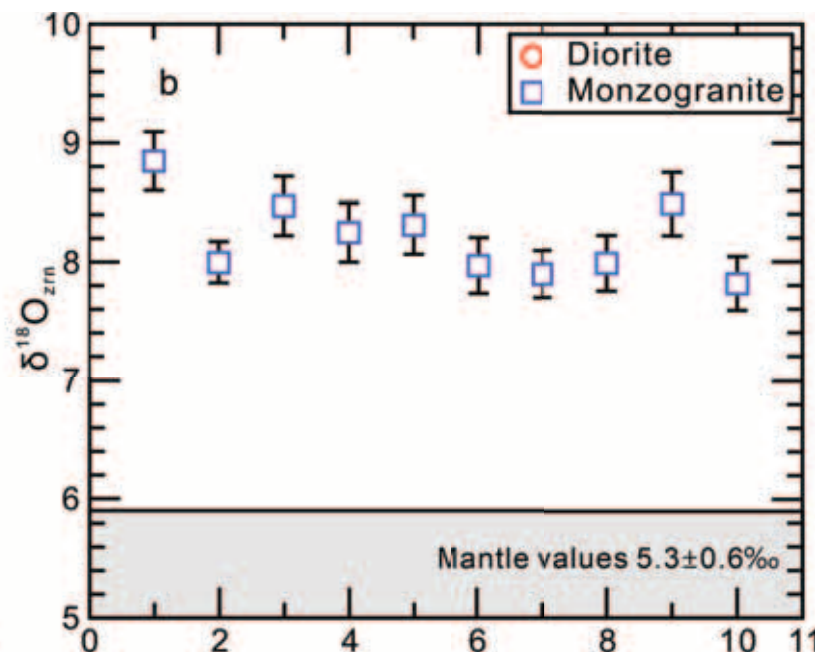
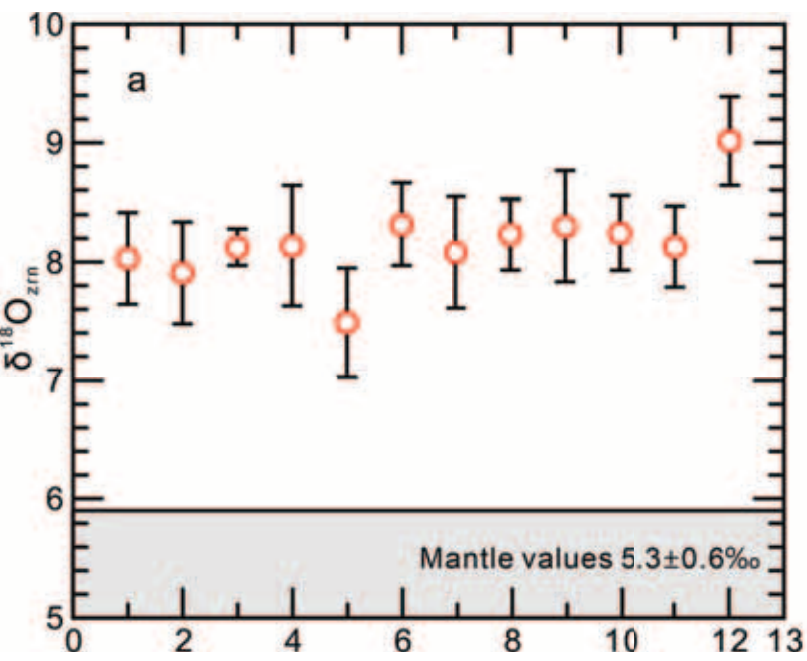


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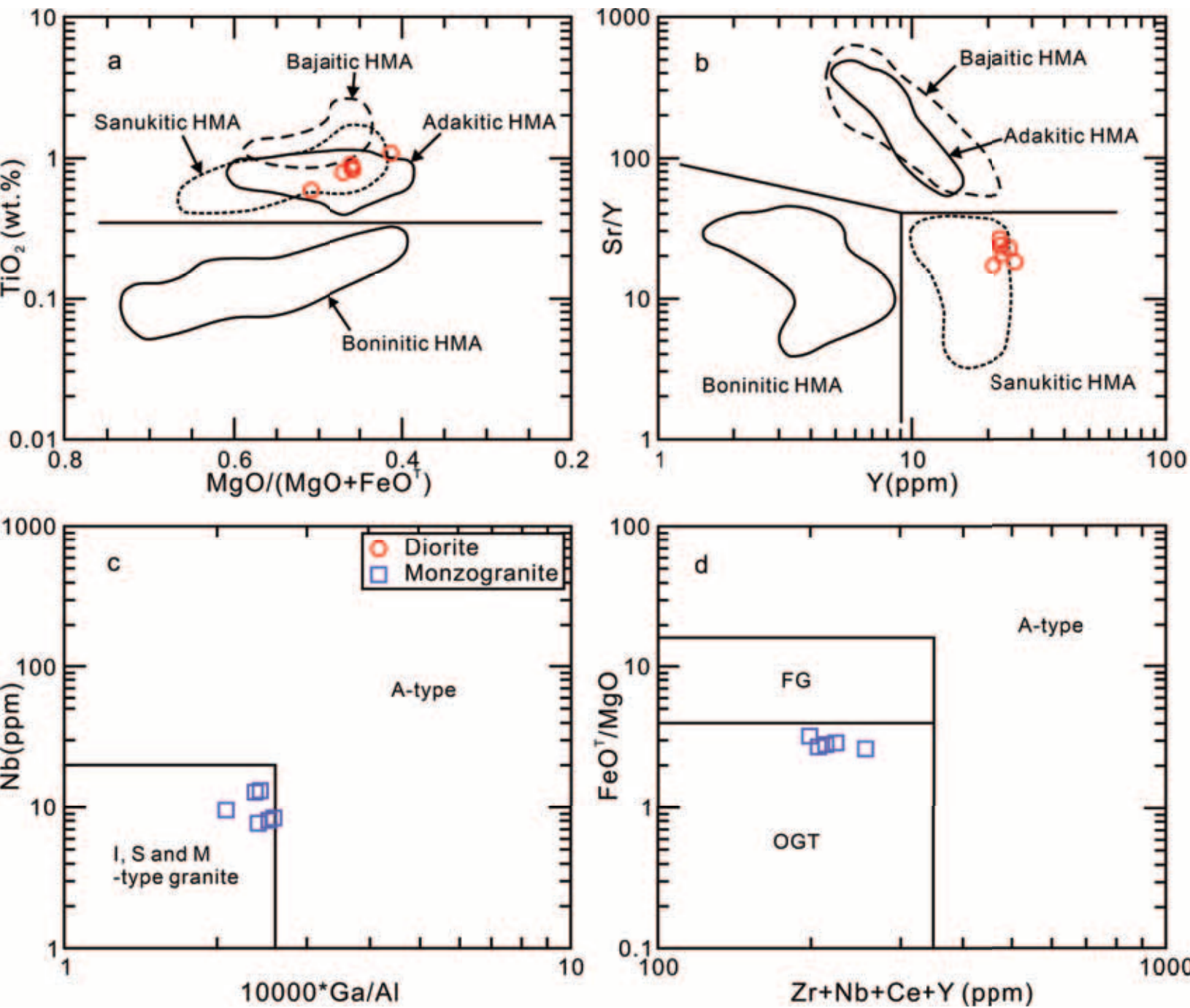


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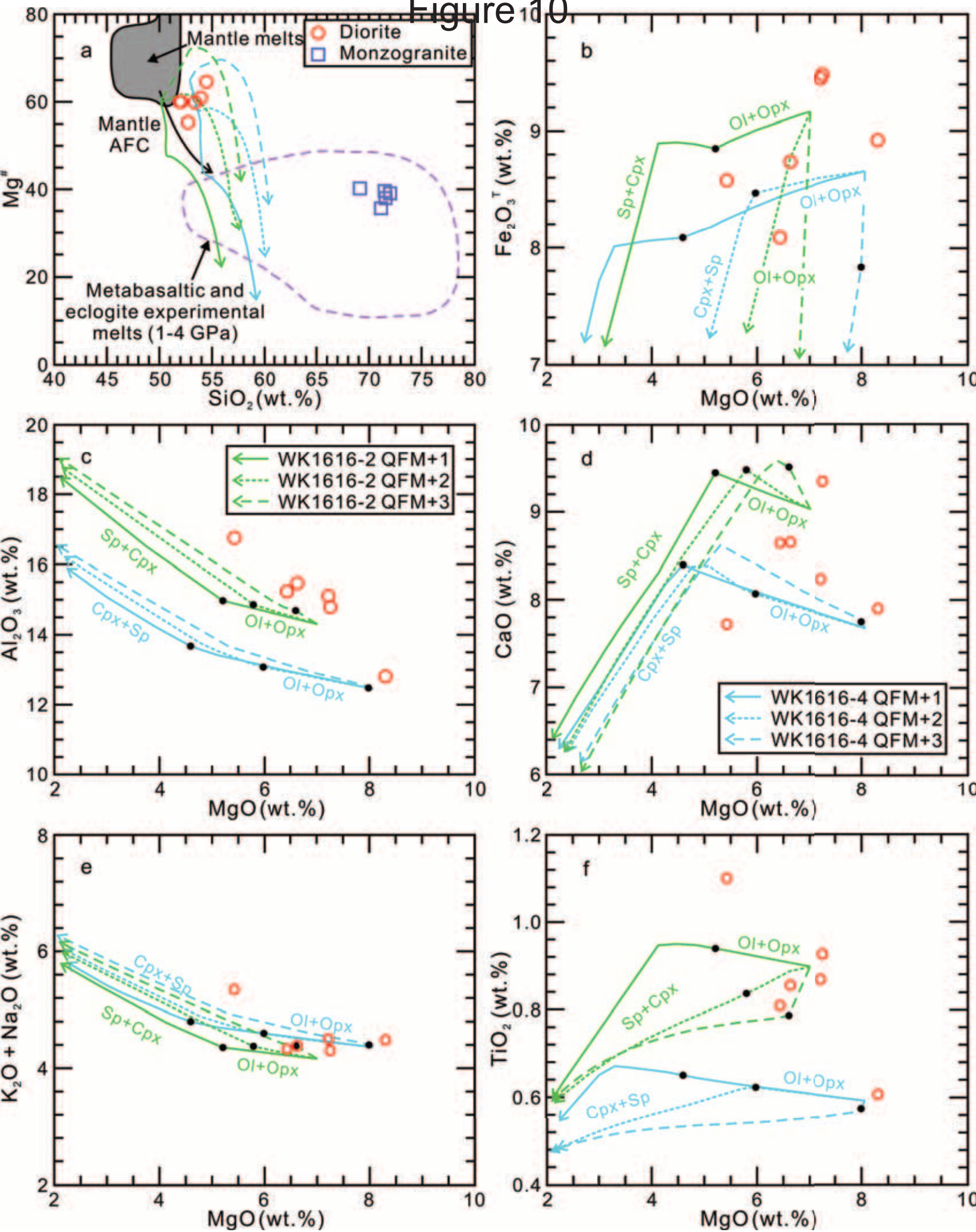


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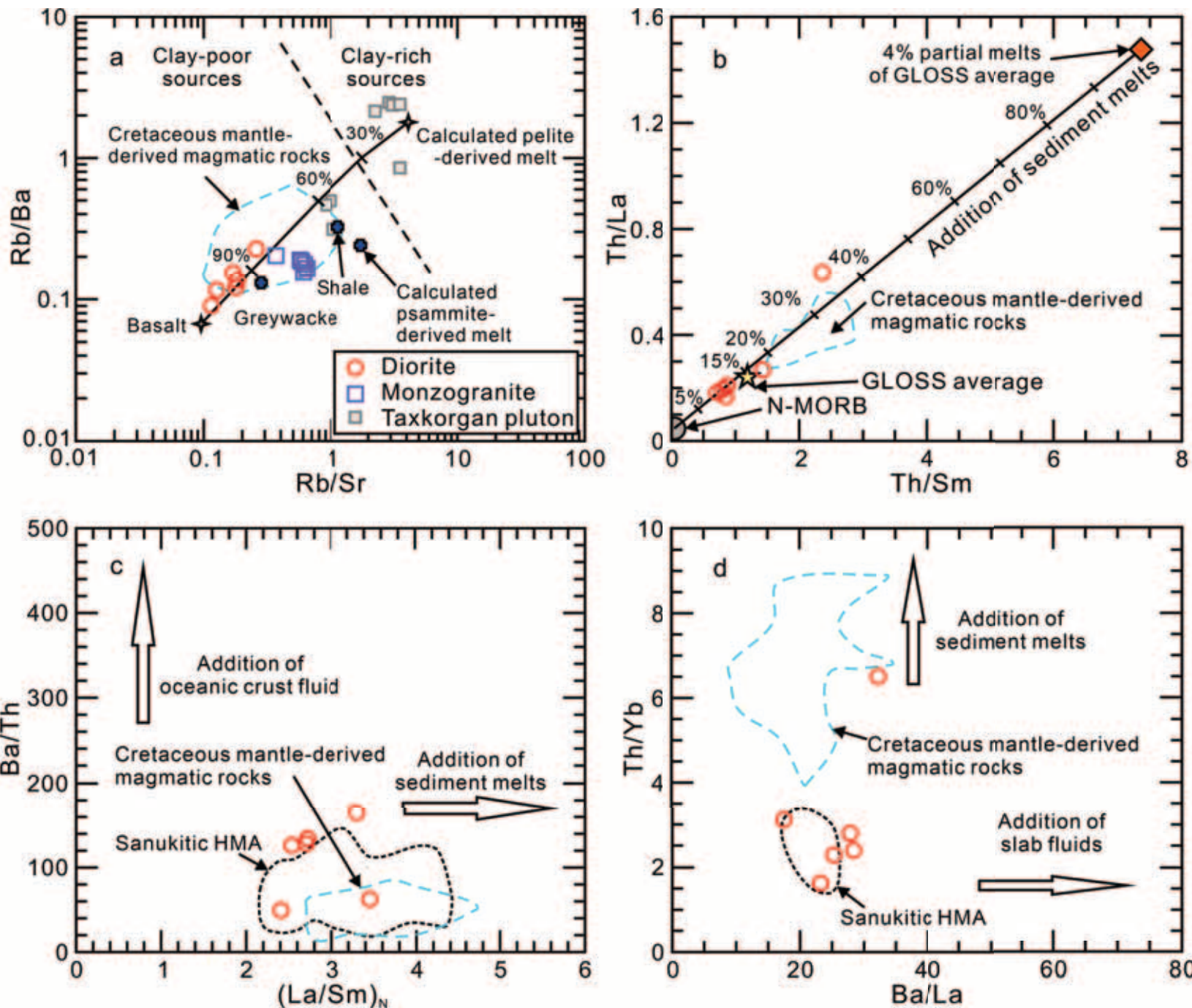


Figure 12

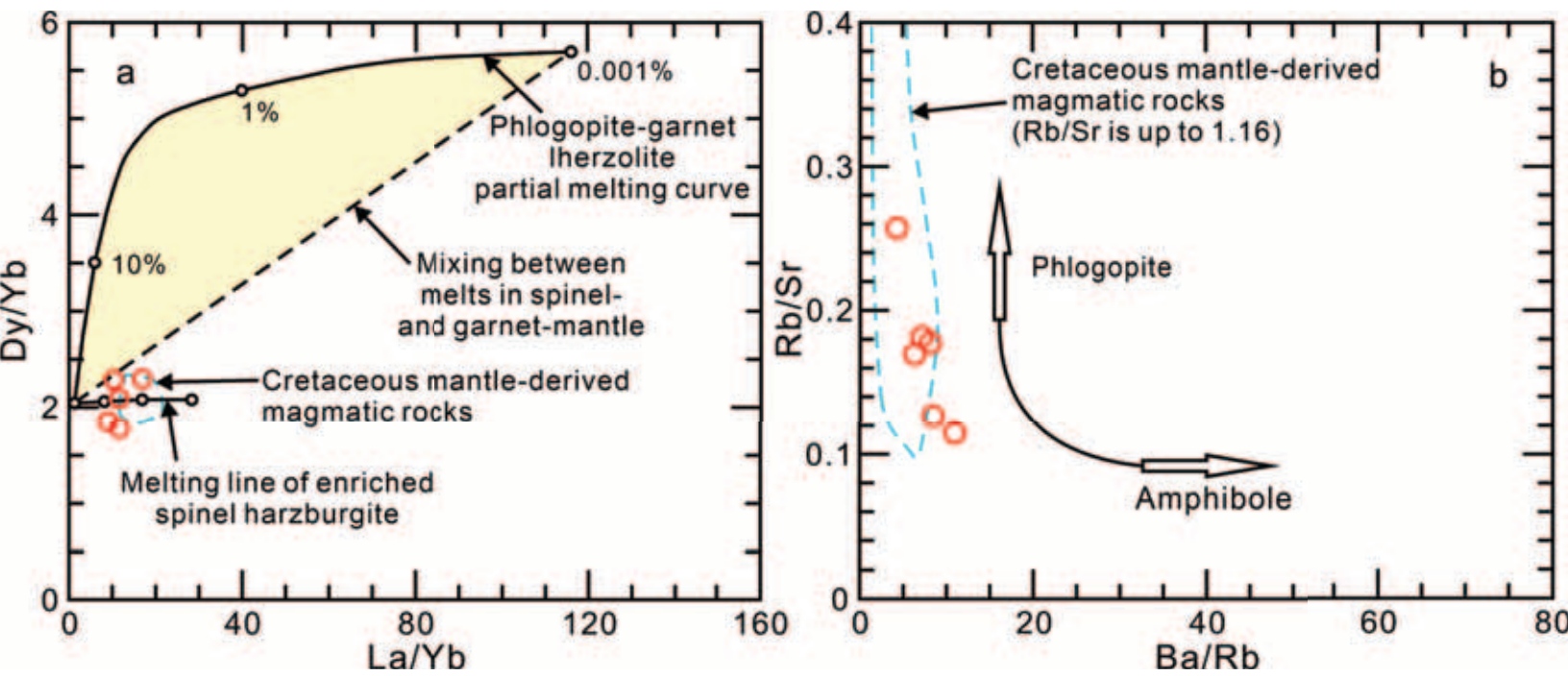


Figure 13

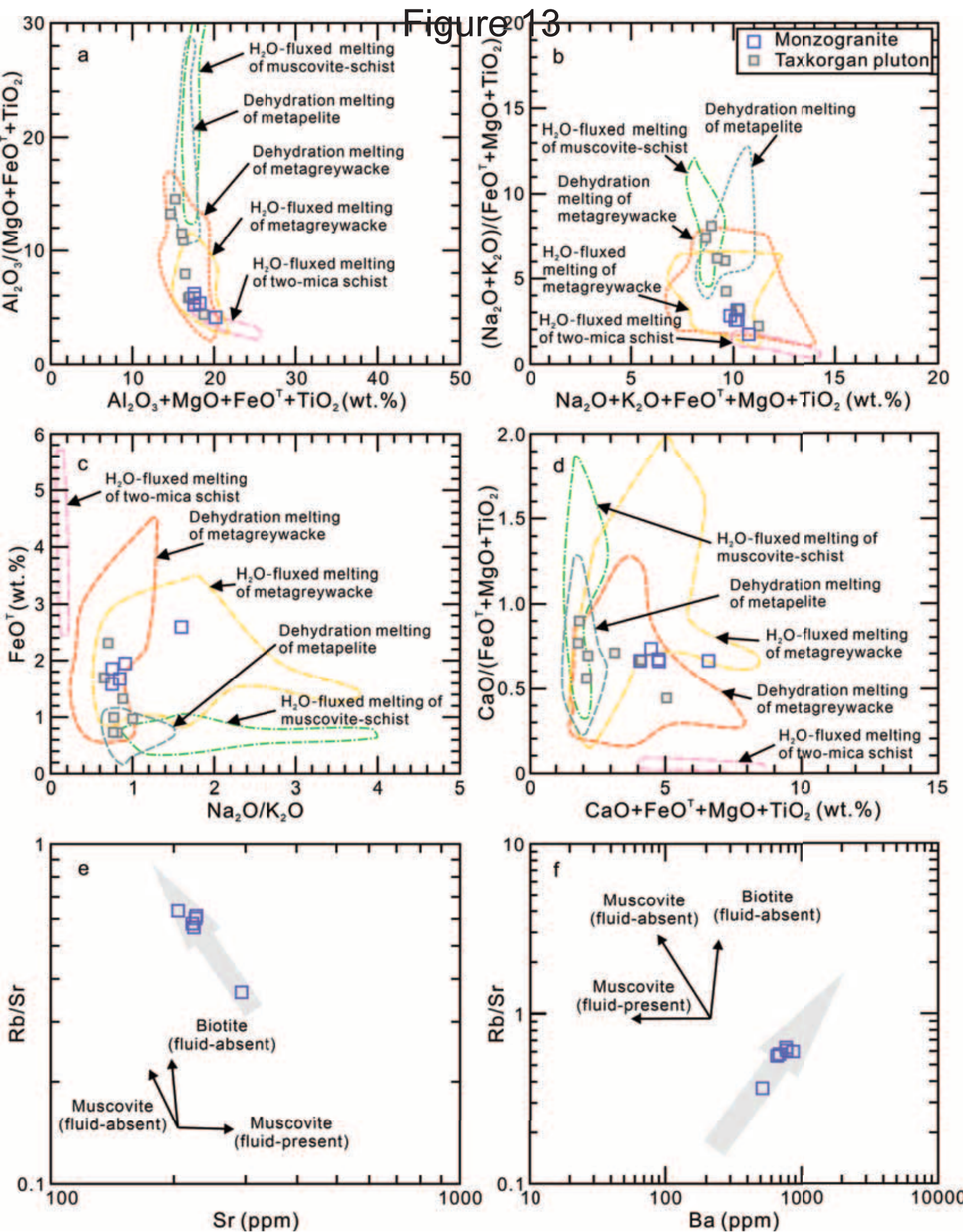


Figure 14

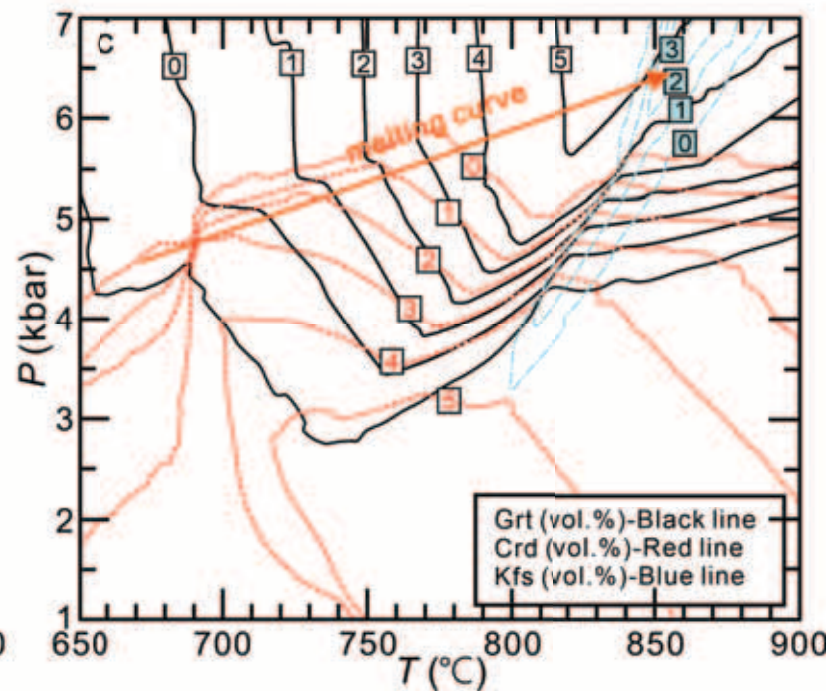
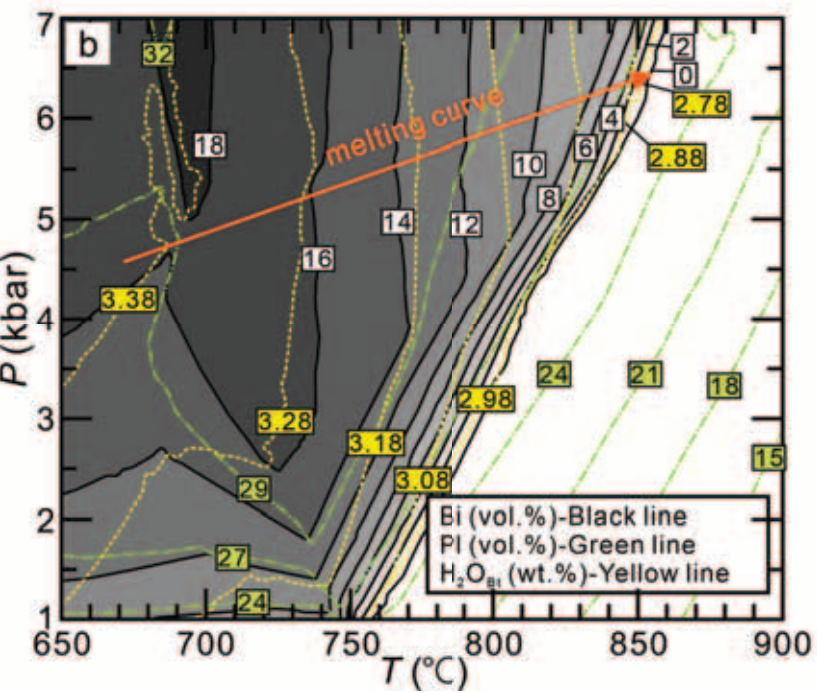
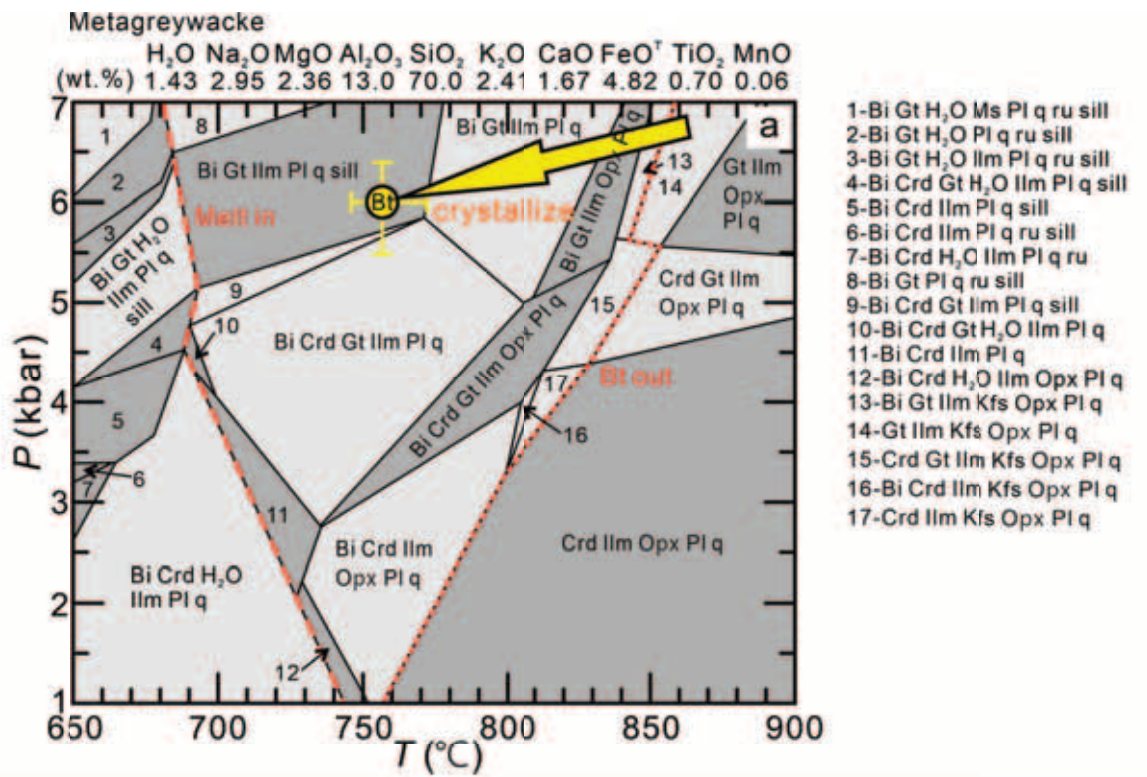


Figure 15

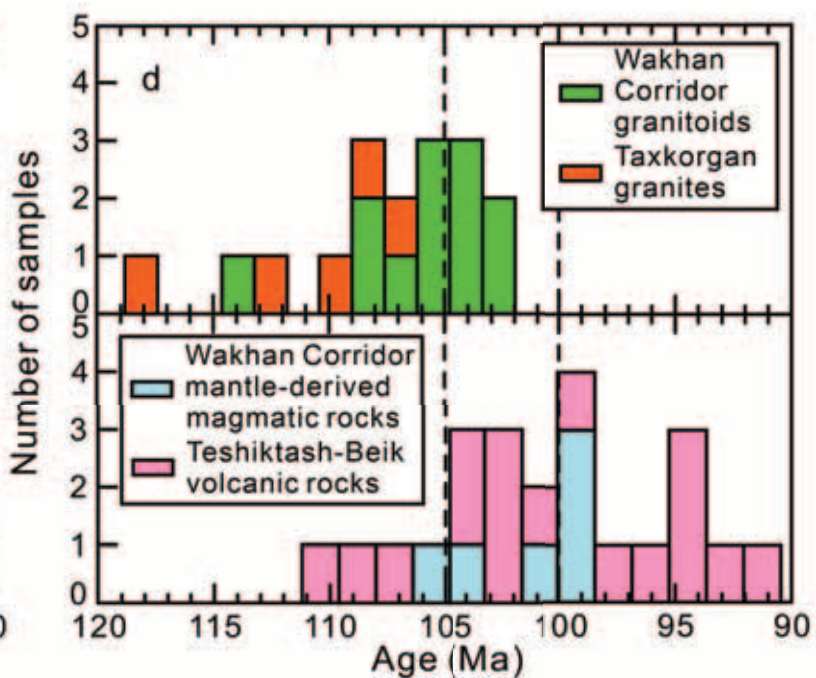
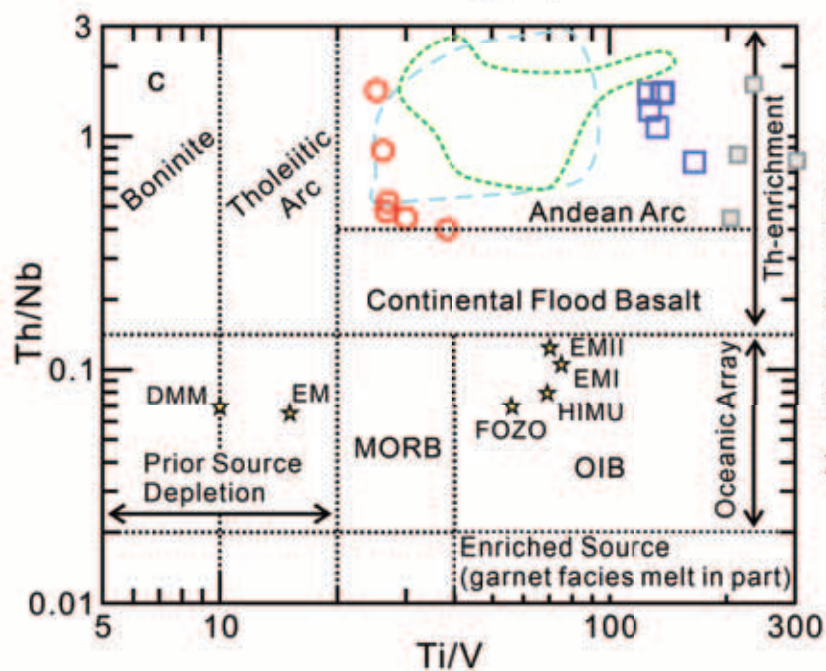
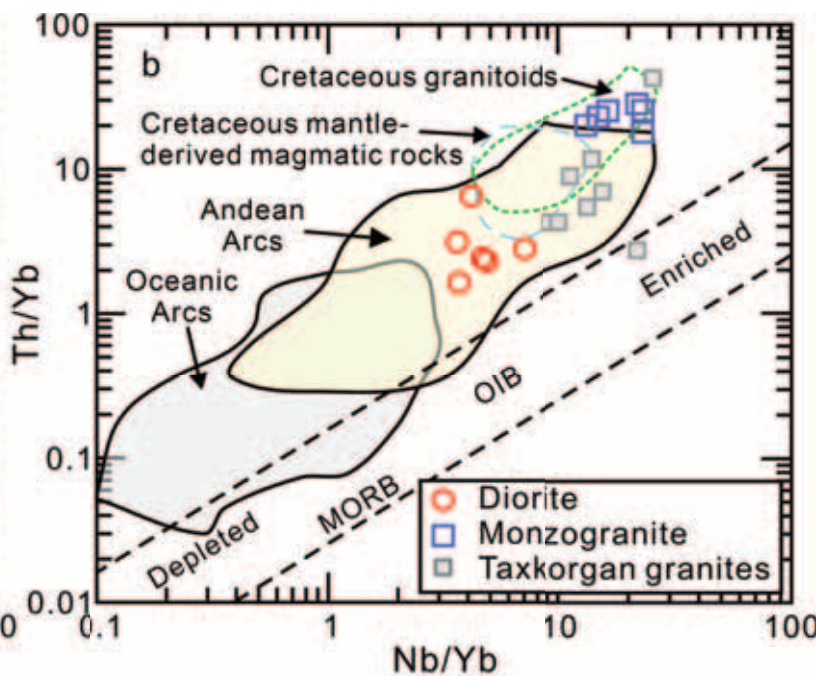
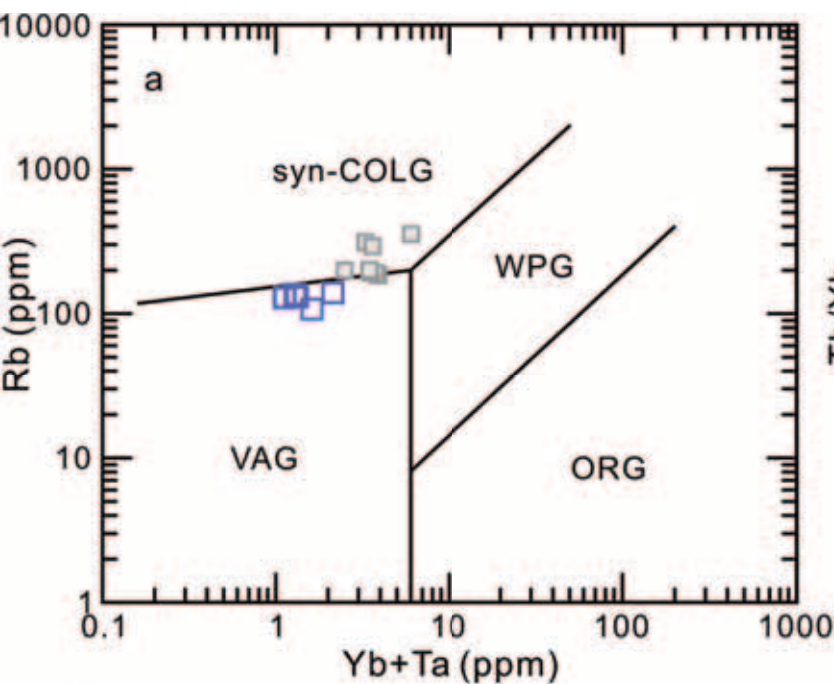
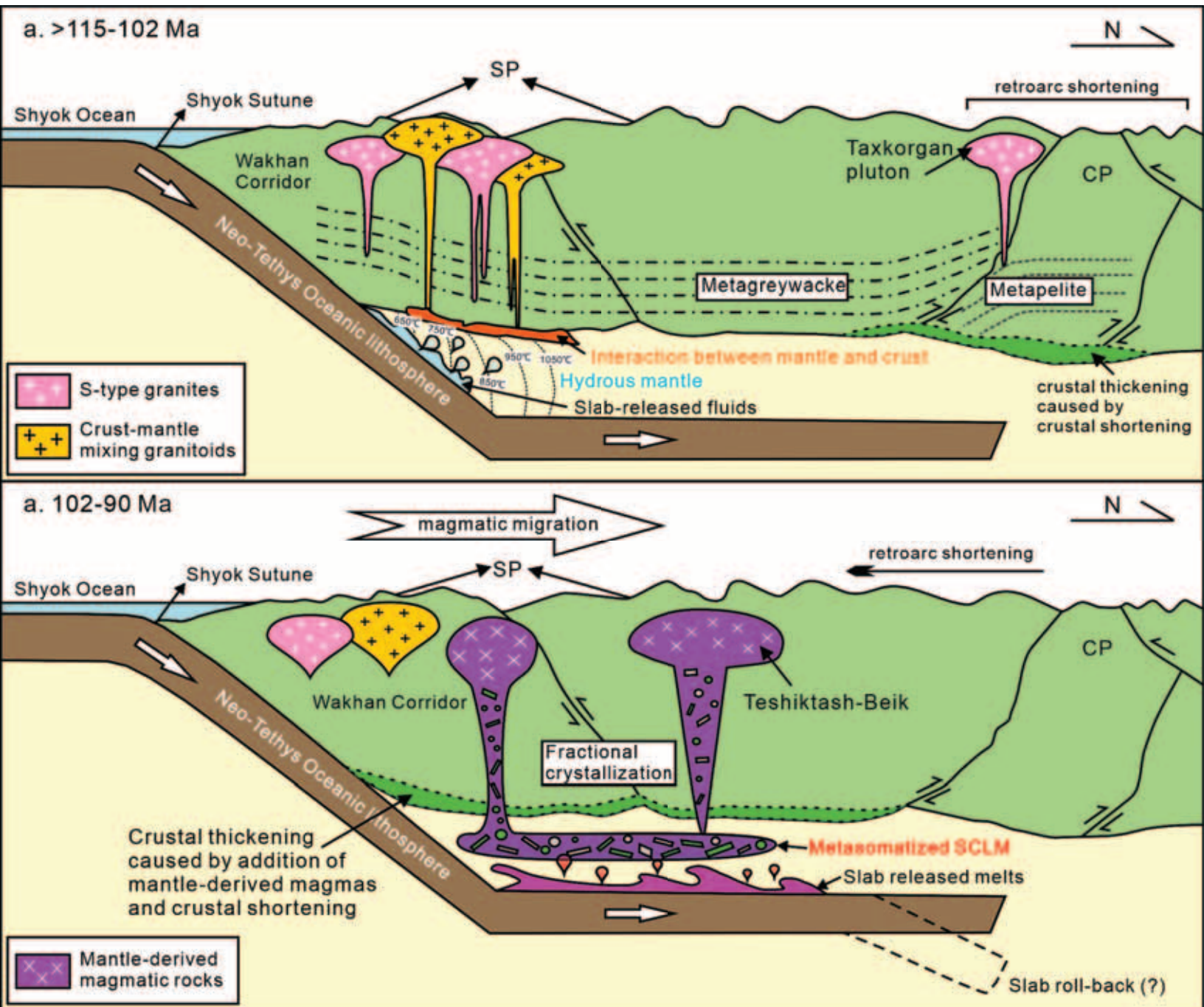


Figure 16



Supporting Information for

Early Cretaceous continental arc magmatism in the Wakhan Corridor,
South Pamir: mantle evolution and geodynamic processes during flat
subduction of the Neo-Tethys oceanic slab

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Data Repository

Supplementary File 1. Analytical methods

Supplementary Table 1. LA-ICP-MS zircon U-Pb isotopic dating data for the Kalaqigu diorite and monzogranite at the Chinese Wakhan Corridor, Southern Pamir

Supplementary Table 2. Representative electron probe analyses of amphibole, biotite and plagioclase from the Kalaqigu pluton

Supplementary Table 3. Whole-rock major (wt.%), trace (ppm) elements and Sr-Nd-Pb isotopic compositions of Kalaqigu diorite and monzogranite at the Chinese Wakhan Corridor and Taxkorgan S-type monzogranites

Supplementary Table 4. Zircon Lu-Hf isotopic compositions of Kalaqigu diorite at the Chinese Wakhan Corridor, Southern Pamir

Supplementary Table 5. Zircon O isotopic compositions of Kalaqigu diorite and monzogranite at the Chinese Wakhan Corridor, Southern Pamir

Supplementary File 1 Analytical methods

Zircon U-Pb geochronology

Zircon grains for U-Pb and Lu-Hf analysis were separated using conventional magnetic and heavy-liquid techniques, followed by hand-picking under a binocular microscope at the MC-ICPMS laboratory of the Institute of Geology and Geophysics, Chinese Academy of Sciences (IGG CAS). Photographs were taken in transmitted and reflected light, followed by cathodoluminescence (CL) imaging to reveal the internal texture of the grains and to select suitable positions for U-Pb dating and Hf isotope analysis.

Zircon U-Pb dating of two samples (WK1616 and WK1617) were carried out using LA-ICP-MS (Laser Ablation Inductively Coupled Plasma Mass Spectrometry). An Agilent 7500a ICP-MS and a Neptune multi-collector (MC-ICPMS with an attached 193 nm excimer ArF laser-ablation system (GeoLas Plus)) were used for simultaneous determination of zircon U-Pb ages. Instrumental settings and detailed analytical procedures for laser ablation system, ICP-MS instrument and data reduction have been described in Xie et al. (2008). Analyses were acquired at a beam diameter of 32 μm , an 8 Hz repetition rate, and an energy of 10–20 J/cm^2 . Helium carrier gas transported the ablated sample materials from the laser-ablation cell via a mixing chamber to the ICPMS. Every spot analysis consisted of ~30 s background acquisition and 40 s sample data acquisition. The zircons 91500 and GJ-1 were used as an external standard and internal standard, respectively. The glass NIST 610 was used as an external standard for trace element compositions calibration. Off-line raw data

selection, integration of background and analytical signals, time-drift correct, and quantitative calibration of U-Pb isotopes were performed using ICPMSDataCal software (Liu et al., 2009). Concordia diagrams and weighted mean calculations were made using ISOPLOT 3.00 program (Ludwig, 2003).

Zircon Lu-Hf isotope analysis

Zircons showing concordant U-Pb ages were selected for in-situ zircon Lu-Hf isotopes in the same dated domains. They were subsequently analyzed using Laser Ablation (LA)-ICPMS at the MC-ICPMS laboratory of IGG CAS, using a beam size of 60 μm (8 Hz laser pulse frequency). Details of instrumental conditions and data acquisition have been given in Wu et al. (2006). During the analytical period, a weighted $^{206}\text{Pb}/^{238}\text{U}$ age and a weighted $^{176}\text{Hf}/^{177}\text{Hf}$ ratio of the sample GJ-1 were determined at 609.7 ± 6.3 Ma (2σ , MSWD = 0.97, $n = 12$) and 0.282015 ± 0.000003 (2σ , MSWD = 1.12, $n = 94$), which are in good agreement with the recommended U-Pb age and Hf isotopic ratios (Black et al., 2003; Wu et al., 2006). The $^{176}\text{Hf}/^{177}\text{Hf}$ ratios of the standard zircon (MUD) were measured to be 0.282504 ± 0.000003 (2σ , MSWD = 0.71, $n = 82$), and it was used for data acquisition of Hf isotopes.

Zircon O isotope analysis

Zircon oxygen isotopes were measured using the Cameca IMS-1280 HR secondary ion mass spectrometer (SIMS) at the at State Key Laboratory of Isotope Geochemistry (SKLaBIG), Guangzhou Institute of Geochemistry, Chinese Academy

of Sciences (GIG CAS). Detailed analytical procedures are described in Li et al. (2010a) and Yang et al. (2018). The measured oxygen isotopic values were corrected for instrumental mass fractionation factor (IMF) using the standard Penglai zircon with $\delta^{18}\text{O}_{\text{VSMOW}} = 5.3 \pm 0.10 \text{ ‰}$ (2σ) and Qinghu standards with $5.4 \pm 0.2 \text{ ‰}$ (2σ) (Li et al., 2010b). The internal precision of single analysis was better than 0.1 ‰ (1σ) for $\delta^{18}\text{O}$ values. Uncertainties of analytical $\delta^{18}\text{O}$ values are quoted at 2σ level. The external precision (0.50 ‰ ; 2SD , $n = 68$), measured by spot-to-spot reproducibility of repeated analyses of the PengLai standard, was adopted for data evaluation.

Mineral geochemistry analysis

Major elements analyses of amphibole, plagioclase and biotite were carried out using JEOL JXA-8100 Electron Probe Micro Analyzer (EPMA) at IGG CAS. A beam current of 20 nA at 15 kV accelerating voltage, a beam size of 5 μm and a counting time of 30 s were used to analyze minerals. The analytical precision for all elements is better than 1.5%.

Whole-rock geochemistry analysis

Representative samples selected on the basis of optical microscopy were cleaned, crushed and homogenization, and then powdered to ~ 200 -mesh size using an agate mill. The resulting powder was used for analyses of major and trace elements, and Sr-Nd-Pb isotopes. Determination of loss on ignition (LOI) was performed at $1000 \text{ }^\circ\text{C}$. Major-element oxides were analyzed using a Rigaku RIX 2000 X-ray fluorescence

spectrometer (XRF) on fused glass beads at SKLaBIG, GIG CAS. Details of procedures are described by Yuan et al. (2010). Analytical uncertainty for major elements is generally < 5%

Trace element concentrations, including rare earth element (REE) concentrations, were determined using a Perkin-Elmer ELAN-DRC-e inductively coupled plasma mass spectrometer at the State Key Laboratory of Ore Deposit Geochemistry (SKLOG), with analytical uncertainty better than 10%. The analytical precision is generally better than 5% for most trace elements. The analytical procedures for the trace elements were described in detail in Liang et al. (2000).

Whole-rock Sr-Nd-Pb isotope analysis

Sr-Nd-Pb isotopic compositions of selected samples were determined using a MC-ICP-MS at SKLaBIG, GIG CAS. Sr and Nd were separated using cation columns, and Nd fractions were further separated by HDEHP-coated Kef columns. Analytical procedures are similar to those described in Wei et al. (2002) and Li et al. (2004). The $^{87}\text{Sr}/^{86}\text{Sr}$ ratio of the NBS987 standard and $^{143}\text{Nd}/^{144}\text{Nd}$ ratio of the Shin Etsu Jndi-1 standard measured are 0.710285 ± 15 (2σ) and 0.512085 ± 10 (2σ), respectively. Measured $^{87}\text{Sr}/^{86}\text{Sr}$ and $^{143}\text{Nd}/^{144}\text{Nd}$ ratios were corrected for fractionation using ratios of $^{87}\text{Sr}/^{86}\text{Sr} = 0.1194$ and $^{143}\text{Nd}/^{144}\text{Nd} = 0.7219$, respectively. A total of 50 mg powder was weighed into a Teflon beaker and dissolved in concentrated HF at 180 °C for 3 days to determinate Pb isotope. Lead was separated and purified by conventional cation-exchange techniques with diluted HBr as an eluant. Analytical procedures for

Pb isotopic compositions were described in Zhu et al. (2001). Pb isotope fractionations were corrected using correction factors based on replicate analyses of international standard NBS981. The results measured for NBS981 are $0.059135 \pm 0.021\%$ (2σ) for $^{204}\text{Pb}/^{206}\text{Pb}$, $0.914174 \pm 0.010\%$ (2σ) for $^{207}\text{Pb}/^{206}\text{Pb}$, and $2.161430 \pm 0.016\%$ (2σ) for $^{208}\text{Pb}/^{206}\text{Pb}$.

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Supplementary Table 1 LA-ICP-MS zircon U-Pb isotopic dating data for the Kalaqigu diorite and monzogranite at the Chinese Wakhan Corridor, Southern Pamir

	Pb	Th	U	Th/U	²⁰⁷ Pb/ ²⁰⁶ Pb	²⁰⁷ Pb/ ²⁰⁶ Pb	²⁰⁷ Pb/ ²³⁵ U	²⁰⁷ Pb/ ²³⁵ U	²⁰⁶ Pb/ ²³⁸ U	²⁰⁶ Pb/ ²³⁸ U	²⁰⁷ Pb/ ²⁰⁶ Pb	²⁰⁷ Pb/ ²⁰⁶ Pb	²⁰⁷ Pb/ ²³⁵ U	²⁰⁷ Pb/ ²³⁵ U	²⁰⁶ Pb/ ²³⁸ U	²⁰⁶ Pb/ ²³⁸ U
	ppm	ppm	ppm	Ratio	Ratio	1sigma	Ratio	1sigma	Ratio	1sigma	Age (Ma)	1sigma	Age (Ma)	1sigma	Age (Ma)	1sigma
Diorite (WK1616)																
WK1616-02	45	180	788	0.23	0.04434	0.00447	0.10446	0.01042	0.01712	0.0003	-53	174	101	10	109	2
WK1616-03	53	305	626	0.49	0.04264	0.00574	0.09489	0.01268	0.01617	0.0003	-142	198	92	12	103	2
WK1616-05	77	504	763	0.66	0.05439	0.0047	0.12164	0.01037	0.01625	0.00027	387	163	117	9	104	2
WK1616-06	72	463	753	0.62	0.0416	0.00624	0.09367	0.01392	0.01636	0.00037	-199	207	91	13	105	2
WK1616-07	64	419	748	0.56	0.04897	0.00494	0.10885	0.01086	0.01615	0.00028	146	191	105	10	103	2
WK1616-08	19	114	260	0.44	0.04597	0.01294	0.10719	0.02999	0.01694	0.00056	-4	409	103	28	108	4
WK1616-09	44	340	380	0.89	0.05606	0.00903	0.12551	0.01999	0.01626	0.00042	455	307	120	18	104	3
WK1616-10	82	578	883	0.66	0.04646	0.0043	0.10506	0.00961	0.01642	0.00027	22	171	101	9	105	2
WK1616-11	65	408	822	0.50	0.0468	0.00442	0.10639	0.00994	0.01651	0.00027	39	176	103	9	106	2
WK1616-12	95	674	1021	0.66	0.04363	0.00379	0.10242	0.0088	0.01705	0.00026	-90	153	99	8	109	2
WK1616-14	50	314	601	0.52	0.04152	0.00581	0.09644	0.0134	0.01687	0.00032	-204	200	93	12	108	2
WK1616-15	71	404	954	0.42	0.0507	0.00371	0.11825	0.00854	0.01694	0.00025	227	137	113	8	108	2
WK1616-16	68	478	762	0.63	0.04633	0.00458	0.10418	0.0102	0.01633	0.00027	15	183	101	9	104	2
WK1616-18	69	466	669	0.70	0.04651	0.00525	0.10857	0.01215	0.01695	0.00031	24	207	105	11	108	2
WK1616-21	65	434	801	0.54	0.03941	0.00442	0.09018	0.01003	0.01662	0.00028	-324	193	88	9	106	2
WK1616-22	71	507	714	0.71	0.05342	0.00501	0.12039	0.01116	0.01636	0.00029	347	177	115	10	105	2
WK1616-24	53	342	667	0.51	0.05019	0.0055	0.11453	0.01241	0.01657	0.00031	204	210	110	11	106	2
WK1616-25	37	226	501	0.45	0.03683	0.00785	0.08266	0.01752	0.01629	0.00041	-1	295	81	16	104	3
Monzogranite WK1617																
WK1617-02	77	341	1396	0.24	0.04662	0.00332	0.10803	0.00755	0.01682	0.00026	30	123	104	7	108	2
WK1617-04	63	480	362	1.33	0.05401	0.00901	0.13039	0.02152	0.01753	0.00045	371	319	124	19	112	3
WK1617-05	499	605	604	1.00	0.05999	0.00148	0.83847	0.02002	0.10147	0.00111	603	33	618	11	623	6
WK1617-06	156	109	157	0.69	0.07098	0.00296	1.53662	0.06251	0.15719	0.00231	957	59	945	25	941	13
WK1617-08	300	231	338	0.68	0.07041	0.00212	1.23401	0.03609	0.12725	0.00158	940	40	816	16	772	9
WK1617-09	105	67	118	0.57	0.07361	0.00398	1.6198	0.0857	0.15978	0.00264	1031	80	978	33	956	15
WK1617-10	96	93	143	0.65	0.06346	0.00624	0.96032	0.09252	0.10988	0.00269	724	164	683	48	672	16
WK1617-11	173	189	163	1.16	0.0615	0.00546	0.79957	0.06937	0.09441	0.00222	657	147	597	39	582	13
WK1617-13	36	218	428	0.51	0.04039	0.01007	0.09209	0.02282	0.01656	0.00049	-267	328	89	21	106	3

WK1617-14	509	672	503	1.33	0.05992	0.00175	0.78631	0.0226	0.0953	0.00109	601	42	589	13	587	6
WK1617-16	72	527	383	1.38	0.0498	0.01074	0.11634	0.02487	0.01697	0.00051	186	347	112	23	108	3
WK1617-17	17	112	108	1.03	0.05251	0.04327	0.13531	0.11073	0.01872	0.00183	308	1159	129	99	120	12
WK1617-18	104	52	705	0.07	0.05515	0.00195	0.40354	0.01418	0.05315	0.00059	418	59	344	10	334	4
WK1617-19	500	987	1293	0.76	0.05734	0.00122	0.51134	0.01105	0.06477	0.00061	505	31	419	7	405	4
WK1617-20	57	396	299	1.33	0.06046	0.01558	0.13778	0.03512	0.01655	0.00065	620	457	131	31	106	4
WK1617-21	403	191	511	0.37	0.07364	0.0014	1.77078	0.03497	0.17469	0.00168	1032	25	1035	13	1038	9
WK1617-23	449	320	560	0.57	0.0687	0.00152	1.35344	0.03072	0.14312	0.00146	890	30	869	13	862	8
WK1617-24	326	424	907	0.47	0.05638	0.00187	0.55852	0.01855	0.07197	0.0008	467	54	451	12	448	5
WK1617-25	43	289	307	0.94	0.04322	0.01616	0.10257	0.03813	0.01724	0.00072	-111	514	99	35	110	5

Supplementary Table 2-1 Representative electron probe analyses of amphibole from the Kalaqigu high-Mg diorite

Sample	Locality	Lithology	Spot no.	Comment	SiO ₂	TiO ₂	Al ₂ O ₃	Cr ₂ O ₃	TFeO	MnO	MgO	CaO	Na ₂ O	K ₂ O	Ni	Total	Mg [#]
WK1616	Kalaqigu	High-Mg diorite	1	core	45.58	1.44	9.30	0.03	14.63	0.28	12.04	11.64	1.16	0.97	0.00	97.06	59
WK1616	Kalaqigu	High-Mg diorite	2		↓	43.97	1.65	10.05	0.06	16.08	0.31	11.33	11.83	1.28	1.32	0.01	97.87
WK1616	Kalaqigu	High-Mg diorite	3	rim	43.19	2.08	11.43	0.09	15.35	0.29	10.78	11.48	1.53	1.22	0.05	97.44	56
WK1616	Kalaqigu	High-Mg diorite	4		↓	45.93	1.08	8.45	0.08	15.50	0.29	12.46	11.88	1.07	1.03	0.01	97.76
WK1616	Kalaqigu	High-Mg diorite	5	core	48.69	0.87	6.34	0.00	14.29	0.34	13.16	12.15	0.74	0.71	0.00	97.29	62
WK1616	Kalaqigu	High-Mg diorite	6		↓	43.28	1.74	10.51	0.06	16.97	0.27	10.37	11.96	1.16	1.43	0.02	97.77
WK1616	Kalaqigu	High-Mg diorite	7	rim	44.32	1.91	10.28	0.04	15.69	0.30	10.75	11.53	1.23	1.03	0.00	97.08	55
WK1616	Kalaqigu	High-Mg diorite	8		↓	47.43	0.85	7.33	0.04	15.42	0.33	12.93	11.63	0.95	0.83	0.00	97.74

Strutural formulae is calculated for 23 oxygens with Fe²⁺/Fe³⁺ estimation assuming Σ15 cations (Leake et al., 1997)

Sample	Si	Al ^{IV}	Tsite	Al ^{VI}	Ti	Cr ³⁺	Fe ³⁺	Mg	Fe ²⁺	Mn ²⁺	Csite	Fe ²⁺	Mn ²⁺	Ca	Na	Bsite	Na	K	Asite	Mg/(Mg+Fe ²⁺)	Temperature (°C)	fO ₂	ΔFMQ	Pressure (kbar)	H ₂ O _{melt} (wt.%)
WK1616	6.77	1.24	8.00	0.39	0.16	0.00	0.16	2.66	1.62	0.00	5.00	0.03	0.04	1.85	0.08	2.00	0.26	0.18	0.44	0.62	847	-12.21	1.22	6.07	6.17
WK1616	6.54	1.46	8.00	0.31	0.19	0.01	0.27	2.51	1.71	0.00	5.00	0.01	0.04	1.89	0.06	2.00	0.31	0.25	0.56	0.59	880	-11.85	0.89	4.66	5.40
WK1616	6.46	1.54	8.00	0.47	0.23	0.01	0.09	2.40	1.79	0.00	5.00	0.04	0.04	1.84	0.09	2.00	0.36	0.23	0.59	0.57	907	-11.71	0.45	7.43	6.60
WK1616	6.77	1.23	8.00	0.24	0.12	0.01	0.37	2.74	1.52	0.00	5.00	0.02	0.04	1.88	0.07	2.00	0.24	0.19	0.43	0.64	834	-12.20	1.54	3.59	5.24
WK1616	7.16	0.84	8.00	0.26	0.10	0.00	0.14	2.89	1.62	0.00	5.00	0.00	0.04	1.91	0.05	2.00	0.17	0.13	0.30	0.64	771	-13.28	1.89	3.86	5.40
WK1616	6.49	1.51	8.00	0.35	0.20	0.01	0.23	2.32	1.90	0.00	5.00	0.00	0.04	1.92	0.04	2.00	0.30	0.27	0.57	0.55	889	-12.01	0.53	5.38	5.82
WK1616	6.64	1.37	8.00	0.45	0.22	0.01	0.09	2.40	1.84	0.00	5.00	0.03	0.04	1.85	0.08	2.00	0.28	0.20	0.47	0.57	869	-12.30	0.64	7.02	6.86
WK1616	6.96	1.04	8.00	0.22	0.09	0.01	0.39	2.83	1.46	0.00	5.00	0.04	0.04	1.83	0.09	2.00	0.18	0.16	0.33	0.66	795	-12.73	1.88	3.26	5.19

TFeO = Total FeO content; Mg[#] = 100*Mg²⁺/(Mg²⁺ + TFe²⁺)

Temperature, fO₂ and ΔFMQ values calculated using the formulations of Ridolfi et al. (2010).

Pressure calculated using an extended calibration of the Larocque and Canil (2010) barometer published by Krawczynski et al. (2012).

H₂O_{melt} (wt.%) calculated using the formulations of Ridolfi et al. (2010).

Supplementary Table 2-2 Representative electron probe analyses of biotite from the Kalaqigu pluton

Sample	Locality	Lithology	Spot no.	Comment	SiO ₂	TiO ₂	Al ₂ O ₃	Cr ₂ O ₃	TFeO	MnO	MgO	CaO	Na ₂ O	K ₂ O	Ni	Total	Mg [#]
WK1616	Kalaqigu	High-Mg diorite	1	core	45.58	1.44	9.30	0.03	14.63	0.28	12.04	11.64	1.16	0.97	0.00	97.06	59
WK1616	Kalaqigu	High-Mg diorite	2		↓	43.97	1.65	10.05	0.06	16.08	0.31	11.33	11.83	1.28	1.32	0.01	97.87
WK1616	Kalaqigu	High-Mg diorite	3	rim	43.19	2.08	11.43	0.09	15.35	0.29	10.78	11.48	1.53	1.22	0.05	97.44	56
WK1616	Kalaqigu	High-Mg diorite	4		↓	45.93	1.08	8.45	0.08	15.50	0.29	12.46	11.88	1.07	1.03	0.01	97.76

WK1616	Kalaqigu	High-Mg diorite	5	core ↓ rim	48.69	0.87	6.34	0.00	14.29	0.34	13.16	12.15	0.74	0.71	0.00	97.29	62
WK1616	Kalaqigu	High-Mg diorite	6		43.28	1.74	10.51	0.06	16.97	0.27	10.37	11.96	1.16	1.43	0.02	97.77	52
WK1616	Kalaqigu	High-Mg diorite	7		44.32	1.91	10.28	0.04	15.69	0.30	10.75	11.53	1.23	1.03	0.00	97.08	55
WK1616	Kalaqigu	High-Mg diorite	8		47.43	0.85	7.33	0.04	15.42	0.33	12.93	11.63	0.95	0.83	0.00	97.74	60

Struturale formulae is calculated for 23 oxygens with Fe²⁺/Fe³⁺ estimation assuming $\sum 15$ cations (Leake et al, 1997).

Sample	Si	Al ^{IV}	Tsite	Al ^{VI}	Ti	Cr ³⁺	Fe ³⁺	Mg	Fe ²⁺	Mn ²⁺	Csite	Fe ²⁺	Mn ²⁺	Ca	Na	Bsite	Na	K	Asite	Mg/(Mg+Fe ²⁺)	Temperature (°C)	fO ₂	ΔFMQ	Pressure (kbar)	H ₂ O _{met} (wt.%)
WK1616	6.77	1.24	8.00	0.39	0.16	0.00	0.16	2.66	1.62	0.00	5.00	0.03	0.04	1.85	0.08	2.00	0.26	0.18	0.44	0.62	847	-12.21	1.22	6.07	6.17
WK1616	6.54	1.46	8.00	0.31	0.19	0.01	0.27	2.51	1.71	0.00	5.00	0.01	0.04	1.89	0.06	2.00	0.31	0.25	0.56	0.59	880	-11.85	0.89	4.66	5.40
WK1616	6.46	1.54	8.00	0.47	0.23	0.01	0.09	2.40	1.79	0.00	5.00	0.04	0.04	1.84	0.09	2.00	0.36	0.23	0.59	0.57	907	-11.71	0.45	7.43	6.60
WK1616	6.77	1.23	8.00	0.24	0.12	0.01	0.37	2.74	1.52	0.00	5.00	0.02	0.04	1.88	0.07	2.00	0.24	0.19	0.43	0.64	834	-12.20	1.54	3.59	5.24
WK1616	7.16	0.84	8.00	0.26	0.10	0.00	0.14	2.89	1.62	0.00	5.00	0.00	0.04	1.91	0.05	2.00	0.17	0.13	0.30	0.64	771	-13.28	1.89	3.86	5.40
WK1616	6.49	1.51	8.00	0.35	0.20	0.01	0.23	2.32	1.90	0.00	5.00	0.00	0.04	1.92	0.04	2.00	0.30	0.27	0.57	0.55	889	-12.01	0.53	5.38	5.82
WK1616	6.64	1.37	8.00	0.45	0.22	0.01	0.09	2.40	1.84	0.00	5.00	0.03	0.04	1.85	0.08	2.00	0.28	0.20	0.47	0.57	869	-12.30	0.64	7.02	6.86
WK1616	6.96	1.04	8.00	0.22	0.09	0.01	0.39	2.83	1.46	0.00	5.00	0.04	0.04	1.83	0.09	2.00	0.18	0.16	0.33	0.66	795	-12.73	1.88	3.26	5.19

The Fe²⁺ and Fe³⁺ were calculated using the software of Geokit (Lu et al., 2004).

The crystallization pressures and temperatures of the biotites were estimated using formulations of Li et al. (2022).

Supplementary Table 2-3 Representative electron probe analyses of plagioclase from the Kalaqigu pluton

Sample	Lithology	Spot no.	Comment	SiO ₂	TiO ₂	Al ₂ O ₃	Cr ₂ O ₃	TFeO	MnO	MgO	CaO	Na ₂ O	K ₂ O	Ni	Total	An (mol.%)	Ab (mol.%)	Or (mol.%)
WK1616	High-Mg diorite	1	rim	55.5	0.03	28.3	0.00	0.12	0.00	0.00	10.08	5.39	0.29	0.00	99.7	50	48	2
WK1616	High-Mg diorite	2		48.7	0.00	32.9	0.00	0.09	0.01	0.00	15.1	2.49	0.07	0.01	99.3	77	23	0
WK1616	High-Mg diorite	3		47.8	0.06	33.5	0.03	0.11	0.01	0.02	15.9	2.10	0.06	0.01	99.6	80	19	0
WK1616	High-Mg diorite	4	↓	47.7	0.00	33.4	0.00	0.14	0.00	0.01	15.8	2.18	0.04	0.00	99.4	80	20	0
WK1616	High-Mg diorite	5		48.2	0.00	33.2	0.04	0.08	0.03	0.00	15.74	2.22	0.05	0.01	99.6	79	20	0
WK1616	High-Mg diorite	6	core	47.0	0.00	33.9	0.00	0.12	0.00	0.00	16.77	1.78	0.05	0.00	99.7	84	16	0
WK1616	High-Mg diorite	7		46.9	0.00	33.7	0.00	0.12	0.01	0.00	16.60	1.78	0.05	0.09	99.2	83	16	0
WK1616	High-Mg diorite	8	↓	55.4	0.01	27.9	0.00	0.17	0.02	0.01	11.4	4.34	0.28	0.03	99.5	58	40	2
WK1616	High-Mg diorite	9		56.1	0.00	27.7	0.02	0.11	0.00	0.00	9.5	5.84	0.33	0.00	99.6	47	52	2
WK1616	High-Mg diorite	10	rim	56.3	0.00	27.7	0.03	0.12	0.02	0.02	9.5	5.90	0.33	0.03	99.9	46	52	2
WK1617	Monzogranite	11	core	62.30	0.00	23.26	0.02	0.00	0.00	0.00	5.02	8.28	0.23	0.00	99.1	25	74	1
WK1617	Monzogranite	12	↓	60.21	0.02	24.79	0.00	0.02	0.03	0.00	6.00	7.62	0.32	0.01	99.0	30	68	2
WK1617	Monzogranite	13	rim	60.25	0.00	24.47	0.01	0.00	0.01	0.00	6.93	7.23	0.30	0.02	99.2	34	64	2

WK1617	Monzogranite	14	63.58	0.00	22.35	0.00	0.02	0.00	0.00	4.47	8.51	0.26	0.05	99.2	22	76	2
WK1617	Monzogranite	15	62.59	0.01	22.69	0.00	0.04	0.00	0.00	4.96	8.62	0.25	0.00	99.2	24	75	1

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Ni	18.17	22.30	22.83	34.61	15.66	7.84	1.18	1.07	0.83	1.47	1.22	1.30	1.90	4.20	2.90	4.70				
Cu	15.00	16.10	11.50	16.30	12.50	14.00	2.89	1.29	1.37	1.34	1.46	1.41								
Zn	85.70	86.00	88.60	82.80	73.40	92.10	130.00	58.90	46.40	61.80	75.00	67.00								
Ga	18.40	16.80	17.70	14.40	16.40	19.40	20.10	20.10	19.30	19.80	20.40	19.70	22.00	19.00	21.00	24.00				
Ge	1.68	1.68	1.64	1.65	1.55	1.55	1.34	1.19	1.30	1.60	1.36	1.31								
Rb	66.20	70.30	82.40	94.00	89.20	108.00	129.00	127.00	136.00	139.00	107.00	130.00	200.00	193.00	192.00	184.00	354.00	291.00	199.00	311.00
Sr	574.00	554.00	485.00	365.00	490.00	607.00	222.00	224.00	226.00	227.00	293.00	205.00	88.00	207.00	190.00	178.00	100.00	99.40	56.00	100.00
Y	24.50	22.70	25.80	20.90	23.10	22.50	9.40	9.10	8.81	10.20	9.83	7.97	12.50	15.80	15.20	9.00	11.65	13.63	9.39	13.42
Zr	80.90	56.30	64.50	71.00	60.00	101.00	124.00	118.00	115.00	109.00	145.00	129.00	47.00	88.00	122.00	231.00	37.74	48.46	38.42	31.45
Nb	10.21	9.68	9.30	7.31	8.09	13.31	8.24	7.94	7.64	13.16	12.71	9.53	20.00	18.00	20.30	21.10	16.91	17.13	11.54	14.87
Cs	1.72	1.93	3.61	2.51	3.37	3.86	2.52	2.48	2.66	2.67	3.23	2.86								
Ba	726.00	602.00	530.00	413.00	648.00	886.00	689.00	662.00	870.00	778.00	518.00	772.00	93.00	412.00	384.00	583.00	147.00	118.00	232.00	130.00
La	25.50	23.70	22.80	23.70	20.00	31.80	41.20	39.90	38.30	32.80	45.30	36.90	13.00	22.00	25.00	54.00	9.59	9.90	5.91	9.57
Ce	60.10	54.60	52.50	49.10	43.40	67.60	82.90	79.70	76.20	66.90	89.20	76.50	25.00	39.00	51.00	96.00	17.44	18.09	10.85	17.79
Pr	7.33	6.67	6.36	5.55	5.38	7.78	8.93	8.50	8.30	7.37	9.70	8.17	3.10	4.80	5.90	13.40	2.17	2.26	1.37	2.23
Nd	29.40	27.40	26.40	22.10	23.50	31.00	32.90	31.20	30.60	27.50	36.50	31.00	11.00	17.00	21.00	44.00	7.51	7.93	5.62	7.70
Sm	6.06	5.66	5.80	4.44	5.37	6.23	5.79	5.37	5.32	5.13	6.45	5.69	2.29	3.61	4.20	7.59	1.99	2.17	1.91	2.16
Eu	1.51	1.48	1.39	1.11	1.35	1.65	1.06	1.01	1.13	1.20	1.36	1.07	0.29	0.57	0.59	0.95	0.31	0.28	0.37	0.29
Gd	5.41	4.99	5.31	4.11	4.86	5.30	3.97	3.87	3.66	3.79	4.78	3.80	2.16	3.08	3.43	4.46	1.58	1.74	1.77	1.72
Tb	0.81	0.79	0.81	0.63	0.83	0.81	0.53	0.55	0.47	0.57	0.61	0.50	0.38	0.51	0.51	0.45	0.34	0.39	0.42	0.40
Dy	4.63	4.28	4.71	3.62	4.47	4.35	2.19	2.12	1.98	2.22	2.35	1.90	2.37	3.03	2.87	2.18	2.10	2.25	2.11	2.26
Ho	0.89	0.80	0.92	0.71	0.85	0.81	0.32	0.31	0.25	0.34	0.33	0.26	0.43	0.58	0.53	0.33	0.40	0.47	0.34	0.47
Er	2.51	2.31	2.64	2.16	2.35	2.26	0.75	0.79	0.73	0.75	0.81	0.65	1.25	1.63	1.44	0.83	1.06	1.31	0.68	1.27
Tm	0.35	0.33	0.38	0.31	0.32	0.33	0.08	0.09	0.09	0.09	0.09	0.08	0.20	0.27	0.23	0.12	0.21	0.26	0.11	0.26
Yb	2.21	2.00	2.54	2.03	1.95	1.89	0.50	0.53	0.58	0.57	0.55	0.44	1.29	1.60	1.46	0.83	1.27	1.71	0.53	1.62
Lu	0.32	0.30	0.37	0.28	0.26	0.28	0.06	0.06	0.06	0.07	0.07	0.05	0.17	0.23	0.20	0.11	0.18	0.25	0.08	0.23
Hf	2.60	2.07	2.14	2.42	2.23	2.51	3.27	3.30	3.19	3.13	4.02	3.70	1.52	2.76	3.43	6.16	2.06	1.90	1.79	1.58
Ta	0.72	0.71	0.69	0.82	0.60	0.74	0.63	0.58	0.71	1.57	1.07	0.89	2.16	1.92	2.29	3.08	4.74	1.92	1.96	1.66
Pb	9.98	7.98	7.57	8.50	8.96	8.82	28.60	27.80	30.40	26.90	20.80	27.00	26.00	33.00	32.00	32.00	26.52	25.15	22.44	28.57
Th	5.34	4.60	4.15	6.35	12.70	5.31	12.70	12.10	11.60	10.20	13.90	12.30	8.90	14.20	17.00	35.20	6.90	7.32	1.44	6.95
U	1.23	1.25	0.98	2.78	2.08	0.87	2.32	2.29	2.37	2.52	2.70	2.54	1.80	18.10	3.20	3.60	1.63	2.77	1.41	2.89
T _z (°C)	536	489	524	512	521	581	769	767	771	761	767	824	697	741	768	829				

$^{87}\text{Rb}/^{86}\text{Sr}$	0.3334	0.3669	1.6800	1.6392	6.5660	2.6980	2.9190	2.9890
$(^{87}\text{Sr}/^{86}\text{Sr})_i$	0.708590	0.707473	0.715400	0.715753	0.70926	0.70864	0.71435	0.70753
$^{147}\text{Sm}/^{144}\text{Nd}$	0.1245	0.1248	0.1063	0.1040	0	0	0	0
$^{143}\text{Nd}/^{144}\text{Nd}$	0.512220	0.512282	0.511936	0.511942	0.1288	0.1298	0.1229	0.1047
$\epsilon_{\text{Nd}}(t)$	-7.18	-5.97	-12.45	-12.30	0.51210	0.51209	0.51190	0.51208
$T_{\text{Nd_DM}}$ (Ma)	1592	1491	1724	1679	9	9	1	3
$T_{\text{Nd_DM}}^2$ (Ma)	1491	1393	1921	1909	-9.40	-9.60	-13.30	-9.50
$(^{206}\text{Pb}/^{204}\text{Pb})_i$	18.4337	18.3246	18.5056	18.4918				
$(^{207}\text{Pb}/^{204}\text{Pb})_i$	15.6938	15.6853	15.7495	15.7481				
$(^{208}\text{Pb}/^{204}\text{Pb})_i$	38.7664	38.6378	38.8710	38.8572				

* is from Jiang et al. (2014).

** is from Li et al. (2019).

The zircon saturation temperatures (T_{Zr}) were estimated using thermobarometers from Watson and Harrison. (1983).

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Supplementary Table 4 Lu-Hf isotopic compositions of Kalaqigu diorite at the Chinese Wakhan Corridor, Southern Pamir

Sample	Age (Ma)	$^{176}\text{Yb}/^{177}\text{Hf}$	$^{176}\text{Lu}/^{177}\text{Hf}$	2σ	$^{176}\text{Hf}/^{177}\text{Hf}$	2σ	ϵ_{Hf}	$f_{\text{Lu/Hf}}$	T_{HfDM1}	T_{HfDM2}
Diorite (WK1616)										
WK1616-01	105.9	0.014094	0.000652	0.000003	0.282428	0.000016	-9.9	-0.98	1154	1791
WK1616-02	105.9	0.014484	0.000600	0.000003	0.282363	0.000014	-12.2	-0.98	1243	1934
WK1616-03	105.9	0.017451	0.000783	0.000003	0.282420	0.000027	-10.2	-0.98	1168	1807
WK1616-04	105.9	0.020054	0.000888	0.000024	0.282415	0.000016	-10.4	-0.97	1180	1820
WK1616-05	105.9	0.018204	0.000820	0.000005	0.282403	0.000016	-10.8	-0.98	1193	1845
WK1616-06	105.9	0.024250	0.001046	0.000008	0.282412	0.000016	-10.5	-0.97	1189	1828
WK1616-07	105.9	0.026881	0.001083	0.000003	0.282395	0.000016	-11.1	-0.97	1213	1864
WK1616-08	105.9	0.018689	0.000734	0.000012	0.282378	0.000015	-11.7	-0.98	1226	1901
WK1616-09	105.9	0.011516	0.000498	0.000008	0.282411	0.000017	-10.5	-0.99	1173	1827
WK1616-10	105.9	0.018273	0.000830	0.000004	0.282417	0.000016	-10.3	-0.98	1174	1814
WK1616-11	105.9	0.018392	0.000835	0.000006	0.282403	0.000015	-10.8	-0.97	1194	1845
WK1616-12	105.9	0.020144	0.000906	0.000004	0.282405	0.000015	-10.7	-0.97	1194	1842
WK1616-13	105.9	0.019794	0.000907	0.000006	0.282398	0.000017	-11.0	-0.97	1204	1858
WK1616-14	105.9	0.018751	0.000810	0.000006	0.282367	0.000014	-12.1	-0.98	1243	1925
WK1616-15	105.9	0.022458	0.000950	0.000015	0.282415	0.000016	-10.4	-0.97	1181	1820
WK1616-16	105.9	0.022262	0.001012	0.000007	0.282387	0.000018	-11.4	-0.97	1222	1882

Supplementary Table 5 Zircon O isotopic compositions of Kalaqigu diorite and monzogranite at the Chinese Wakhan Corridor, Southern Pamir

Analysis spot	Age (Ma)	Intensity O ¹⁶	O ¹⁸ /O ¹⁶ Mean	δ ¹⁸ O (‰)	2σ
Diorite (WK1616)					
WK1616-01		979518000	0.002055	8.03	0.38
WK1616-03	103	986419500	0.002055	7.91	0.43
WK1616-04		975533400	0.002055	8.13	0.15
WK1616-05	104	976108100	0.002055	8.14	0.51
WK1616-06	105	983103700	0.002054	7.49	0.46
WK1616-07	103	983825500	0.002056	8.32	0.35
WK1616-09	104	967447600	0.002055	8.08	0.47
WK1616-10	105	952977200	0.002056	8.23	0.30
WK1616-11	106	987485400	0.002056	8.30	0.47
WK1616-12	109	969187100	0.002056	8.24	0.31
WK1616-14	108	962150800	0.002055	8.13	0.34
WK1616-15	108	942879200	0.002057	9.01	0.37
Monzogranite (WK1617)					
WK1617-01		1523824000	0.002032	8.85	0.25
WK1617-02	108	1543593000	0.002030	8.00	0.17
WK1617-03		1549577000	0.002031	8.47	0.25
WK1617-04	112	1554741000	0.002031	8.25	0.25
WK1617-05	623	1565975000	0.002031	8.31	0.25
WK1617-06	941	1569459000	0.002030	7.97	0.24
WK1617-07		1571741000	0.002030	7.90	0.20
WK1617-08	772	1566809000	0.002030	7.99	0.23
WK1617-09	956	1576714000	0.002031	8.49	0.27
WK1617-10	672	1578725000	0.002030	7.82	0.23