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Post-collisional Ediacaran A2-type granitic magmatism
at the southern margin of the Saharan Metacraton**

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1 **Crystalline inliers near Lake Iro (SE Chad): post-collisional Ediacaran**
2 **A₂-type granitic magmatism at the southern margin of the Saharan**
3 **Metacraton**

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18

19 **Abstract**

20 The structure and evolution of the continental crust in the southern part of the Saharan
21 Metacraton (central Africa) is poorly known due to extensive laterite and sediment
22 cover as well as geographic isolation. We report on a series of five crystalline inliers
23 newly discovered near Lake Iro (south-eastern Chad) that offered the opportunity to

24 unravel the nature of the local basement in this area. Two spatially determined
25 petrographic types are exposed in the inliers: a medium-grained porphyritic
26 amphibole–biotite granite and a biotite microgranite with typical embayed (rhyolitic)
27 quartz phenocrysts, both containing fine-grained melanocratic igneous enclaves.
28 Some porphyritic granites display evidence for low temperature deformation along N60
29 shear zones. The Lake Iro igneous rocks are typified by their ferroan mineralogy, define
30 a ferroan alkali-calcic metaluminous to weakly peraluminous potassic association, and
31 exhibit elevated HFSE (Zr, Nb, Y) and REE contents. Therefore, they are best
32 classified as A₂-type granites. They are variably ferromagnetic with total magnetization
33 lower than 1 A/m. The Lake Iro granites most likely correspond to a single subvolcanic–
34 plutonic silicic complex because: (i) first-order geochemical modelling indicates that
35 the microgranites can be generated by 30% fractional crystallization of a two feldspar–
36 amphibole assemblage from a porphyritic granite melt and (ii) both granite types yield
37 zircon U–Pb emplacement ages consistent within a 575.3 ± 5.6 to 581.3 ± 3.8 Ma time
38 frame. The Lake Iro granites thus represent the oldest of a series of post-collisional
39 igneous associations exposed in southern Chad and nearby countries which,
40 collectively, show that the amalgamation of the constituent blocks of the southern
41 Saharan Metacraton is older than 580 Ma.

42

43 **Keywords:** A-type granites, post-collisional magmatism, Saharan Metacraton, Pan-
44 African orogeny, southern Chad

45

46 **Introduction**

47 The geodynamic framework of Central Africa is not completely understood. In this area,
48 the continental crust was assembled in the late Neoproterozoic (0.68 to 0.58 Ga) and

49 built up by two main Neoproterozoic to Paleoproterozoic cratonic blocks (the West African
50 Craton and the Congo Craton) ligatured by the so-called Pan-African mobile belts (Fig.
51 1a). These include the Trans Saharan Orogen, the East African Orogen and the
52 Oubanguides (Abdelsalam et al., 2002; Black and Liégeois, 1993; Ennih and Liégeois,
53 2008; Garfunkel, 2015; Kennedy, 1964; Kröner, 1980; Liégeois et al., 2013; Stern,
54 1994). These mobile belts are intensively deformed crustal segments resulting from
55 the amalgamation of reworked microcratonic domains and juvenile crust (extracted
56 from the mantle not long before its incorporation into the belts). The juvenile crust
57 formed between 0.8 and 0.65 Ga, during the closure of several oceanic domains, and
58 encompasses a set of (meta)igneous and (meta)sedimentary rocks which collectively
59 represent remnants of active margins (Bouyo et al., 2015; Caby, 2003; Liégeois et al.,
60 2003; Triantafyllou et al., 2016). The intrinsically composite origin of the mobile belts
61 is best reflected by the heterogeneous radiogenic isotope signature of their late
62 Neoproterozoic igneous rocks which testifies to the tapping of two contrasting
63 geochemical reservoirs: an old basement (Neoproterozoic to Paleoproterozoic cratonic
64 fragments) and Neoproterozoic arc rocks (Errami et al., 2009; Liégeois et al., 2003,
65 1998; Toteu et al., 2001). The Arabian Nubian Shield within the East African Orogen
66 (Fig. 1a) stands as a notable exception as it lacks any large remnant of pre-
67 Neoproterozoic crust (Stern, 2002). These mobile belts also contain significant
68 volumes of post-collisional igneous rocks and have been dissected by continent-scale
69 (100s of km long) strike-slip shear zones which can be traced in the adjacent South
70 American continent (Ngako et al., 2003; Njonfang et al., 2008; Toteu et al., 2004).
71 Among this geodynamic puzzle lies the Saharan metacraton (Abdelsalam et al., 2002)
72 which is also termed the Central Saharan ghost craton (Black and Liégeois, 1993).
73 This is a poorly known crustal domain which would encompass several Pan-African

74 mobile belts suturing cratonic blocks. The latter have been imaged using geophysical
75 methods (Murzuq, Al-Kufrah and Chad cratons, Fezaa et al., 2010; Liégeois et al.,
76 2013) but detailed petrographic data on their constituent lithologies are restricted to
77 one report (Bea et al., 2011). The S-SE region of Chad is located in a critical domain
78 at the southern margin of the Saharan Metacraton and within the so-called
79 Oubanguides or Central African Orogenic Belt (Fig. 1) which connects it to the Congo
80 Craton (Abdelsalam et al., 2002; Pin and Poidevin, 1987; Toteu et al., 2004). However,
81 the geology of the S-SE region of Chad is relatively unknown due to its geographical
82 isolation and the large extension of laterite and recent sediment cover.

83 Field trips were conducted during the GELT Program on Lake Iro (10°10'N - 19°42'E)
84 located in the southern part of Lake Chad Basin, about 100 km north of the border with
85 the Central African Republic (Fig. 1b). The goals of the GELT project included
86 description of the geological environment of the lake, assessment of its putative impact
87 crater origin (Reimold and Koeberl, 2014), characterization of its sedimentary filling
88 and determination of the hydrological and hydrogeological features of the area (Poulin
89 et al., 2019). The present study focused on exploration of the lake surroundings and
90 presents a description of newly discovered crystalline inliers where granitoids are
91 exposed. Petrographic and geochemical investigations, as well as U-Pb zircon dating
92 were conducted on selected samples. We also recorded basic magnetic properties of
93 the constituent lithologies for application to aeromagnetic anomaly interpretation. This
94 new set of data, coupled with the examination of geophysical regional anomalies,
95 offers the opportunity to correlate the newly described igneous suite with geological
96 formations located in neighboring sectors of the Oubanguides and position the Lake
97 Iro region within the geodynamic framework of Central Africa.

98

99 **1. Geological setting**

100 Four main lithotectonic domains of the Oubanguides Pan-African mobile belt are
101 exposed in southern Chad and nearby countries (Fig. 1b): the Ouaddaï massif to the
102 North-East, the Guéra Massif to the North, the West Cameroon Domain and its
103 Chadian continuation (termed the Mayo Kebbi massif and hereafter referred to as MK–
104 WCD) and the Adamawa-Yadé domain (AYD) to the South. The AYD, mainly exposed
105 in Cameroon and the Central African Republic (CAR), is composed of Neoproterozoic
106 metavolcanosediments (Toteu et al., 2006) and features remnants of Archean to
107 Paleoproterozoic crust metamorphosed during Neoproterozoic times (Ganwa et al.,
108 2016; Penaye et al., 1989; Toteu et al., 2001). Abundant syn- to post-tectonic (mostly
109 Ediacaran) granitoid plutons have been described in the AYD (Bessoles and
110 Trompette, 1980; Soba et al., 1991; Tchameni et al., 2006; Toteu et al., 2004). Finally,
111 the Central Cameroon Shear Zone (Fig. 1b) corresponds to a major late Pan-African
112 structure dissecting the AYD (Njonfang et al., 2008).

113 The MK–WCD is regarded as a predominantly juvenile Neoproterozoic crust segment
114 formed in an active margin setting (Bouyo et al., 2015; Penaye et al., 2006; Pouclet et
115 al., 2006). It includes: (i) (meta)sediments deposited in the Cryogenian–Ediacaran and
116 metamorphosed shortly afterwards (Bouyo et al., 2009; Toteu et al., 2006); (ii) very
117 scarce fragments of reworked Paleoproterozoic crust (Bouyo et al., 2009); and (iii)
118 several magmatic calc-alkaline plutonic suites, the earliest consisting of mafic to
119 intermediate granitoids emplaced between 737 and 723 Ma while the second consists
120 of more silicic terms and is dated at 665–640 Ma (Bouyo et al., 2016; Penaye et al.,
121 2006). The MK–WCD was involved in a polyphase tectonic collage during the Pan-
122 African orogeny. First the Mayo-Kebbi arc was juxtaposed against the AYD at c. 650-
123 640 Ma, a contact corresponding to the sinistral Tcholliré-Banyo Shear Zone (Nomo et

124 al., 2017; Pinna et al., 1994) (Fig. 1b). Then, this newly accreted block collided with a
125 more western segment, the Neoproterozoic magmatic arc of Poli. As noted for the
126 AYD, the MK–WCD rocks were intruded by Ediacaran c. 570 Ma-old post-collisional
127 granites (Isseini et al., 2012; Pouclet et al., 2006).

128 Little is known about the Guéra massif which is primarily composed of granitoids (with
129 minor gabbroic bodies) and metamorphic rocks (Isseini et al., 2013; Kusnir and
130 Moutaye, 1997). A recent U–Pb dating survey of the igneous rocks revealed
131 emplacement ages ranging between 595 and 545 Ma (Shellnutt et al., 2020, 2019,
132 2018, 2017) while a Neoproterozoic age is inferred for the metamorphic lithologies.
133 Similarly, limited data are available on the southern Ouaddaï massif (Fig. 1b) which
134 encompasses a range of c. 635–540 Ma-old granitoids (Djerossem et al., 2020; Kusnir
135 and Moutaye, 1997; Liégeois, 1993) along with metamorphic rocks including
136 amphibolites and Neoproterozoic metasediments which would have experienced
137 metamorphism at 620–600 Ma (Djerossem et al., 2020). Results of detrital zircon
138 geochronology performed on alluvium from the Chari river, which drains a large
139 catchment covering most of SE Chad, yielded a date distribution clustered in the
140 interval 550 – 650 Ma. This further indicates that the bulk of the local crust has been
141 strongly affected by an Ediacaran magmatic episode (Shellnutt et al., 2019).

142 The post-Pan-African cover of the study area includes Cretaceous to Quaternary
143 terrigenous sedimentary formations (Genik, 1993; Guiraud et al., 2005) which
144 constitute notably thick, i.e. pluri-kilometric, sequences in four extensional to
145 transtensional basins (Bongor, Doba, Doseo and Salamat, Fig. 1b). Sedimentation in
146 these basins was partly controlled by normal to dextral brittle deformation along crustal
147 structures parallel to the Pan-African shear zones thus suggesting a strong structural

148 inheritance (Genik, 1993). Cretaceous to Pliocene protracted alkaline volcanic activity
149 has also been recorded in the area (e.g. Nkouandou et al., 2008)

150

151 **2. Materials and methods**

152 **2.1 Bedrock exposures around the Lake Iro**

153 Most of the crystalline basement around the Lake Iro is hidden by an extensive lateritic
154 cover (locally as thick as 20 m) and by recent sediments. Five bedrock occurrences
155 were newly identified in the area and investigated during this study (Fig. 2, 3).

156 Two occurrences were discovered respectively 2 km S and 5 km SW of Lake Iro, close
157 to the Massadjanga village (Fig. 2). Site G₁ (*10°01'36.84N/19°25'44.173E*) (Fig. 3a) is
158 a massive outcrop forming two roughly 300 m-long and N60-trending ridges, together
159 with smaller exposures being found southwards. Site G₂ (*10°00'59.02N/19°23'04.39E*)
160 corresponds to a 50 m-round subcircular massif surrounded by large boulders. The
161 Massadjanga outcrops are built-up by a medium-grained locally K-feldspar-porphyritic
162 granite (samples A2, A3, A4, A5, Iro08) cut across by 15- to 20-cm wide aplite veins
163 (sample Iro17, Fig. 3c) and hosting various enclaves (Fig. 3b and d). These encompass
164 30- to 40 cm large, fine- (Iro15) to medium-grained (Iro18), light-coloured and quartz-
165 bearing varieties along with smaller, darker, fine-grained mafic types (Iro09) akin to
166 mafic microgranular enclave (MME) (Didier and Barbarin, 1991).

167 Two outcrops are located c. 10.5 km NW of the lake, near the Karou village. Site G₃
168 (*10°12'50N/19°18'40E*) (Fig. 3e) corresponds to a rocky tip built-up by more-or-less
169 loose blocks lacking clear orientation. Site G₄ (*10°12'52N/19°19'10E*) (Fig. 3f) is a 300
170 m-round subcircular massive outcrop showing N110 to N140-trending joints. Both
171 outcrops feature as main rock type an unsheared biotite microgranite (samples A1, A6,

172 A7, A8). The microgranite hosts rounded fine-grained melanocratic enclaves (Iro04)
173 (Fig. 3g) and are locally cut across by pegmatite veins.

174 Site G₅ (09°58.623'N/19°04.315'E) (Fig. 3h) is located SW of Lake Iro, near the village
175 of Malé. It features dome-shaped outcrops capped by large boulders composed of a
176 medium-grained porphyritic granite (sample A9) locally cut across by aplite (Fig. 3i)
177 veinlets. Two groups of joints strike N30 to 40 and N120, respectively.

178 **2.2 Mineral chemistry**

179 Mineral major element compositions were measured at the Laboratoire Magmas &
180 Volcans (Clermont-Ferrand, France) using a Cameca SX100 electron microprobe with
181 an accelerating voltage of 15 kV, a beam current of 15 nA, a beam size of 1 µm, and
182 20s counting times for all elements. The standards used were natural and synthetic
183 minerals. Biotite and amphibole structural formulas (including ferric contents) were
184 calculated using the approach and spreadsheets of Li et al. (2020a, 2020b) and the
185 grains named following the classification schemes of Rieder et al. (1999) and
186 Hawthorne et al. (2012), respectively. Feldspar formula were calculated using a
187 Gabbrosoft spreadsheet. Representative mineral analyses are available in
188 Supplementary Table 1.

189

190 **2.3 Whole-rock geochemistry**

191 Whole-rock major and trace element compositions were analysed at the Service
192 d'Analyse des Roches et Minéraux (SARM, CNRS, Nancy) from powdered samples.
193 Major element compositions were measured using an emission spectrometer ICP-OES
194 (ICap 6500 Thermo Fischer), and trace elements by ICP-MS (Thermo Elemental X7)
195 following the method detailed by Carignan et al., (2001). Typical analytical precisions
196 were c. 2% for major elements and 5-8% for trace elements.

197

198 **2.4 Zircon U-Pb dating**

199 Zircon grains were separated from crushed rock samples at Saint-Etienne University
200 using standard techniques (magnetic separator, heavy liquids), cast into epoxy mounts
201 and polished down to equatorial sections. U–Th–Pb isotope data were measured by
202 laser ablation inductively coupled mass spectrometry at the Laboratoire Magmas &
203 Volcans (Clermont-Ferrand, France). Zircons were ablated using a Resonetics
204 Resolution M-50 equipped with a 193 nm Excimer laser system coupled to a Thermo
205 Element XR Sector Field ICP-MS. Helium carrier gas was supplemented with N₂ prior
206 to mixing with Ar for sensitivity enhancement (Paquette et al., 2014). The laser was
207 operated with a spot diameter of 27 µm, a repetition rate of 3 Hz, and a laser fluence
208 of 2.5 J/cm². The alignment of the instrument and mass calibration were performed
209 before every analytical session using a NIST SRM 612 reference glass, by inspecting
210 the signals of ²³⁸U, ²³²Th and ²⁰⁸Pb and by minimising the ThO⁺/Th⁺ ratio. The analytical
211 method for isotope dating with laser ablation ICPMS is similar to that reported in Hurai
212 et al., (2010) and Mullen et al., (2018). The ²³⁵U signal is calculated from ²³⁸U on the
213 basis of the ratio ²³⁸U/²³⁵U = 137.818 (Hiess et al., 2012). Single analyses consisted of
214 30 seconds of background integration with the laser off followed by 60 seconds
215 integration with the laser firing and a 20 seconds delay to wash out the previous sample
216 and prepare the next analysis.

217 Data were corrected for U-Pb fractionation occurring during laser sampling and for
218 instrumental mass bias by standard bracketing with repeated measurements of GJ-1
219 zircon reference material (Jackson et al., 2004). Repeated analyses of 91500 zircon
220 reference material (Wiedenbeck et al., 1995) during each analytical session and
221 treated as unknown, independently controlled the reproducibility and accuracy of the

222 corrections. Data reduction was carried out with the software package GLITTER® (Van
223 Achterbergh et al., 2001). Concentrations of U, Th, and Pb were calculated by
224 normalization to the certified composition of GJ-1 (Jackson et al., 2004). Data were not
225 corrected for common Pb. Tera-Wasserburg $^{207}\text{Pb}/^{206}\text{Pb}$ vs. $^{238}\text{U}/^{206}\text{Pb}$ diagrams were
226 generated for each sample using Isoplot/Ex v. 2.49 software of Ludwig (2008). Error
227 ellipses for each point are shown at the 2σ level and incorporate both internal and
228 external uncertainties. Data points were pooled and Isoplot was used to calculate a
229 date and associated 2σ error for each sample.

230 **2.5 Magnetic properties**

231 Low-field susceptibility (K, dimensionless in SI) was measured on large sub-samples
232 (circa 24 to 72 g) using the KLY2 Kappabridge. Natural and saturation isothermal
233 remanences (NRM and IRM, in A/m) were measured using a 2G cryogenic
234 magnetometer on smaller subsamples (7 to 21 g). Measurements were normalized by
235 mass but transformed into volume normalized units using an arbitrary density of 2.7.
236 Using a present day ambient field H of 28 A/m, induced magnetization $M_i=K*H$ was
237 used to compute the total in situ magnetization ($M_i + \text{NRM}$) and Koenigsberger ratio
238 NRM/M_i , which are typical parameters for magnetic anomaly interpretation. To
239 evaluate the origin of NRM and possible bias by lightning, NRM/IRM ratios were
240 computed before and after demagnetization in a 30 mT alternating field (AF).

241 **3. Results**

242 **3.1 Petrography and mineral chemistry**

243 3.1.1 The Massadjanga-Malé suite

244 The Massadjanga and Malé granites share a similar mineralogy and both will be
245 described concomitantly in the following. They exhibit medium-grained (>1 mm)
246 porphyritic textures and the common igneous assemblage includes quartz, microcline

247 (Or₉₀₋₉₄, Fig. 5a), oligoclase (An₁₄₋₂₇, Fig. 5a), amphibole, biotite and a large amount of
248 accessories such as titanite, Fe-Ti oxides, zircon, apatite and epidote (Fig. 4a to c and
249 h to i). The samples collected near Massadjanga display local evidence for strong
250 deformation leading to protoclastic to mylonitic textures. Quartz appears as neo-
251 crystals within polycrystalline mylonitic bands (Fig. 4c) or as partially restored
252 phenoclasts. The K-feldspar occurs as phenocrysts and small laths in the matrix. The
253 former are perthitic and poikilitic with inclusions of quartz and plagioclase. The
254 plagioclase grains embedded in the K-feldspar are slightly more calcic than the crystals
255 of the matrix. Calcic amphibole is a common mineral in the granite and is notably Fe-
256 rich (3.5–4 a.p.f.u.) and ferroan (Mg-number lower than 25). This mineral displays
257 several habitus. A former green-brown amphibole (Ti-rich hastingsite) partially
258 recrystallized in blue-green amphibole (hastingsite). Part of the green-brown
259 amphibole could derive from the breakdown of earlier Ca-pyroxene as suggested by
260 the presence of amphibole-quartz symplectites. Finally, amphibole is partially replaced
261 by biotite and epidote. Fe-rich (1.9–2.1 a.p.f.u.) and ferroan (Mg# <25) biotite (Fig. 4a)
262 classified as annite (Fig. 5b) develop as large crystals that contain numerous zircons.
263 Igneous compositions are commonly preserved but a third of the analysed grains plot
264 in the field of reequilibrated biotite (Fig. 5c) in the diagram of Nachit et al. (2005).
265 Feldspars, amphibole and biotite locally show undulatory extinction and/or fracturing.
266 Apatite, oxides and zircons appear in clusters associated with ferromagnesian
267 minerals. They are usually euhedral and well prismatic. Fe-Ti oxides are partially
268 replaced by titanite.

269 The mafic microgranular enclave (MME Iro09, Fig. 3b) has sliced contacts with the
270 host granite. It shows a sub-doleritic texture well marked by plagioclase laths.
271 Oligoclase, amphibole, and biotite are the dominant minerals while quartz and K-

272 feldspar are subordinate. Post-magmatic deformation is underlined by undulatory
273 extinction of crystals. Epidote, apatite, Fe-Ti oxides together with secondary titanite are
274 abundant. However the thin section reveals a particular, several mm-thick zone at the
275 granite–MME contact. This zone is enriched in amphibole, biotite, needle-shape
276 apatite and large prisms of zircon. It could result from thermal–chemical exchanges
277 between the MME and its host, related to magma mixing processes.

278 The aplite dyke (Iro17, Fig. 3c) and the leucocratic enclave (Iro15, Fig. 3d) both display
279 sharp contacts with the host granite. They are similar to the main granite facies,
280 regarding mineral assemblage and deformation features, but typified by a finer grain
281 size of less than one mm.

282

283 3.1.2 The Karou suite

284 The Karou microgranite (A1, A6, A7, A8) is characterized by a microgranular texture
285 with hypovolcanic features such as globular embayed (rhyolitic) quartz crystals and,
286 less commonly, a slight “fluidal” fabric underlined by oriented euhedral laths of sodic
287 plagioclase. No significant post-magmatic deformation features were observed (Fig.
288 4e and f). The mineral assemblage is consistent with a high degree of differentiation:
289 quartz, microcline (Or_{92-97} , Fig. 5a) and albite (An_{1-18}) are abundant and occur both as
290 phenocrysts and as major components of the mesostasis (Fig. 4e,f). Microcline
291 phenocrysts often include globular quartz inclusions and plagioclase is mainly
292 restricted to the groundmass. The only Fe-Mg mineral is Fe-rich (3.8–4.7 a.p.f.u.) and
293 highly ferroan ($Mg\# < 10$) magmatic biotite (annite, Fig. 5b,cd). Epidote and titanite are
294 present but always in lesser amounts. Zircon is embedded in biotite or occurs as
295 anhedral grains in the groundmass. Sample A8 notably contains fluorite, which
296 displays an interstitial habitus. Sample A6 is slightly coarser-grained than the typical

297 facies (Fig. 4e). The melanocratic enclave Iro04 (Fig. 3g) is composite with a
298 microgranular inner part surrounded by a narrow 3 cm-thick rim, with a coarser sub-
299 millimeter grain, at the contact with the host granite. The mineralogy is mainly
300 composed of quartz, dominant plagioclase (albite-oligoclase), K-feldspar and
301 abundant biotite as large crystals.

302

303 **3.2 Whole-rock geochemistry**

304 The limited influence of post-magmatic alteration and weathering on whole-rock
305 chemical signatures is evidenced by: (i) the relatively fresh aspect of the rocks at the
306 hand-sample and thin-section scale; (ii) the weathering indexes W , calculated following
307 the method of Ohta and Arai (2007), are always lower than 30% in a FMW diagram
308 (see Couzinié et al., 2017); (iii) the positive correlations between fluid-mobile (Ba, Sr)
309 and fluid-immobile (Zr, Hf) incompatible trace elements. Therefore, we will consider in
310 the following that the range of observed chemical compositions exclusively results from
311 igneous processes.

312 **3.2.1 The Massadjanga-Malé suite**

313 This suite shows a large compositional variation with SiO_2 contents mostly ranging
314 between 68.4 and 77.0 wt, the MME Iro09 being less silicic (SiO_2 : 61.9 wt.%). All
315 samples plot in the field of granite and notably define a “subalkaline” trend in the P–Q
316 cationic diagram (Fig. 6a) of Debon and Le Fort (1983). In the Q’–ANOR normative
317 classification diagram of Streckeisen and Le Maitre (1979), all but two samples plot in
318 the field of syenogranites (Fig. 6b). The Massadjanga–Malé granites are metaluminous
319 to subaluminous (A/CNK : 0.90–1.04; A/NK : 1.08–1.36; Fig. 6c), alkali-calcic to alkalic
320 (in the sense of Frost et al., 2001, Fig. 6d) and define a potassic and ferriferous
321 association based on the cationic $\text{Fe}+\text{Mg}+\text{Ti}$ vs. $\text{K}/(\text{Na}+\text{K})$ and $\text{Mg}/(\text{Mg}+\text{Fe})$ diagrams

322 of Debon and Le Fort (1988) (Fig. 6e,f). Their agpaitic indexes ($\text{Na}+\text{K}/\text{Al}$) are in the
323 range 0.74–0.86 apart for the aplite sample Iro 17 and the granite sample collected at
324 Malé which was 0.93 (Supplementary Table 2).

325 The LREE are moderately fractionated with $(\text{La}/\text{Sm})_{\text{N-chondrite}}$ ratios in the range 2.5–
326 6.6 (Fig. 7). HREE patterns are nearly flat and Eu negative anomalies weak to marked
327 (Eu/Eu^* : 0.31–1.21). HFSE and REE contents are elevated (Zr: 210–717 ppm; Nb: 17–
328 45 ppm; Ce: 105–403 ppm; Sum_{REE} : 249–884 ppm). All samples show Ti, Nb and Sr
329 negative anomalies but only two granites display a Ba negative anomaly (Fig. 7). Three
330 notable exceptions were pinpointed within the Massadjanga dataset: (i) the aplite
331 sample (Iro 17) shows a strong Eu negative anomaly (Eu/Eu^* : 0.24), a flat LREE and
332 negative HREE patterns with a $(\text{Gd}/\text{Yb})_{\text{N}}$ as low as 0.5 and a lower Zr content (96 ppm),
333 consistent with fractionation of feldspar and accessory minerals including zircon; (ii)
334 two leuco- (B values <38) granite samples (A5 and enclave Iro18) display strongly
335 positive Eu anomalies ($\text{Eu}/\text{Eu}^* > 2.5$, Fig. 9a), a pronounced depletion in HREE (Yb
336 contents below 1.4 ppm) and low HFSE and REE contents (Nb: 3–4 ppm; Zr: 59–68
337 ppm; Sum_{REE} : 165–168 ppm). These two leucogranites have markedly higher Ba and
338 Sr compared to other fractionated specimens from the suite (A4, A9, Fig. 9a,b). Such
339 signatures collectively indicate that they encompass a significant component of
340 accumulated feldspar and deviate from melt compositions.

341 3.2.2 The Karou suite

342 Microgranite samples from the Karou suite show a narrower range of high SiO_2
343 contents (74.2–75.0 wt.%) and plot in the field of granites in the P–Q cationic diagram
344 (Fig. 6a) of Debon and Le Fort (1983) and in the field of syenogranites in the Q' –
345 ANOR classification diagram of Streckeisen and Le Maitre (1979) (Fig. 6b). They are
346 subaluminous (A/CNK : 0.99–1.02; A/NK : 1.15–1.16, Fig. 6c), calc-alkalic to alkali-

347 calcic (in the sense of Frost et al., (2001), Fig. 6d) and also define a potassic and
348 ferriferous association (Fig 6e,f) (Debon and Le Fort, 1988). The enclave Iro04 is
349 slightly less silicic (SiO_2 : 69.5 wt.%), metaluminous (A/CNK : 0.93; A/NK : 1.22) but still
350 exhibits the same ferroan and potassic character as its host (Fig. 6e,f).
351 As for major elements, trace element compositions are clustered (Fig. 7). LREE
352 patterns are moderately fractionated with $(\text{La}/\text{Sm})_{\text{N-chondrite}}$ ratios in the range 4.4–5.2.
353 HREE patterns are nearly flat with $(\text{Gd}/\text{Yb})_{\text{N-chondrite}}$ ranging between 1.1 and 1.36. All
354 samples show a negative Eu anomaly (Eu/Eu^* : 0.25–0.29) and exhibit elevated
355 concentrations in HFSE (Zr: 203–249 ppm; Nb: 26–32 ppm) and REE (Ce: 211–242
356 ppm; Sum_{REE} : 473–533 ppm). The Karou microgranites display Sr, Nb, Ti and Ba
357 negative anomalies (Fig. 7). Enclave Iro04 shows overall similar patterns but differs by
358 its enrichment in Zr, Hf, Sr, Ba, Eu and depletion in U and Th compared to the host
359 (Fig. 7, 9, 10).

360

361 **3.3 Zircon U-Pb geochronology**

362 Two samples (one from each igneous suite) were selected for zircon U–Pb
363 geochronological investigations (Supplementary Table 3). The zircon crystals from the
364 A3 Massadjanga granite sample are large, euhedral and often broken. They are poorly
365 translucent, yellowish and filled with numerous undetermined dark inclusions. U and
366 Th concentrations are generally lower than 100 ppm. High Th/U ratios are consistent
367 with the magmatic origin of the zircons. Thirteen analyses were performed on the 7
368 available zircon grains defining a Discordia line with a lower intercept date of $575.3 \pm$
369 5.6 Ma (Fig. 8) interpreted as the crystallisation age of the zircon crystals.

370 In the A6 Karou granite sample, the zircon crystals are smaller but still euhedral,
371 translucent and colorless. They contain rare clear undetermined inclusions. Th and U

372 content are often lower than 50 ppm whereas Th/U ratios are high and up to 1.5.
373 Nineteen analyses were performed on 13 zircon grains and yielded a Concordia date
374 of 581.3 ± 3.8 Ma (Fig. 8) interpreted as the crystallization age of the zircon grains.

375

376 **3.4 Magnetic properties**

377 Table 1 summarizes the results of measurements made on the A1 to A9 samples. The
378 three granitic suites Karou, Massandjana and Malé appear distinct regarding the
379 intensity of their ferromagnetic signal. Based on susceptibility (Gleizes et al., 1993;
380 Rochette, 1987) the Karou samples appear mostly paramagnetic ($K < 0.6$ mSI, apart
381 from A6), while Massandjana is mildly ferromagnetic (K from 1 to 2.3 mSI) except A5
382 which is paramagnetic, in agreement with its very low iron content (see Supplementary
383 Table 2). Accordingly the NRM and IRM intensities increase from Karou to Malé. To
384 propose a preliminary database for Chad granites aeromagnetic interpretation (such
385 data do not currently exist) we first note that the Q ratio is always $\gg 1$ except in the
386 two weakest Karou samples (A1, A7), thus justifying the need to consider remanence
387 in anomaly interpretation. Total magnetization varies between 0.1 and 4.6 A/m
388 (maximum for Malé). Such values are similar (although in the lower range) to values
389 found in granitoids from central CAR and Western Cameroun (Ouabego et al., 2013).
390 To confirm that the measured NRMs are not biased by lightning-induced remanence
391 and thus reliable as a proxy for in-situ crustal magnetization, we computed the
392 NRM/IRM ratio, that should be a few percent for non-lightning natural processes. Four
393 samples (A3, 4, 8, 9) yield a ratio in the 12-16% range, and appear to be contaminated
394 by lightning (Verrier and Rochette, 2002). After a 30 mT AF demagnetization these
395 ratios decreased below 4% except for A3. Therefore we propose (as done in Ouabego
396 et al., 2013) to use a modeled NRM by assuming that in situ $NRM = 0.02 * IRM$. This

397 leads to lower total magnetization, at most 0.7 A/m (for the Malé granite). Still, Q ratio
398 is $\gg 1$ (except for sample A7), consistent with data from granites in W Cameroon and
399 CAR (Ouabego et al., 2013).

400 **4. Discussion**

401 **4.1 Petrogenetic relationships between the Lake Iro granites**

402 Several lines of evidence support a common origin for the granites exposed North
403 (Karou) and South (Massadjanga-Malé) of Lake Iro. First, both igneous suites exhibit
404 a ferroan mineralogy (including Fe-rich biotite) and define similar trends in geochemical
405 diagrams (see section 4.3). One of the most silicic sample from the Massadjanga-Malé
406 suite (A4) overlaps, in terms of major and trace element contents, with the Karou
407 microgranites. The Karou enclave Iro04 is also chemically identical to the main
408 Massadjanga-Malé granite type (Fig. 7, 8, 9). Second, trace element modelling
409 indicates that the most fractionated samples (the Karou microgranites and sample A4)
410 can be generated by c. 30% fractional crystallization of an assemblage of amphibole
411 + plagioclase + K-feldspar + ilmenite starting from an average Massadjanga-Malé
412 porphyritic granite composition (Fig. 9b). The fractionated granites exhibit lower zircon
413 saturation temperatures (762–820°C depending on the calibration considered) than
414 the porphyritic types (815–884°C). In the absence of inherited zircon cores among
415 magmatic zircon grains, those values are regarded as minimum magma temperatures.
416 These indicate that zircon was part of the fractionating assemblage and that the Karou
417 microgranites, along with Massadjanga sample A4, represent lower temperature
418 residual melts related to the main Massadjanga-Malé porphyritic granite type. Third, in
419 addition to their geochemical kinship, the Karou and Massadjanga granites yielded
420 zircon U–Pb emplacement ages which are consistent, within analytical uncertainties,

421 at c. 580–575 Ma. We therefore suggest that the Lake Iro granites were part of a single
422 igneous system active at the end of the Precambrian.

423 Petrographic evidence strongly suggests that, despite their petrogenetic relationships,
424 the Karou and Massadjanga granites were emplaced at different crustal levels. The
425 presence of “rhyolitic” globular quartz and the fine-grained microgranular texture of the
426 Karou samples collectively demonstrate that the Karou magmas emplaced at very
427 shallow crustal levels, which is consistent with their fractionated character. In contrast,
428 the medium-grained texture of the Massadjanga-Malé granites is indicative of slow
429 cooling in the middle crust. The aluminum content of calcic amphiboles is generally
430 considered to be a reliable indicator of the pressure at which granite magmas
431 crystallize (review in Anderson et al., 2008), but a calibration designed for ferroan calcic
432 amphibole compositions is unavailable. Application of the Mutch et al. (2016)
433 formulation to the hastingsite grains of the Massadjanga granites suggest
434 emplacement pressures of 3.4 to 5.0 kbar, i.e. 13 to 19 km (considering an average
435 crustal density of 2700 kg.m^{-3}).

436 In addition to their contrasting emplacement depths, the Karou and Massadjanga
437 granites also differ with respect to their subsolidus tectonic evolution. The Karou
438 samples display no evidence of solid-state deformation, in marked contrast to the
439 Massadjanga (but not the Malé) granites. Petrographic features including recrystallized
440 quartz and fractured feldspar/amphibole show that the latter experienced strong
441 shearing at low temperature. The linear shape of the largest Massadjanga outcrop (G₁,
442 section 2.1) may be closely related to this deformation event. The N60 trend of the two
443 massive ridges that constitute the outcrop G₁ matches that of the main shear zones in
444 the Oubanguides, notably the Central Cameroon Shear Zone (Toteu et al., 2004). It is
445 most likely that the Lake Iro crustal segment experienced heterogeneous ductile

446 shearing subsequent to the cooling of the late Ediacaran granites which had previously
447 been emplaced at different crustal levels. The age of this deformation remains
448 undetermined.

449

450 **4.2 Typology of the Lake Iro igneous association**

451 The Lake Iro granites are typified by their ferroan mineralogy and their major element
452 composition matches that of A-type granitic suites (Bonin, 2007; Bonin et al., 2020;
453 Collins et al., 1982; Eby, 1990; Frost and Frost, 2011) which feature: (i) high alkali and
454 low CaO contents (the rocks are alkali-calcic to alkalic, Fig. 6d); (ii) low Mg# (Fig. 6f);
455 (iii) moderate aluminosity (rocks are metaluminous to weakly peraluminous, Fig. 6c);
456 (iv) a potassic character (Fig. 6e). The analogy is also supported by the trace element
457 systematics and notably the elevated HFSE (Zr, Nb, Y) and REE concentrations.
458 Consistently, the Lake Iro granites (discarding cumulate compositions) plot in the field
459 of A-type granites in the discrimination diagrams (Fig. 10a,b) of Whalen et al. (1987)
460 and in the field of Within Plate Granites (which encompasses A-type varieties) in the
461 Nb vs. Y diagram (Fig. 10c) of Pearce et al. (1984). The affinity with A-type granite
462 suites is further substantiated by the discrimination diagram (Fig. 10f) of Abdel-
463 Rahman (1994) which is based on biotite compositions, regarded as repositories of
464 magma chemistry. Similarly, the elevated Fe³⁺ contents of magmatic biotite (Fig. 10g)
465 plot above the NNO buffer indicating that the magma were relatively oxidized, as
466 commonly observed in A-type suites (Shabani et al., 2003). The presence of magmatic
467 fluorite is also a typical feature of A-type granites (Collins et al., 1982). Finally, the Th–
468 U contents of magmatic zircon grains from the Lake Iro suite are also consistent with
469 those of alkalic magmas (Paquette et al., 2019).

470 Following the subdivision scheme of Eby (1992) and Grebennikov (2014), the Lake Iro
471 samples are best classified as A₂-type granites (Fig. 10d,e). This group of igneous
472 rocks shares chemical characteristics with a suite of metaluminous to weakly
473 peraluminous alkali-calcic granitoids variously defined in the literature as “monzonitic”,
474 “subalkaline” or “transalkaline” associations (de La Roche, 1980; Lameyre and Bonin,
475 1991; Stussi, 1989) and corresponding to the “K-rich calc-alkaline granitoids” (KCG) of
476 Barbarin (1999). These associations encompass ferriferous varieties (sensu Debon
477 and Le Fort, 1988) which are relatively similar to the A₂-type granites (Bonin et al.,
478 2020) and have been extensively studied in the Variscan belt. Iconic examples include
479 the Ploumanac’h complex (Barrière, 1977) or the Mont-Blanc granites (Debon and
480 Lemmet, 1999). The kinship between A₂-type granites and the Fe–K “subalkaline” suite
481 is well-illustrated in granite discrimination diagrams originally designed based on
482 Variscan case studies, notably those based on biotite compositions (Nachit et al., 1985;
483 Rossi and Chèvremont, 1987). For example, in the FeO_t–MgO–Al₂O₃ ternary and Mg
484 vs. Al^{tot} binary discriminant diagrams (Fig. 10g,h), the Lake Iro samples dominantly plot
485 in the field of “subalkaline” granites (or close to the subalkaline–alkaline boundary).

486

487 Several petrogenetic models have been proposed to account for the genesis of A-type
488 granite suites. These include: (i) melting of mafic (D’Souza et al., 2006; Frost et al.,
489 1999; Tagne-Kamga, 2003) to quartz–feldspar-rich (Breiter, 2012; Dall’Agnol and de
490 Oliveira, 2007; Patiño Douce, 1997), and possibly residual (Collins et al., 1982;
491 Creaser et al., 1991; Landenberger and Collins, 1996; Skjerlie and Johnston, 1992)
492 crustal rocks; (ii) fractionation of mantle-derived magmas (Bogaerts et al., 2006;
493 Duchesne et al., 2010; Eby, 1990; Ferré et al., 1998; Frost and Frost, 2011; Liégeois
494 et al., 1998; Skridlaite et al., 2003; Turner et al., 1992). For the so-called “Fe-K

495 subalkaline” associations, models involving melting of the lithospheric to
496 asthenospheric mantle and subsequent fractionation of basic melts are preferred
497 (Debon and Lemmet, 1999; Laurent et al., 2014; Schaltegger and Corfu, 1992).
498 Discriminating between the different models for the Lake Iro granites is beyond the
499 scope of this study.

500

501 **4.3 Regional correlations and tectonic setting**

502 Addressing the geodynamic significance of the Lake Iro association and its bearings
503 on the evolution of the southern margin of the Saharan Metacraton would require a
504 regional structural sketch of the Central African Orogen in southern Chad. Such a
505 sketch is currently unavailable. In this poorly exposed area, geophysical data provide
506 an opportunity to delineate crustal blocks and attempt regional correlations. In Central
507 Chad, a major N45 lineament (the so-called Chad Lineament) has been identified by
508 Bayer and Lesquer (1978) based on a gravimetric Bouguer anomaly map and its
509 importance was reassessed by Braitenberg et al. (2011). Garfunkel (2015) interprets
510 this structure as a possible oceanic suture zone. A revisit of this map reveals an
511 overlooked, although prominent, N70 lineament in southern Chad (Fig. 11a; Bonvalot
512 et al., 2012) that approximately connects the Lake Iro area with the Doba basin and
513 the AYD massif to the SW and overlaps with the area of maximal Cretaceous–
514 Paleogene extension. The magnetic anomaly map (Fig. 11b; Dyment et al., 2015,
515 wdmam.org) is less resolved due to the lack of aeromagnetic surveys, but also shows
516 N70 trending dipolar anomalies in southern Chad. This direction matches that of the
517 major Bangui magnetic anomaly, which covers CAR just south of the present study
518 area (Ouabego et al., 2013). These authors suggested that a Pan-African iron-rich
519 crust may be the source of this anomaly. Therefore, we interpret the N70 lineament

520 observed in southern Chad as the prolongation of the Central Cameroon Shear Zone,
521 where the regional crust may possess reduced densities and be notably iron-rich. This
522 major structure would thus juxtapose a northern block (comprising the MK–WCD, the
523 northern part of the AYD, the Ouaddaï, Guéra massifs and Lake Iro inliers) against a
524 southern block encompassing most of the southern part of the AYD and associated
525 formations in CAR (see Fig. 2 of Toteu et al., 2004). Our new dataset on the Lake Iro
526 inliers provide novel constraints on the orogenic evolution of this northern block.

527

528 As demonstrated by Bonin (2007), A₂-type granites are the hallmarks of the post-
529 collisional stage of orogens (sensu Liégeois, 1998). The emplacement of the Lake Iro
530 granites consistently post-dates the main tectonic-metamorphic events that shaped the
531 northern block, which could be as old as 620–590 Ma (Bouyo et al., 2009; Djerosse
532 et al., 2020, Fig. 12). Igneous suites similar to that of the Lake Iro inliers are exposed
533 throughout this northern block (yellow stars, Fig. 11a). A-type volcanic-plutonic
534 associations with emplacement ages ranging between 568 ± 6 and 545 ± 6 Ma have
535 been described in the Guéra massif and two granitic inliers located 100 km to the NW
536 (Pham et al., 2017; Shellnutt et al., 2018, 2017). In the WCD-MK, the A₂-type Zabili
537 granite pluton was emplaced at 567 ± 10 Ma (Isseini et al., 2012) even though recent
538 geochronological investigations performed on a hydrothermally altered facies (albitite)
539 suggest it might be older (Vanderhaeghe et al., in press). No such rock types have
540 been described in the Ouaddaï massif but the latter exposes 538 ± 5 Ma-old
541 magnesian alkali-calcic potassic granites along with basic and intermediate igneous
542 rocks (including two pyroxene monzonites). Together, they define a KCG-type
543 (Barbarin, 1999) or “Mg–K subalkaline” (de La Roche, 1980) association (Djerosse
544 et al., 2020, blue stars, Fig. 11a). As for A₂-type granites, such igneous suites are

545 typically emplaced in post-collisional settings (Couzinié et al., 2016, 2014; Janoušek
546 et al., 2019; Laurent et al., 2017). Similar rocks have been described in the MK–WCD
547 (Pala suite, Penaye et al., 2006) and dated at 571 ± 1 and 567 ± 15 Ma. Besides, a
548 diorite sample retrieved from the Karin-1 well in the Doba basin also yielded an
549 intrusion age of 572.6 ± 0.3 Ma (Shellnutt et al., 2017). Collectively, the available
550 petrological and geochronological data support a protracted post-collisional magmatic
551 activity at the southern margin of the Saharan Metacraton, spanning the range 580–
552 540 Ma (Fig. 12). The Lake Iro granites, first described in this contribution, would thus
553 represent the oldest post-collisional intrusions in southern Chad.

554

555 **Conclusion**

556 The basement rocks exposed in the Lake Iro inliers correspond to cogenetic ferroan
557 alkali-calcic A₂-type granites emplaced at contrasting crustal levels. The different
558 components of this association are related by fractional crystallization processes.
559 Investigation of the granite magnetic properties indicates that they are variably
560 ferromagnetic, as observed in similar granites from West Cameroon and CAR, and
561 may thus generate moderate magnetic anomalies (total magnetization of the order of
562 1 A/m). Emplacement of the Lake Iro granites at c. 580–575 Ma, i.e. less than 20 Ma
563 after peak metamorphism related to arc accretion–collision (e.g. Bouyo et al., 2009).
564 This demonstrates a rapid change in magma sources and geodynamic setting at the
565 end of the Ediacaran in southern Chad. The onset of the post-collisional period was
566 marked by large lateral displacements along shear zones such as the CCSZ. Further
567 work is needed to identify the temporal evolution south of this structure, in CAR, and
568 to clarify the timeframe of post-collisional magmatism in the southern part of the
569 Saharan Metacraton

570 **Figure and Table captions**

571 Fig. 1: (a) Geological sketch delineating the main crustal domains of northern–central
572 Africa (age and Nd–Hf isotope signature). Inspired from Garfunkel (2015) and adapted
573 from Couzinié et al. (2019). (b) Geological map of southern Chad and adjacent
574 countries. The map was redrawn and adapted from the geological map of Africa at a
575 scale of 1/10 000 000 (Thieblemont, 2016). Dotted lines depict the outlines of the
576 Cretaceous–Tertiary rift basins in southern Chad (redrawn from Genik, 1993).
577 Abbreviations: C.A.R., Central African Republic; C., Cameroon; MK–WCD, Mayo
578 Kebbi–West Cameroon domain, AYD, Adamawa–Yadé domain, TBSZ, Tcholliré–
579 Banyo shear zone; CCSZ, Central Cameroon shear zone.

580

581 Fig. 2: Geological map centered on Lake Iro, depicting the location of the crystalline
582 inliers.

583

584 Fig. 3: Field pictures of the studied bedrock exposures. (a) outcrop of the Massadjanga
585 granite (G₁); (b) melanocratic enclave Iro09 in the Massadjanga granite; (c) aplite vein
586 Iro17 cutting across the Massadjanga granite; (d) leucocratic enclave Iro15; (e,f)
587 outcrops of the Karou microgranite (respectively G₃ and G₄); (g) melanocratic enclave
588 Iro04 in the Karou microgranite; (h) outcrop of the Malé granite (G₅); (i) aplite dyke
589 cutting across the Malé granite;

590

591 Fig. 4: Representative photomicrographs of studied samples. Massadjanga granite:
592 (a,b) variably deformed granites with two feldspars, Ca-amphibole, biotite, titanite and
593 apatite assemblage (plane-polarized); (c) strongly deformed, mylonitic facies with
594 quartz ribbons (cross-polarized); (d) euhedral zircon crystal in a weakly deformed

595 facies (cross-polarized). Karou granites: (e) isotropic and medium-grained type (cross-
596 polarized); (f) isotropic and fine-grained microgranite type (plane-polarized); (g)
597 anhedral zircon within the fine-grained facies (plane-polarized). Malé granite (cross-
598 polarized light): (h) weakly deformed two-feldspars, biotite, titanite and Fe-Ti oxides
599 granite; (i) euhedral titanite crystal hosted by biotite.

600

601 Fig. 5: Mineral chemical compositions. (a) Feldspar ternary classification diagram; (b)
602 binary classification diagram for micas after Tischendorf et al. (2007),
603 $feal = {}^{VI}Fe_{tot} + Mn + Ti - {}^{VI}Al$ and $mgli = Mg - Li$ a.p.f.u. ; (c) $FeO_t + MnO - 10TiO_2 - MgO$ ternary
604 diagram of Nachit et al. (2005) discriminating between magmatic and reequilibrated
605 biotites.

606

607 Fig. 6: Whole-rock geochemistry of the Lake Iro samples. (a) P–Q cationic
608 classification diagram of Debon and Le Fort (1983); (b) Q'–ANOR classification
609 diagram of Streckeisen and Le Maitre (1979); (c) B–A cationic classification diagram
610 of Debon and Le Fort (1983); (d) SiO_2 vs. $Na_2O + K_2O - CaO$ (MALI) diagram of Frost et
611 al. (2001); (e) B vs. $Mg/(Mg+Fe)$ diagram of Debon and Le Fort (1988) allowing
612 discrimination between ferriferous and magnesian associations; (e) $K/(K+Na)$ vs. B and
613 (f) B vs. $Mg/(Fe+Mg)$ cationic classification diagrams of Debon and Le Fort (1988). The
614 subdivisions of Villaseca et al. (1998) are indicated in plot (c). SALK refers to the
615 “potassic subalkaline” differentiation trend according to Debon and Le Fort (1988).

616

617 Fig. 7: Chondrite-normalized (Boynnton, 1984) Rare Earth Elements and N-MORB-
618 normalized trace elements patterns (Sun and McDonough, 1989) for the Karou

619 microgranites (left), Massadjanga-Malé granites (center) and peculiar facies
620 (leucogranites A5 et Iro18 and aplite Iro17) of the Massadjanga suite (right).

621

622 Fig. 8: Tera–Wasserburg diagrams ($^{238}\text{U}/^{206}\text{Pb}$ vs. $^{207}\text{Pb}/^{206}\text{Pb}$) showing
623 Neoproterozoic zircon data for the Lake Iro granites. Error ellipses/ages are quoted at
624 the 2σ level of uncertainty.

625

626 Fig. 9: Petrogenesis of the Lake Iro granitoids. (a) Eu/Eu^* and (b) maficity (B parameter
627 of Debon and Le Fort (1983) vs. Sr content for the Lake Iro granitoids. The red line
628 depicts the evolving composition of a sample A2 melt undergoing fractional
629 crystallization of an assemblage: 0.40 plagioclase An_{18} + 0.35 hastingsite + 0.23 K-
630 feldspar + 0.02 ilmenite (in mol.%). The considered partition coefficients for Sr are 8.0,
631 4.2 and 0.6 for plagioclase, K-feldspar and amphibole, respectively. They correspond
632 to the average values calculated from the results of Ewart and Griffin (1994). The blue
633 dotted lines connect the melt compositions to those of the fractionating assemblage.
634 (c) Zircon saturation temperatures calculated based on the calibrations of Watson and
635 Harrison (1983) and Boehnke et al. (2013). Same legend as Fig. 7.

636

637 Fig. 10: Granitoid discrimination diagrams based on whole-rock (a to e) and biotite (f
638 to h) chemical compositions (only grains plotting in the field of “magmatic biotite” in the
639 diagram of Nachit et al. (2005) were retained). (a,b): $\text{Zr}+\text{Nb}+\text{Ce}+\text{Y}$ (ppm) vs. FeO_i/MgO
640 and $(\text{Na}_2\text{O}+\text{K}_2\text{O})/\text{CaO}$ discriminant diagrams for A-type granites from Whalen et al.
641 (1987), OTG and FG stand for ordinary-type and fractionated granites, respectively;
642 (c) $\text{Rb}-(\text{Y}+\text{Nb})$ geotectonic diagram of Pearce (1996) separating syn-collision (syn-
643 COLG), post-collision (post-COLG), within-plate (WPG) and ocean ridge (ORG)

644 granites; (d,e) triangular plots of Eby (1992) and Grebennikov (2014) discriminating
645 between A₁ and A₂-type granites; (f) FeO_t–MgO–Al₂O₃ diagram after Abdel-Rahman
646 (1994) separating biotite crystallized in alkaline (A), calc-alkaline (C) and peraluminous
647 (P) granites; (g) Fe²⁺–Fe³⁺–Mg ternary diagram of Wones and Eugster, (1965)
648 indicative of the redox state of the melt from which biotite crystallized, with respect to
649 the three common oxygen fugacity buffers: fayalite–magnetite–quartz (FMQ), nickel–
650 nickel oxide (NNO) and hematite–magnetite (HM). (h) Mg vs Al^{tot} discriminant diagram
651 (atoms per formula unit) after Nachit et al. (1985) and Stussi and Cuney (1996); (i)
652 FeO_t–MgO–Al₂O₃ discriminant diagram after Rossi and Chèvremont (1987).

653

654 Fig. 11: Maps of the Bouguer gravimetric (a) and magnetic (b) anomalies in southern
655 Chad and adjacent countries. Source grids correspond to Bonvalot et al. (2012)'s
656 model for gravimetric anomaly, and to the WDMAM 2.0's grid for magnetic field
657 anomaly (Dyment et al., 2015, wdmam.org). Yellow and blue stars depict the location
658 of post-collisional igneous suites identified in the area: A-type granites and magnesian
659 K-rich calc-alkaline granitoids, respectively (see references in text).

660

661 Fig. 12: Timeline summarizing available chronological constrains on the orogenic
662 events in southern Chad and nearby countries based on the geological record of each
663 massif (from the MK–WCD to the West to the Ouaddaï massif to the East). Yellow and
664 blue stars depict the emplacement ages of typical post-collisional igneous
665 associations: A₂-type granites and magnesian KCG suites, respectively. See
666 references in text.

667

668 Table 1: Magnetic properties of Lake Iro granite samples. $\alpha(30)$ is initial NRM/IRM
669 (after 30 mT AF demagnetization).

670

671 Supplementary Table 1: Mineral major element compositions for the Lake Iro granitoid
672 samples. Feldspar structural formulas were calculated with a Gabbrosoft spreadsheet
673 on an 8 O basis. Amphibole formulas are in the form $A_{0-1}B_2C_5T_8O_{22}W_2$ and were
674 calculated using the method of Li et al. (2020a). Biotite has a general formula of
675 $A_1M_3T_4O_{10}W_2$ estimated following Li et al. (2020b).

676

677 Supplementary Table 2: Whole-rock geochemical data for the Lake Iro granitoid
678 samples in this study.

679

680 Supplementary Table 3: Zircon U-Pb data for the Lake Iro granites (SE Chad), obtained
681 by in situ Laser Ablation ICP-MS.

682

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