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## Discovery of a Devonian mafic magmatism on the western border of the Murzuq basin (Saharan metacraton): Paleomagnetic dating and geodynamical implications

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1 **Discovery of a Devonian mafic magmatism on the western border of the Murzuq basin**  
 2 **(Saharan metacraton): paleomagnetic dating and geodynamical implications.**

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30  
 31 **Abstract**

32  
 33 **Keywords:** Basic magmatism, Isotopic (K-Ar, U-Pb) dating, paleomagnetic dating,  
 34 Devonian, Murzuq craton, Saharan metacraton

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## Abstract

Intraplate deformation is most often linked to major stress applied on plate margins. When such intraplate events are accompanied by magmatism, the use of several dating methods integrated within a multidisciplinary approach can bring constraints on the age, nature and source mobilized for generating the magma and in turn on the nature of the intraplate deformation. This study focuses on the large gabbro Arrikine sill (35 km in extension) emplaced within the Silurian sediments of the western margin of the Murzuq cratonic basin in southeastern Algeria. Its emplacement is dated during the early Devonian (415-400 Ma) through the determination of a reliable paleomagnetic pole by comparison with the Gondwana Apparent Polar Wander Path (APWP). This age can be correlated with deep phreatic eruptions before Pragian time thought to be at the origin of sand injections and associated circular structures in Algeria and Libya. For the sill, the K-Ar age of  $325.6 \pm 7.7$  Ma is related to a K-rich aplitic phase that has K-enriched by more than 20% the Devonian gabbro. Laser-ICP-MS U-Pb method dates only inherited zircons mostly at c. 2030 Ma with additional ages at c. 2700 Ma and younger ones in the 766-598 Ma age range. The Arrikine sill is a high-Ti alkaline gabbro having the geochemical composition of a hawaiite akin to several intraplate continental and oceanic provinces, including the contemporaneous Aïr ring complexes province in Niger, but also to the Mauna Loa volcano in Hawaii. This peculiar composition akin to that of the contemporaneous Aïr province is in agreement with a lower Devonian age for the Arrikine sill.

The lower Devonian Arrikine sill emplacement is related to a "Caledonian" transtensive reactivation of the western metacratonic boundary of the Murzuq craton. This event also generated in the Saharan platform the so-called "Caledonian unconformity" of regional extension, the Aïr ring complexes and magmatic rocks that produced sand injections. It could be related to rifting of the Hun terranes that occurred at the plate margin to the north (Stampfli and Borel, 2002, Blackey, 2008 and references therein). The mid-Carboniferous (c. 326 Ma) reactivation corresponds to Variscan compression on NW Africa generating aplitic fluids, but also to the major "Hercynian unconformity" of regional extension. The generation of the Arrikine magma is attributed to partial melting through adiabatic pressure release of uprising asthenosphere along tectonically reactivated mega-shear zones, here bordering the relictual Murzuq craton enclosed in the Saharan metacraton.

## 68 **1. Introduction**

69 The Murzuq basin located in central North Africa, in Algeria, Libya and Niger (Fig. 1)  
70 is a key area within the Saharan metacraton, being located on a relictual cratonic nucleus, the  
71 Murzuq craton (Liégeois et al., 2013). It is one of the largest cratonic sedimentary basins of  
72 the Saharan platform, filled by different Phanerozoic series covered by a very large sandy  
73 desert, the Murzuq edeyen or erg. On its western border, we discovered a very large sill of  
74 mafic rocks interbedded in the Silurian sedimentary series belonging to the so-called Tassilis  
75 series. The sedimentary sequences unconformably rest on the late-Neoproterozoic Pan-  
76 African basement of the Tuareg Shield (Fig. 2). Known post-Pan-African magmatism in the  
77 In-Ezzane area is limited to Cenozoic lava flows ( $2.86 \pm 0.07$  Ma; Yahiaoui et al., 2014).  
78 Around the Murzuq basin, several large Cenozoic volcanic areas are known: Hoggar, Tibesti,  
79 Libyan basaltic fields (Liégeois et al., 2005; Fezaa et al., 2010). On the southwestern border  
80 of the Murzuq basin in Niger (between  $11^{\circ}53'$  E/ $22^{\circ}27'$  N and  $12^{\circ}04'$  E/  $21^{\circ}53'$  N),  
81 Paleozoic dolerites are mapped on the “Carte géologique du Sahara central” (Lelubre et al.,  
82 1962). These dolerites seem to be interbedded between Silurian and Devonian sedimentary  
83 sequences and correspond to a very large volume of mafic rocks (the biggest outcrop is  
84 indicated to be 90 km long on this map). Several hundred kilometers far to the west and  
85 southwest of In-Ezzane, Lower Carboniferous basic intrusions (Djellit et al., 2006; Derder et  
86 al., 2006) in the Tin Serririne basin and Devonian ring complexes in Aïr region (Black et al.,  
87 1967, Moreau et al., 1994) have been evidenced. In addition, Moreau et al. (2012) suggested  
88 that three areas with “sand injections”, which generated circular structures along the Murzuq  
89 basin boundaries during lower Devonian, could be related to deep phreatic eruption before  
90 Pragian times (Lower Devonian, 407-410 Ma). One of these structures is very close to the sill  
91 in the In-Ezzane area (Fig. 1A). We note that the In-Ezzane circular structures proposed to be  
92 meteoritic impacts (Bonin et al., 2011; Reimold and Koeberl, 2014) are located more to the  
93 West, within the Ordovician sandstones.

94 Such an intraplate magmatism is a key event for understanding the geodynamical  
95 evolution of the Saharan platform. The aim of this paper is to date this event and to determine  
96 its different implications.

97

## 98 **2. Geological setting**

99 The Murzuq Basin is the principal feature of Central Sahara. This basin lies across  
100 Libya, Algeria and Niger. From a geomorphological point of view, this intracratonic  
101 sedimentary basin can be described as an enormous flat hollow of approximately 900 km in

102 diameter from Libya in the north to Niger in the south. This basin is filled by different  
103 Phanerozoic series covered by a very large sandy desert (the “Murzuq edeyen or erg”). The  
104 main morphological units of this basin consist of plateaus, oueds and sand dunes cover. The  
105 Tassili-n-Ajjer (Algeria) and Djado (Niger) uplifted plateaus constitute the southwestern and  
106 southern margins of this basin. Primary geological formations dip gently from the flanks  
107 towards the basin centre in a concentric pattern. In the In Djerane and Tadrart areas (Figs. 1  
108 and 2), the different Paleozoic series (Cambro-Ordovician, Silurian, Devonian, Carboniferous  
109 and Permian – Henry et al., 2014) outcrop in several trenches of oueds, which produce  
110 remarkable rock cliffs (Fig. 2).

111 To the west, the Murzuq craton is bordered by its metacratonic margin, corresponding  
112 to the Djanet terrane, made of greenschist Ediacaran sediments (Fezaa et al., 2013) crosscut  
113 by 575-545 Ma old high-level plutonic bodies, batholiths, plutons and dyke swarms, (Fezaa et  
114 al., 2010). Further west, on the other side of a major NW-SE trending shear zone, the Edebo  
115 terrane resulted from a stronger intracontinental metacratonic reactivation that generated  
116 amphibolite-facies migmatitic gneisses and aluminous granitoids during the same period (c.  
117 568 Ma; Fezaa et al., 2010). This intracontinental reactivation (575-545 Ma  
118 metacratonization) occurred to the east of the Raghane mega-shear zone (western boundary of  
119 the Saharan metacraton; Abdelsalam et al., 2002) in a region previously stable, that was  
120 covered by sediments deriving from the more western Tuareg Shield in the 595-575 Ma time  
121 span (Fezaa et al., 2010). These metacratonic reactivations occurred mostly along pre-existing  
122 Eburnean (c. 2 Ga) lithospheric-scale terrane boundaries, along the Murzuq craton, of  
123 probable Archean age (Fezaa et al., 2010).

124 These metacratonic zones were reactivated during the Mesozoic, with deposition of  
125 several thousand meters of sediments in troughs, and during the late Cenozoic, with the  
126 emplacement of the In-Ezzane basaltic field (Liégeois et al., 2013). The Phanerozoic  
127 sediments of the Murzuq basin are located to the East, partly resting on the Murzuq  
128 metacratonic margin, the Djanet terrane.

129 The Arrikine area consists of a monoclinical structure dipping  $6^\circ$  toward E to ENE. This  
130 structural orientation gradually changes. The Murzuq geological map (Geological Map of  
131 Libya, 1985) shows that its borders, including Cretaceous levels, were uplifted and slightly  
132 tilted. Very locally, folding has been observed in the In-Ezzane area (Amenna et al., 2014),  
133 probably related to far-field effects of the Hercynian orogeny (Haddoum et al., 2001), again  
134 reactivating the metacratonic structures.

135 In this area, we discovered a large sill of mafic rocks (maximum observed NNW-SSE  
136 outcrop extension of about 35 km) interbedded in Silurian sediments (Fig. 2). The current  
137 description of the local stratigraphy (Freulon, 1964; Bellini and Massa, 1980; Legrand, 2000;  
138 Ghienne et al., 2013), together with the published maps of this area, do not mention any  
139 occurrence of mafic rocks.

140 The sill intruded the upper member of the *Oued In Djerane Formation*, mainly made  
141 of shales. This upper member displays signs of a more marine environment than the  
142 underlying levels, but in its top part, presents facies that are more and more regressive  
143 (Legrand, 2000). It is overlain by the detritic *Acacus Formation*. The host-rocks constituting  
144 the roof of the sill are affected by a contact metamorphism.

145

### 146 **3. Petrography and mineralogy**

147 The Arrikine sill (Fig. 2) has a maximum thickness of c. 250 m and is dipping gently  
148 to the ENE similarly to the host-rocks series. The Silurian host-rocks, both at the floor and  
149 roof of the sill, show local folding due to the intrusion. Farther from the intrusion, in the  
150 Ordovician series, deformations become brittle. The sill is constituted of a pyroxene-rich  
151 gabbro. On its borders, it is light gray in color due to a strong weathering, while in the other  
152 parts, it is dark in appearance and not or slightly altered. These dark-facies rocks are medium-  
153 grained gray-green gabbros. The main mineral is plagioclase appearing as large lengthened  
154 intertwined laths. The mafic mineral assemblage is dominated by granular clinopyroxene.  
155 Minor constituents are thin laths of biotite associated with oxides, few small quartz grains and  
156 relatively large granules of opaque minerals. Weathering renders plagioclase brownish in  
157 color and is at the origin of the incomplete replacement of some pyroxene by chlorite.

158 Microprobe analyses (Table 1) confirm that the magmatic plagioclase is a labrador  
159 ( $An_{55}$ ), the pyroxene a diopside, the presence of biotite and apatite and that the oxides are  
160 ilmenite and magnetite. The biotite is particularly rich in  $TiO_2$  (c. 7%), which is also a  
161 characteristic of the biotite from the Iskou ring complex in Aïr (Léger, 1985). These analyses  
162 also show that Fe-Mg minerals are often altered but above all, that an aplitic component is  
163 present and is represented by small spots of K-feldspar, with diverse K compositions ranging  
164 from 2 wt % to 15 wt %. This K enrichment is accompanied by a Ba-enrichment (Table 1)  
165 developed within acid plagioclase ( $An_{13-17}$ ), two minerals that cannot be in equilibrium with  
166 the gabbroic magma. The development of this phase enriched in K and Ba, has consequences  
167 for the whole rock geochemistry and its K-Ar dating (see below).

168

## 169 **4. Isotopic dating**

170

### 171 **4.1 K/Ar geochronology**

172 Analyses were done by one of us (H.B) at the K-Ar dating platform of UMR 6538  
173 “Domaines Océaniques” CNRS, UBO at Brest.

#### 174 *4.1.1 Analytical procedure*

175 This sample of gabbro (AF-ALG/Dj-Site58 from site 58 – Fig. 2) was also considered  
176 for dating on a whole-rock fraction made of grains 0.3 to 0.15 mm in size, prepared after  
177 hand-crushing and sieving of the solid rock, then cleaned using ultrapure water. One aliquot  
178 of grains was powdered in an agate grinder. Then two independent dissolutions of 80 mg of  
179 powdered sample were realized in two well-sealed synthetic bottles using 4 cm<sup>3</sup> of  
180 hydrofluoric acid at a temperature of 80°C for 8 hours. K content of the two HF solutions was  
181 determined by Atomic Absorption Spectrometry after dilution (1/2500).

182 A second aliquot of grains was reserved for under vacuum extraction of all gases by  
183 heating in a molybdenum crucible, using a high frequency generator. Cleaning of active gases  
184 from the whole extracted gases was realized during and after their extraction by gettering on  
185 hot titanium sponge contained in three quartz traps successively activated when their  
186 temperature was decreasing from 800°C to the ambient one during 10 minutes. The final step  
187 was the ultra-purification using an Al-Zr SAES getter.

188 Isotopic composition of argon and concentration of <sup>40</sup>Ar<sub>R</sub> were measured in a stainless  
189 steel mass spectrometer with a 180° geometry. Isotopic dilution was realized during the fusion  
190 step, using precise concentrations of <sup>38</sup>Ar buried as ions in aluminium targets (Bellon et al.,  
191 1981).

192 A huge degasing of grains was observed and measured for the two first extractions  
193 from 0.4 to 0.3 g of sample and led to a long and difficult purification of argon; consequently  
194 it was less easy to reduce the effects of signal evolution through time . Two complementary  
195 extractions of gases from significantly lighter weights (0.07 and 0.06 g) have permitted to  
196 restore better conditions for purification and for isotopic analysis.

#### 197 *4.1.2 Data*

198 Resulting data of the four extractions and isotopic analyses are presented in Table 2.  
199 Ages are calculated using Steiger and Jäger's (1977) constants and errors following the  
200 equation of Cox and Dalrymple (1967).

### 201 4.1.3 Interpretation

202 A mean age of  $325.6 \pm 7.7$  Ma results from the analyses, i.e. a Serpukhovian age (from  
203 uppermost Visean to upper Bashkirian, if taking error margins into account). The Silurian age  
204 of the host sedimentary series questions the significance of the K-Ar age particularly when  
205 considering the huge degassing of samples under vacuum that may indicate the effects of  
206 secondary processes that affected the Arrikine sill.

207 Such a result points to an event occurring during the Variscan period. Therefore, it is  
208 legitimate to raise the question whether a geochemical remobilization of the sill, leading to a  
209 potassium and barium enrichment in the plagioclases (Table 1), has occurred. This alteration  
210 could have been promoted by cryptocirculations of fluids (supposed hot but not exactly  
211 known) that may have reset the K-Ar system by complete external diffusion of radiogenic  
212  $^{40}\text{Ar}$ . Following such a hypothesis, the 325.6 Ma age marks the time to which the gabbro  
213 should have been rejuvenated during the Variscan evolution.

214

## 215 4.2 U-Pb geochronology

### 216 4.2.1 Analytical procedure

217 Zircons were separated from 4 kg of rocks and were subsequently processed according  
218 to standard technique (e.g. Bosch et al., 1996), including magnetic separation and careful  
219 hand-picking under a binocular microscope. U-Pb dating on zircon was performed by laser  
220 ablation sector field ICP-MS at Geosciences Montpellier (France). The ICP-MS  
221 (ThermoFinnigan Element XR high resolution) was coupled to a Lambda Physik CompEx  
222 102 excimer laser generating 15 ns duration pulses of radiation at a wavelength of 193 nm.  
223 Data were acquired at low resolution ( $\Delta M/M = 300$ ) in the fast electrostatic scan mode.  
224 Signals were measured in pulse counting mode using 3 points per peaks and a 20% mass  
225 window. A pre-ablation step consisting of 7 pulses was applied to clean the sample surface.  
226 The acquisition step began with 15 s of gas blank measurement followed by 45 s of laser  
227 ablation at a frequency of 4 Hz with a spot size of 26  $\mu\text{m}$  and an energy density of 10  $\text{J}/\text{cm}^2$ .  
228 Measured isotopic ratios were monitored with reference to the G91500 zircon standard with a  
229  $^{206}\text{Pb}/^{238}\text{U}$  ratio of 0.17917 and a  $^{207}\text{Pb}/^{206}\text{Pb}$  ratio of 0.07488 equivalent to ages of 1062 Ma  
230 and 1065 Ma, respectively (Wiedenbeck et al., 1995). The standard was measured four times  
231 each five unknowns in a sequence of 2 standards, 5 unknowns and 2 standards (see Bosch et  
232 al., 2011 for details). Analytical data are plotted and ages calculated using the IsoplotEx



233 program (Ludwig 2000). Analyses in Table 3 and in the Concordia plots are  $\pm 1\sigma$  errors and  
234 uncertainties in ages quoted in the text are given at  $\pm 2\sigma$ .

#### 235 4.2.2 Results

236 Only 17 zircons were recovered, which is not so surprising in a basic rock. They  
237 display a large range of shapes and colors, indicating they do not form a homogeneous  
238 population. Most grains are subhedral to rounded, translucent and colorless to pink in color.  
239 Although some grains have euhedral shapes, the variety of shapes and the occurrence of  
240 rounded grains indicate that most grains may have been snatched from the country rocks,  
241 which possibly include a detrital component. Reported in the Concordia diagram (Fig. 3),  
242 analyzed grains display a wide range of ages, from Neoproterozoic to Archean, ranging from,  
243  $598\pm 14$  Ma to  $2707\pm 38$  Ma ( $2\sigma$ , see Table 3). The main population is Paleoproterozoic with a  
244 mean age of around 2030 Ma ( $n=7$ ). The Neoproterozoic population is dominated by a  
245 coherent group of four grains plotting concordantly at *c.* 682 Ma, while other grains are  
246 concordant at 766, 632 and 598 Ma. Lastly the recovered zircon population includes a  
247 homogeneous Archean component dated at *c.* 2700 Ma.

#### 248 4.2.3 Interpretation

249 The Arrikine sill cannot be older than the intruded Silurian sediments. The dated  
250 zircons are therefore xenocrysts and not new minerals crystallized within the sill. This agrees  
251 with its low-silica geochemical composition and high temperature of crystallization,  
252 conditions during which the zircon saturation is very rarely attained (Watson and Harrison,  
253 1983; Boehnke et al., 2013). The inherited zircons can have been incorporated in the Arrikine  
254 magma by incorporating (1) unmelted zircons from the source, (2) zircons from the basement  
255 during its ascent or (3) zircons from the Ediacaran or Paleozoic sediments during its final  
256 emplacement, these possibilities being not mutually exclusive.

257 Incorporating unmelted zircon crystals from the source is unlikely, again because such  
258 magmas are far from zircon saturation and are thus efficient to dissolve zircon (Watson and  
259 Harrison, 1983). Moreover, the source of the Arrikine gabbro (the mantle or the lower mafic  
260 crust) itself most probably bears very few zircons for the same reason.

261 The ages of the inherited zircons in the Arrikine gabbro (0.6, 0.63, 0.68, 2.03 and 2.7  
262 Ga) match those of the detrital zircons from the Ediacaran sediments of the Djanet terrane,  
263 just to the west and probably present at depth below the Arrikine area (Fezaa et al., 2010).  
264 However, 2.7 Ga zircons are, strictly speaking, not present in these Ediacaran sediments (only  
265 2.65 and 2.8 Ga old zircons are present). In addition, it is noteworthy that they bear other age

266 peaks, especially at 0.7, 1.9 and 2.5 Ga, which are not present in the Arrikine sill. The source  
267 of these sediments is Central Hoggar, especially its closest eastern part, explaining that some  
268 samples have a proximal immature source with few detrital zircon ages (Fezaa et al., 2010;  
269 2013). So, the Ediacaran sediments could have provided some of the inherited zircons in the  
270 Arrikine sill, but other sources should be found.

271 The detrital zircons from the Tassilis Ordovician sandstones just to the West  
272 (Linnemann et al., 2011) display much more numerous age peaks. These sediments are  
273 considered as deriving from sources located all over West Africa and even Amazonia  
274 (Linnemann et al., 2011). However, the 2.7 Ga age is systematically absent. Incorporation of  
275 detrital zircons from these Ordovician mature sediments is thus not favored. The same  
276 conclusion can be reached for the Silurian series in which the Arrikine sill intruded. Its source  
277 is also distal and it is made of shales whose fine granulometry does not favor the presence of  
278 zircons.

279 The basement is not outcropping in the Easternmost Hoggar. The oldest known rocks  
280 are the Ediacaran sediments, metamorphosed in the greenschist facies (Djanet terrane) or  
281 migmatized in the amphibolite facies (Edembo terrane), intrusive and subvolcanic rocks being  
282 younger (Fezaa et al., 2010). Ediacaran magmatic rocks could have had extracted material but  
283 few old ages have been measured in their zircons. The Tin Amali dyke swarm for example  
284 provided a  $2090 \pm 40$  Ma age (Fezaa et al., 2010) for an inherited component. This age is not  
285 present in the detrital zircon age spectrum of the Ediacaran sediments and is likely to be  
286 related to the age of one basement component. More information can be obtained from the Nd  
287 isotopes. The mean  $T_{DM}$  model ages for the various magmatic bodies (batholith, pluton, and  
288 dyke swarm in the 575-545 Ma age range) fall in the 1.6-1.8 Ga age interval (Fezaa et al.,  
289 2010). This points to a Paleoproterozoic and/or Archean basement in the area. More to the  
290 west, but still to the east of the Raghane shear zone that constitutes the western boundary of  
291 the Saharan metacraton, the basement of the Aouzegueur terrane (the third Eastern Hoggar  
292 terrane; Fig. 1A) has been dated at  $1918 \pm 5$  Ma (LA-ICP-MS U-Pb zircon age; Nouar et al.,  
293 2011). This basement is intruded by various granitoids close to the Raghane shear-zone at c.  
294 790 Ma, c. 590 and c. 550 Ma (Henry et al., 2009; Nouar et al., 2011). All these ages do not  
295 correspond to the inherited zircons dated in the Arrikine sill.

296 So few informations are available on the basement below the Arrikine sill. Nd isotopes  
297 indicate that the basement is Paleoproterozoic and/or Archean in age and an inherited zircon  
298 point to an age of  $2090 \pm 40$  Ma. On the other side, the Arrikine inherited zircons have two  
299 well-defined old ages at  $2698 \pm 22$  Ma (3 zircons) and  $2028 \pm 15$  Ma (7 zircons) with no other

300 zircons older than 1 Ga. These zircons can be considered as having been snatched from the  
301 underlying basement, but in very small quantities due to the low viscosity of this magma, and  
302 not from the superficial sediments that bear many ages but these ones. On the other hand, the  
303 younger inherited zircons, at  $682 \pm 7$  Ma (4 zircons),  $632 \pm 10$  Ma (1 zircon) and  $598 \pm 15$  Ma  
304 (1 zircon) could be sourced from the Ediacaran sediments that derived from the erosion of  
305 Neoproterozoic units close to the Raghane shear zone.

306 The Arrikine sill, through its inherited zircons, is thus currently the best probe for  
307 getting information on the basement of the Murzuq craton, at least of its western boundary,  
308 indicating the existence of Archean (c 2.7 Ga) and Paleoproterozoic (c. 2.03 Ga) lithologies at  
309 depth. On the other hand, due to the absence of magmatic zircons from the sill itself, the  
310 zircon dating was not suitable to date emplacement of the Arrikine sill.

311

## 312 **5. Paleomagnetic dating**

### 313 ***5.1 Sampling and analysis procedure***

314 The sampling sites for paleomagnetism are located near the oued Arrikine (Fig. 2).  
315 110 standard paleomagnetic samples, taken using a portable gasoline-powered drill and  
316 oriented with magnetic and sun compasses, were collected from the 12 sites (sites 50 to 61 on  
317 Fig. 2) in the gabbro (including one site in the weathered gray facies on the sill border and  
318 another one within a late dyke). 9 samples were chosen in one site (site 49 on the Figure 2) of  
319 Silurian host-rocks affected by contact metamorphism close to the roof of the sill. One to  
320 three specimens were cut from each core in order to have additional specimens for pilot  
321 studies and rock magnetic analysis. Prior to any demagnetization analysis, the specimens were  
322 stored in a zero field for at least one month, in order to reduce possible viscous components.  
323 The Natural Remanent Magnetization (NRM) of the specimens was measured using a JR5  
324 spinner magnetometer (AGICO, Brno). Several pilot specimens from each site were subjected  
325 to a stepwise alternating field (AF) demagnetization up to 100 mT or thermal (TH)  
326 demagnetization up to  $670^{\circ}\text{C}$  in order to characterize the components of magnetization.  
327 Following combined demagnetization process, some other specimens were subjected to  
328 combined AF and TH treatments. The directions of the magnetization components were  
329 plotted on orthogonal vector plots (As and Zijderveld, 1958; Wilson and Everitt, 1963;  
330 Zijderveld, 1967) and the remaining vectors and vectorial differences of the magnetization on  
331 equal-area projections. The direction of the different components was computed using  
332 principal component analysis (Kirschvink, 1980). Fisher (1953) statistics was used to  
333 determine the mean directions.

334

335 **5.2. Rock magnetism**

336 The NRM has a relatively high intensity in dark facies (mean 3.2 A/m) compared to  
337 the weathered site close to the sill border (mean  $2.7 \cdot 10^{-2}$  A/m) and the late dyke (mean 5.3  
338  $10^{-2}$  A/m). AF demagnetization yields a relatively regular decrease of magnetization intensity.  
339 For most samples, magnetization decrease during thermal demagnetization (Fig. 4) is  
340 progressive (IZ647, IZ668) or steep above 400-450°C treatment up to about 530-560°C  
341 (IZ654). These maximum blocking temperature values suggest the presence of Ti-poor  
342 titanomagnetite as the principal mineral carrier. In few samples (Fig. 4), a strong increase of  
343 the magnetization for about 400°C is followed by a sharp decrease around 450°C up to about  
344 530-560°C (IZ718, IZ729). No increase appears during AF demagnetization of these samples,  
345 for which complementary studies of the magnetic properties have to be performed and will be  
346 the subject of an independent paper.

347 Thermomagnetic curves (low-field magnetic susceptibility as a function of  
348 temperature) have been measured for representative samples by heating in air using CS2–3  
349 oven and KLY3 Kappabridge (AGICO, Brno). For the gray weathered facies, during heating,  
350 the partial cooling curves are often reversible with the heating ones (Fig. 5 - IZ696). In such  
351 cases, irreversibility of the thermomagnetic curves appears only after heating at the highest  
352 temperatures. During heating at temperatures higher than 400°C, a decrease of susceptibility  
353 is observed, giving Curie temperature of about 550°C. This confirms that the principal  
354 magnetic carrier is a Ti-poor titanomagnetite. However for some samples, a slight decrease  
355 for temperatures higher than 580°C points out the presence of a small amount of hematite  
356 besides the titanomagnetite (Fig. 5 - IZ696). In the dark facies, the sample thermomagnetic  
357 curves are not reversible since 400 to 500°C heating, highlighting a relatively complicated  
358 mineralogical alteration during heating at higher temperature (Fig. 6 IZ716).

359 Hysteresis loops were obtained for small cores (cylindrical samples of about 3 cm<sup>3</sup>)  
360 using a laboratory-made translation inductometer within an electromagnet capable of reaching  
361 1.6 T. The loops show that the remanent coercive force determined from several  
362 representative samples is low (around 30 mT) and that a partial saturation of the  
363 magnetization is at least reached between approximately 0.2 to 0.4 T (IZ678, IZ725),  
364 confirming the presence of a low coercivity component (like Ti-poor titanomagnetite). For  
365 weathered gray facies, susceptibility is very low and almost linear hysteresis loops have been  
366 obtained, indicating largely dominating effects of the paramagnetic minerals. For dark facies  
367 (Fig. 6), a much higher susceptibility is associated with a clear dominance of ferrimagnetic

368 minerals. For these sites, Day plot (Day et al., 1977) indicates pseudo single-domain (PSD)  
369 grain size for the titanomagnetite (Fig. 6).

370 In conclusion, Ti-poor titanomagnetite is the main magnetic carrier in the non-  
371 weathered gabbro. It is also present in the weathered gray facies, but in much lower amount  
372 and associated with few hematite.

373

### 374 **5.3. Paleomagnetic analyses**

375 TH or AF or mixed AF-TH demagnetizations (Fig. 7) were performed on all samples.  
376 Very similar Characteristic Remanent Magnetization (ChRM B) data have been obtained  
377 from all methods, in addition to a viscous component A. During the demagnetization process,  
378 three different main kinds of evolution were observed for gabbro sites:

379 - The first type, obtained on half the samples, is illustrated by a single stable magnetic  
380 direction. Figures 7a, b and g show examples of this behavior obtained from an AF method  
381 (IZ659), thermal process (IZ 644) or mixed AF-TH (IZ660A). In this case, after elimination  
382 of the viscous component A, the analysis of the magnetization shows that the linear segments  
383 towards the origin on Zijdeveld (1967) plots, define the ChRM B.

384 - The second type of behavior shows magnetic directions evolving first along a great circle  
385 (remagnetization circle), which indicates superimposition of the unblocking temperature (or  
386 unblocking fields) spectra for at least two remanent magnetization components. This  
387 evolution always reaches a “stable end point” direction yielding the ChRM B. This case is  
388 illustrated on the figures 7c and d for the AF method (IZ 709) and the thermal method (IZ  
389 739). Here also, similar results are obtained by the two method of demagnetization. The  
390 secondary component at the origin of the variation along the great circles cannot be properly  
391 identified. It could correspond to a Cenozoic overprint (Henry et al., 2004; see also for  
392 example Hargraves et al., 1987), but other origins cannot be rejected by lack of constraints.

393 - The third case (Figs. 7e, f) displays similar variation than the first case, except that, during  
394 the thermal treatment, a component, with the same direction as the main magnetization but  
395 with opposite polarity, appears at c. 400°C and progressively disappears at higher temperature  
396 (from c 500°C), yielding a well-defined ChRM magnetization. During AF demagnetization,  
397 this intermediate component of opposite polarity is not present. The final ChRM directions  
398 obtained for both demagnetization processes are the same. This additional component is  
399 therefore related to a mineralogical alteration due to heating. These samples will be the  
400 subject of a complementary analysis about magnetic properties.

401 The ChRM B obtained from samples with these three behaviors was isolated between  
 402 200 and 500 °C for the lowest temperatures and 400–560°C for the highest ones (thermal  
 403 process), and between 10-70 mT for the lowest blocking field and 60 and 102 mT for the  
 404 highest ones (AF demagnetization).

405 Results from the sites sampled in weathered facies on the sill border (14 samples) and  
 406 within a late dyke (12 samples) are mostly not reliable. The obtained results from all the other  
 407 samples are homogeneous (Table 4). The mean direction of this ChRM B computed from 84  
 408 specimens (from 12 sites) has the following directions:  $D = 355.4^\circ$ ,  $I = -38.4^\circ$ ,  $k = 38$ ,  
 409  $\alpha_{95} = 2.5^\circ$  and  $D = 350.7^\circ$ ,  $I = -38.6^\circ$ ,  $k = 38$ ,  $\alpha_{95} = 2.5^\circ$ , before and after dip  
 410 correction, respectively (Table 4 – Fig. 8). This result is similar to that obtained from the mean direction  
 411 of the 10 sites (two sites with less than 3 ChRMs being not considered):  $D = 355.5^\circ$ ,  
 412  $I = -38.8^\circ$ ,  $k = 86$ ,  $\alpha_{95} = 5.2^\circ$  and  $D = 350.6^\circ$ ,  $I = -38.9^\circ$ ,  $k = 86$ ,  $\alpha_{95} = 5.2^\circ$ , before and after dip  
 413 correction, respectively (Fig. 8).

414 The paleomagnetic data obtained from the site of Silurian host-rocks affected by  
 415 contact metamorphism have similar characteristics with neighboring ChRM orientation (N=9  
 416 samples,  $D = 358.6^\circ$ ,  $I = -42.8^\circ$ ,  $k = 402$ ,  $\alpha_{95} = 2.6^\circ$  and  $D = 349.5^\circ$ ,  $I = -41.7^\circ$ ,  $k = 402$ ,  
 417  $\alpha_{95} = 2.6^\circ$ , before and after dip correction, respectively) but with lower magnetization  
 418 intensity. Other sites, taken in the Ordovician and Silurian rocks far from the sill (sites 43 to  
 419 48 on Fig. 2), yield only weak Cenozoic remagnetization of normal and reversed polarity (see  
 420 Henry et al., 2004), totally overprinting any possible previous earlier magnetization. That is  
 421 then not the case for the site affected by contact metamorphism. The samples from this site  
 422 carried a much more stable magnetization than the other sites. That is attributed to an effect of  
 423 the contact metamorphism generated by the sill. This kind of positive contact test strongly  
 424 argues for a primary magnetization of the gabbro.

425 The corresponding paleomagnetic pole determined from the sill data is defined by  
 426  $\lambda = 43.6^\circ\text{S}$ ,  $\phi = 17.0^\circ\text{E}$ ,  $K = 88$  and  $A_{95} = 4.7^\circ$  and  $\lambda = 42.8^\circ\text{S}$ ,  $\phi = 22.9^\circ\text{E}$ ,  $K = 88$  and  
 427  $A_{95} = 4.7^\circ$ , before and after bedding correction, respectively.

428

#### 429 ***5.4 Paleomagnetic dating of the Arrikine sill***

430 A gabbro magnetization age younger than the post-Cretaceous tilting of the formations  
 431 is unrealistic when comparing the Arrikine poles with the known Apparent Polar Wander Path  
 432 - APWP - (Fig. 9) and data can only be used after applying a bedding correction. The so  
 433 corrected Arrikine pole (Fig. 9) corresponds to the Gondwana APWP segment for the 400 Ma  
 434 period of Derder et al., 2006, Amenna et al., 2014 and Henry et al., 2015 and to that of

435 Torsvik et al., 2012 (overlapping of confidence zones) for the 410 Ma mean pole. In the latter,  
436 the uncertainty on the paleomagnetic mean pole age is not taken into account while it  
437 corresponds to a single pole (Aïr intrusives, Hargraves et al., 1987), which can explain the  
438 difference existing with the former one. Indeed, integration of such uncertainties in the first  
439 APWP adds a weighted effect for poles having a large age uncertainty and thus covering  
440 several windows in time, smoothing the curve. This is particularly critical for periods affected  
441 by strong changes of the APWP drift such as the 400-410 Ma period (Fig. 9). Geologically,  
442 the Arrikine sill is post-Silurian and pre-Cretaceous. The Arrikine pole is very far from all  
443 mean paleopoles from Middle Devonian to Present times, especially for Carboniferous times.  
444 It is also far from Silurian mean paleopoles (Fig. 9). By contrast, the Arrikine pole is  
445 consistent with Early Devonian poles of Gondwana: Mt. Leyshon Devonian dykes in  
446 Australia (Clark, 1996), Snowy River Volcanics in Australia (Schmidt et al., 1987), Herrada  
447 Member, Sierra Grande in Argentina (Rapalini and Vilas, 1991) and Aïr alkaline ring-  
448 complexes in Niger (Hargraves et al., 1987). In particular, the coincidence of the Arrikine  
449 pole with this last pole ( $\lambda = 43.4^\circ\text{S}$ ,  $\phi = 8.6^\circ\text{E}$ ,  $K = 50$  and  $A_{95} = 6.2^\circ$ ) is remarkable. The  
450 angular difference between these paleomagnetic poles ( $10.7^\circ$ ) is of the order of what can be  
451 ascribed to the secular variation. In both the Arrikine sill and the Aïr alkaline ring-complexes,  
452 the paleomagnetic directions have the same single polarity. So the Aïr ring-complex age of  
453  $407 \pm 8$  Ma (Moreau et al., 1994) can be considered as a good approximation of the age of the  
454 Arrikine sill, which can be fixed to the 400-415 Ma age interval, i.e. at the base of the  
455 Emsian, from mid-Lochkovian to mid-Emsian (Cohen et al., 2013), thus globally Lower  
456 Devonian.

457 The loop of the Gondwana APWP between the Late Ordovician (pole 450 Ma) and  
458 Late Devonian (370 Ma) was not generally deemed as very reliable. The Aïr and Arrikine  
459 coherent mean is "unique" and supports the existence of this loop that has to be considered for  
460 the paleocontinental reconstructions for this period.

461

## 462 **6. Geochemistry**

463 Only one analysis (sample AF-ALG/Dj-Site58 from site 58 – Fig. 2) is available and  
464 listed in Table 5. However, its geochemical signature (Fig. 10) is sufficiently diagnostic to  
465 allow a proper geochemical characterization of the Arrikine sill and to deduce constrained  
466 conclusions about its origin. All oxides are considered in column a\* as deduced from the  
467 geochemical analysis by ICP and in column b\* are recalculated to 100% on an anhydrous  
468 basis. Oxide % are always wt. %, unless otherwise stated. The Arrikine sill is a fine-grained

469 gabbro but (see below) its chemistry is similar to intraplate alkali basaltic rocks in continental  
470 flood basalt (CFB) and ocean island basalt (OIB) provinces. So, in order to make the  
471 comparison more efficient, geochemical diagrams dedicated to volcanic rocks and volcanic  
472 nomenclature are applied here, which is justified by the subvolcanic character of the sill and  
473 by the difficulty of counting a mode on this kind of rock.

474 The Arrikine gabbro having 45.99% SiO<sub>2</sub> and 5.79% Na<sub>2</sub>O+K<sub>2</sub>O, on an anhydrous  
475 basis, belongs to the alkali series, being an alkali gabbro. It has the composition of a  
476 trachybasalt (Le Bas et al. 1986), more precisely of an hawaiiite (wt%Na<sub>2</sub>O-2 > wt%K<sub>2</sub>O), i.e.  
477 a sodic trachybasalt (Le Maitre et al., 2002). The depleted MgO content (4.1%) corresponds  
478 to typical hawaiiite (3 - 5.5 % MgO; MacDonald and Katsura, 1964) as is the absence of  
479 chromite (Cr: 3 ppm; Ni: 39 ppm), a characteristic of hawaiiite (Cr: 5-150 ppm; Ni: 5-80 ppm;  
480 Nelson and Carmichael, 1984 and references therein), in agreement with the differentiated  
481 character of hawaiiites. Immobile trace elements confirm that the Arrikine gabbro has the  
482 composition of an alkali basalt with Nb/Y= 0.85 and Zr/TiO<sub>2</sub>\*0.0001= 0.0064 (diagram not  
483 shown; Winchester and Floyd, 1977). Low concentration of Ni, Cr and absence of positive Eu  
484 anomaly do not point to a significative cumulative component. Even the very low Cr content  
485 in Arrikine gabbro points to a differentiated character.

486 In addition, with TiO<sub>2</sub> = 5.01%, the Arrikine gabbro belongs to the high-Ti series  
487 (TiO<sub>2</sub> >2%); it has the composition a high-Ti alkali basalt. CIPW norm points to 60%  
488 plagioclase (An<sub>30</sub>), 11% K-feldspar, 9% pyroxene, 6% olivine, 2% apatite and 11% Fe-Ti  
489 oxides (Table 5) in rough agreement with the petrography except that the plagioclase is acid  
490 (actually Labrador) and the absence of biotite, not computed in the norm. This norm is in  
491 agreement with a composition of a high-Ti hawaiiite. The important difference in the An  
492 content of the plagioclase can be ascribed to the late “aplitic” phase determined by the  
493 microprobe measurements to be acid oligoclase (An<sub>13-17</sub>). The An<sub>30</sub> value determined by  
494 the norm suggests that this aplitic phase was quite pervasive at least in the analyzed sample.  
495 This implies that concerned elements (i.e. K, Ba but also Rb, Sr) must be used with caution  
496 when the original magma is concerned as well as the K-Ar age for which we have calculated  
497 (method based on an initial age of 410 Ma given by paleomagnetic considerations and the  
498 rejuvenated result at 326 Ma) a K enrichment of up to 20 %.

499 The contemporaneous (407 ± 8 Ma; Moreau et al., 1994) large province of alkaline  
500 ring-complexes of Aïr bears also high-Ti gabbros, especially the large Ofoud ring-complex  
501 where they are associated with numerous layers and/or lenses of titanomagnetite up to 10 m  
502 thick (Karche and Moreau, 1978) and the smaller Abontorok ring-complex, associated with



503 massive anorthosites (Brown et al., 1989). These gabbros are of uniform grain size (2-7 mm)  
504 and massive in texture and titanomagnetite and ilmenite can make up as much as 10 percent  
505 of the rock (Hush and Moreau, 1982). The Aïr ring-complex province is intrusive along the  
506 western boundary of the Saharan metacraton (Liégeois et al., 1994, 2013) as a consequence of  
507 its reactivation during early Devonian (Moreau et al., 1994). The Arrikine gabbro is compared  
508 to the available analysis from Abontorok ring complex (Brown et al., 1989) and to several  
509 intraplate continental high-Ti alkali basalt or gabbro provinces: the Trias rift-related from SW  
510 China, similar to the high-Ti basalts from the Emeishan LIP (Zhang et al., 2013), the  
511 Ordovician Suordakh and Cambrian Kharaulakh mafic rocks from Siberia (Khudoley et al.,  
512 2013) or the 1.5 Ga Yangtse mafic province (Fan et al., 2013). We added also for comparison  
513 the intraplate high-Ti potassic basalts from the Virunga (Rogers et al., 1998).

514         When compared to these high-Ti alkali basalts from different settings and different  
515 ages, the Arrikine gabbro has globally a similar pattern (Fig. 10). Compared to Mauna Loa  
516 volcano in Hawaii (Wanless et al., 2006), the Arrikine gabbro displays a subparallel REE  
517 pattern, being slightly more enriched, especially in LREE (Fig. 3a). Similarly, its pattern in  
518 the spidergram (Fig. 10b) is parallel, only very slightly enriched in HFSE, more enriched in  
519 LILE especially in Ba. High-Ti basalts have a largely preponderant mantle origin by contrast  
520 to low-Ti basalts (e.g. Deckart et al., 2005), explaining similar compositions in the oceans and  
521 in the continents. This characteristic renders, however, this kind of magmas relatively rare  
522 within the continents, which is an important fingerprint of the Arrikine sill. The spatially close  
523 Aïr high-Ti gabbros (Abontorok ring complex; Brown et al., 1989) display REE patterns close  
524 to that of Arrikine (Fig. 10a) except for Eu, probably due to a plagioclase cumulative pattern  
525 (Aïr ring-complexes comprise massive anorthosites - Brown et al., 1989; Demaiffe et al.,  
526 1991). Abontorok spidergrams are also close to that of Arrikine, the latter being on the most  
527 enriched side. This can be related to its low Cr content suggesting a differentiated character,  
528 the differentiation being able to concentrate incompatible elements. However, in the Arrikine  
529 gabbro Rb, Ba and K appear particularly enriched; these enrichments can be related to the  
530 presence of the late metasomatic aplitic phase (See K-Ar section; Table 5) probably favored  
531 by the sedimentary environment. This is particularly visible when compared with the  
532 Suordakh high-Ti basalt from Siberia (Khudoley et al., 2013) in a primitive mantle-  
533 normalized graph (Fig. 11): the Arrikine basalt is very similar to the Suordakh basalts but  
534 with concentrations in K as high as the richer Suordakh basalts and with Ba contents much  
535 higher than any Suordakh basalts. This indicates that the K, Rb and Ba concentrations in the  
536 pristine Arrikine gabbro were probably lower than that currently observed. By opposition, the

537 other elements (HFSE, Th) can be considered as pristine, being not mobilized by an aplitic  
538 fluid.

539 If the Arrikine gabbro composition is similar to the intraplate alkaline continental  
540 high-Ti alkali basalt or gabbro provinces, it is distinct from the intraplate high-Ti potassic  
541 basalts from the Virunga (Rogers et al., 1998), except for Ba. By opposition to the latter, the  
542 sources of the formers are considered to be a more or less enriched asthenospheric source with  
543 very low lithospheric participation, similarly to Hawaii volcanoes but in a continental setting.  
544 The similarity of the geochemistry of the Arrikine sill with the lower Devonian Air  
545 magmatism concurs to support the lower Devonian age determined in the paleomagnetic  
546 section.

547 Based on the silica content, a simple calculation leads to an estimated liquidus  
548 temperature of 1280°C. When considering the experimental work of Botcharnikov et al.  
549 (2008) determining a liquid line of descent for ferro-basalts, the Arrikine concentrations in  
550 SiO<sub>2</sub>, MgO, TiO<sub>2</sub>, K<sub>2</sub>O, Na<sub>2</sub>O, Al<sub>2</sub>O<sub>3</sub>, FeO, CaO point all to a temperature of c. 1080°C,  
551 which is the temperature where the maximum concentration in FeO and TiO<sub>2</sub> is attained. This  
552 is true when considering nearly anhydrous magmas, which is the case of Arrikine (0.2% H<sub>2</sub>O,  
553 Table 5). Considering the temperature of the mantle partial melting at 1280°C and the  
554 crystallization of the Arrikine sill at 1080°C seems reasonable. This is indeed the temperature  
555 range observed in experimentation concerning Fe-rich basalts (Chevrel et al., 2014). The  
556 calculated viscosity at liquidus temperature is very low (0.17 Pas), as it is the case for Fe-rich  
557 basalts whose viscosity is less than 1 Pas (Chevrel et al., 2014). In a 3 m-width dyke, a  
558 magma rise speed higher than 10 m/s can be proposed, a speed that limits fractional  
559 crystallization and interaction with country-rocks during ascent (Chevrel et al., 2014). This  
560 implies that the differentiation that affected the Arrikine magma (low Cr contents) occurred at  
561 depth.

562 Most of these provinces are considered to be or having been fed by a mantle plume.  
563 Notwithstanding the fact that the existence of mantle plumes is currently challenged (e.g.  
564 Foulger and Hamilton, 2014), there is no one argument for a plume origin in the case of the  
565 Arrikine sill: very small magmatic volume, no swell, no rift, no magmatic track, even if a  
566 similar asthenospheric mantle source can be inferred for the Arrikine sill. The proposed  
567 alternative will be discussed below in the geodynamic section.

568

## 569 **7. Emplacement and reactivation of the Arrikine sill: why there and why at these times?**

570

### 571 **7.1 Regional considerations**

#### 572 *7.1.1. The time of emplacement, the Lower Devonian*

573 During the Pan-African orogeny and the Phanerozoic, the Murzuq craton played an  
574 important role as a rheologically rigid body. Magmatism and compressive (Pan-African) or  
575 extensional (Phanerozoic) tectonics occurred around it, and the Murzuq basin developed as an  
576 intracontinental sag basin in a similar way as Al Kufra and Chad basins (Heine et al., 2008).  
577 Formation of the Murzuq basin did not require accommodation of tectonic subsidence through  
578 faulting and rifting but resulted from the negative buoyancy caused by thick cratonic  
579 lithosphere (Ritzmann and Faleide, 2009). As a consequence, the Murzuq craton is devoid of  
580 magmatism except along its metacratonic boundaries where Cenozoic volcanic fields are  
581 particularly present all around it (Fezaa et al., 2010; Liégeois et al., 2013). Indeed, the West  
582 African Cenozoic volcanism is attributed to tectonic reactivations induced by the Europe-  
583 Africa convergence (Liégeois et al., 2005; Beccaluva et al., 2007; Lustrino et al., 2012;  
584 Radivojević et al., 2015). These reactivations, even induced by the rather low stress  
585 propagating in remote intracontinental regions, must affect the whole lithospheric thickness  
586 allowing for small mantle melt volumes to reach the surface. This is the case of the mega-  
587 shear zones within metacratonic areas that are rigid but fractured (Liégeois et al., 2003, 2005,  
588 2013; Bardintzeff et al., 2011) through vertical planar delamination as recently evidenced by  
589 magnetotelluric studies in Hoggar (Bouzig et al., 2015).

590 The Arrikine sill is lower Devonian in age and intruded Silurian sediments. It can be  
591 correlated with Paleozoic dolerites in Niger (Fig. 1B), which are also intrusive within Silurian  
592 sediments, suggesting a regional event along the southwestern border of the Murzuq basin,  
593 even if limited in volume.

594 By opposition, no direct evidence of magmatism, even as detrital pebbles, were  
595 described within the sedimentary Devonian formations of the Murzuq basin (Bellini and  
596 Massa, 1980). This fact is tentatively related to the so-called Caledonian event (base of the  
597 Devonian) that prevented the deposition of lower Devonian sediments (Ghienne et al., 2013).  
598 This “Caledonian” event” was marked in the Murzuq basin by large scale uplifts and  
599 secondary highs roughly oriented SSE–NNW in SW Libya, i.e. parallel to the Pan-African  
600 major shear zones in the Djanet and Edembo terranes (Davidson et al., 2000; Ghienne et al.,  
601 2013) constituting the western metacratonic boundary of the Murzuq craton (Fezaa et al.,  
602 2010).

603 This means that the Arrikine sill intruded during a period of reactivations of the mega-  
604 shear zones of the region, known to be former terrane boundaries and as such of lithospheric  
605 scale (Fezaa et al., 2010). The Aïr ring complex province also resulted from the reactivation  
606 of a major shear zone, the Raghane shear zone (Moreau et al., 1994), which corresponds to  
607 the western boundary of the Saharan metacraton (Liégeois et al., 1994; Abdelsalam et al.,  
608 2002). The shift from the rapid vertical ascent along the mega-shear zone of the low-viscosity  
609 Arrikine magma to the widespread horizontal sill development can be tentatively attributed to  
610 the characteristics of the Silurian Acacus formation. The latter is a massive level (about 100 m  
611 thick) overlying lagunal deposits that include gypsum-rich beds (Freulon, 1964). We thus  
612 suggest that the Arrikine magma intruded within evaporitic-rich facies, which, at the contact  
613 with the high temperature magma, became much more easily deformable and moreover  
614 released water due to gypsum transformation, largely blocking the magma ascent. However,  
615 looking at the age of the magmatism and at the stratigraphic series, the Arrikine sill should  
616 have intruded the cover at shallow depth (of the order of less than 300 m), in agreement with  
617 the moderate size (PSD) of the magnetite grains. These renders the presence of Arrikine aerial  
618 products likely but in that case it is likely that they were eroded and transported far away by  
619 the large fluvial system that developed at this period in the Murzuq basin (Ghienne et al.,  
620 2013) as it was the case for most of the Lower Devonian deposits in the Murzuq basin  
621 (Davidson et al., 2000).

622 This kind of emplacement would be favorable to suggest a link between the Arrikine  
623 magmatism and the magmatic rocks proposed to be at the origin of sand injections in the  
624 Murzuq area (Fig. 1A; Moreau et al., 2012). Indeed, these peculiar structures occur in the  
625 Silurian sediments around the Murzuq basin in a similar manner as the Arrikine sill and are  
626 considered as Devonian in age. These sand injections would be the result of the combined  
627 effects of tectonic uplift, igneous intrusions in the sand-rich Cambrian-Ordovician sediments  
628 and overpressures generated by the paleo-relief existing below the Silurian strata (Moreau et  
629 al., 2012). The Arrikine sill would correspond to the magma that intruded the Silurian  
630 sediments, lower Devonian sediments having not deposited in the area due to uplift. Let us  
631 remark that the three known occurrences of sand injections (Mt Telout, In-Ezzane to the west  
632 and El Meherschema to the east; Fig. 1A) are also located around the Murzuq craton, within  
633 its metacratonic margin, the Tibesti area being the equivalent of the Djanet and Edembo  
634 terranes from Eastern Hoggar (Fezaa et al., 2010).

### 635 7.1.2. *The time of rejuvenation, the Mid-Carboniferous*

636 The K-Ar method has given an age of  $325.6 \pm 7.7$  Ma that we consider as close to that  
637 of a rejuvenation event during the Serpukhovian, i.e. from uppermost Viséan to upper  
638 Bashkirian when taking into account the error bracket-attributed to aplitic fluid percolation.  
639 Such an age is also that of the erasure of the primary magnetization in the older sedimentary  
640 rocks of the Saharan platform (Henry et al., 2004). It corresponds thus to the second main  
641 tectonic event in the area. This Mid-Carboniferous compression event led to the major so-  
642 called Hercynian unconformity that affected the upper part of the Paleozoic sequence in the  
643 Murzuq Basin but whose initial phases started during middle Carboniferous (Bellini and  
644 Massa, 1980), or at the Viséan-Serpukhovian boundary (Fröhlich et al., 2010 and references  
645 therein). This event was responsible for the uplift of the Tiririne High (NW Murzuq basin) at  
646 the end of the Viséan, removing the Assedjefar-Marar (Lower Carboniferous) sequence  
647 (Echikh and Sola, 2000). It began with transpressional movements along a series of NNE-  
648 SSW trending wrench faults (parallel to the Murzuq craton boundary), producing a NE-SW en  
649 echelon pattern of folds on the Tiririne High (Echikh and Sola, 2000). At the scale of the  
650 Murzuq basin, the last marine deposits are of Moscovian age (e.g. Massa et al., 1988; Amenna  
651 et al., 2014) and the upper Carboniferous is nearly entirely lacking (due to uplift or to erosion)  
652 and only some red lacustrine mudstone (e.g. upper Carboniferous Tiguentourine Formation)  
653 can be rarely found (Davidson et al., 2000).

654 Such a major compressive event, with transpressional movements and various uplifts  
655 at the regional scale is favorable for fluid movements, able to reset the K-Ar geochronometer  
656 that in turn is able to date it. A Serpukhovian reactivation (“Hercynian”) occurred at the same  
657 place as the early Devonian (“Caledonian”) reactivation. This indicates that emplacement of  
658 the Arrikine sill was related to the lithospheric character of the mega-shear zones of the region  
659 (Fezaa et al., 2010). This implies that these intraplate reactivations are quite important and  
660 must be considered.

661

### 662 7.2 *Global considerations*

663 For a long time, the use of the “Caledonian” and “Hercynian” expressions was  
664 suggested for the intraplate intense tectonic reactivations that occurred around the Murzuq  
665 craton. Causes must be found in events that occurred at the plate margins at these periods. In  
666 addition to tectonic reactivations, volcanism may be generated by adiabatic pressure release of  
667 uprising asthenosphere in response to stress resulting from far field collisions such as, for the  
668 West African Cenozoic volcanism, the Africa-Europe collision (Liégeois et al., 2005). Indeed,

669 these “Caledonian” and “Hercynian” effects are known all over the Saharan platform, to the  
670 north of Hoggar (Haddoum et al., 2001; Galeazzi et al., 2010), around the West African  
671 craton (Michard et al., 2008) or within the Algerian petroleum basins (Galeazzi et al., 2010)  
672 and in the whole Northern Africa (Guiraud et al., 2005). The Alpine reactivation is generally  
673 superimposed on the Hercynian reactivated structures (Smith et al., 2006, Derder et al., 2009).  
674 These “Caledonian” and “Hercynian” tectonic events are also recorded at the regional scale by  
675 detrital zircons in the Murzuq basin (Morton et al., 2011).

676 The Arrikine sill emplacement, as well as that of the Aïr ring-complex province  
677 (Moreau et al., 1994) at c. 410 Ma, occurred during a continental breakup, when the Hun  
678 terranes separated from the northern Gondwana margin (Stampfli and Borel, 2002, Blackey,  
679 2008 and references therein). Such a geodynamical environment is favorable to generate  
680 intraplate tectonics and magmatism as it was the case, for West Africa, during the continental  
681 breakup. This led to the Atlantic ocean opening that generated the Central Atlantic Magmatic  
682 Province (CAMP; Marzoli et al., 1999), which extends in West Africa (e.g. Chabou et al.,  
683 2007, 2010), SW Europe (Callegaro et al., 2013) and even Sicily (Cirrincione et al., 2014),  
684 and East America, including Bolivia (Bertrand et al., 2014). This province, even tholeiitic,  
685 comprises abundant high-Ti basalts related to an asthenospheric mantle source (Deckart et al.,  
686 2005). We thus suggest that the Arrikine sill, and associated magmatism, emplaced due to this  
687 continental breakup that also generated the Caledonian unconformity.

688 The mid-Carboniferous reactivation with an age at  $325.6 \pm 7.7$  Ma as determined by  
689 the K-Ar method is easily related to the late Variscan episode that resulted from the collision  
690 between Gondwana, especially NW Africa, and southern Europe from 330 Ma to early  
691 Permian (Schulmann et al., 2014). The fact that this important event generated no magmas  
692 (except the aplitic metasomatism) can be related to the transpressive character of this  
693 reactivation, unable to allow magmas to rise if no major horizontal movements occurred, as it  
694 is the case in an intraplate setting.

695 To sum up, we propose that the Arrikine sill was emplaced along a major lithospheric  
696 weakness zone, the margin of the Murzuq craton during the Gondwana-Hun continental  
697 breakup. Similarly, the reactivation of the major Raghane shear zone bordering the Saharan  
698 metacraton that allowed the emplacement of the Aïr ring complex province (Moreau et al.,  
699 1994), can be attributed to this continental break-up. This lithosphere weakness zone was  
700 reactivated during the late Variscan orogeny that involved Gondwana. These two events  
701 correspond to the two major unconformities with important sedimentary hiatus known in the  
702 region, the so-called Caledonian and Hercynian unconformities.

703

704 **8. Conclusions**

705

706 The newly discovered Arrikine sill is a large magmatic object (pyroxene-rich gabbro)  
707 of Lower Devonian age emplaced in Silurian sediments, rather unexpected in far-east Hoggar.  
708 Although emplaced in an intraplate environment, its datation was not straightforward: all  
709 zircons are older than the country rocks and are thus inherited (most zircons are at c. 2030 Ma  
710 with additional ages at c. 2700 Ma and in the 766-598 Ma age range) and its rejuvenated K-Ar  
711 age of  $325.6 \pm 7.7$  Ma relates to a remobilization event due to a K-rich aplitic fluid that reset  
712 the isotopic system and that is recorded by secondary minerals (K-feldspar, acid plagioclase,  
713 quartz). By contrast, the paleomagnetic pole is reliable and allows dating of the Arrikine sill  
714 emplacement in the 400-415 Ma age interval (mid-Lochkhovian-mid-Emsian).

715 Geochemically, the Arrikine gabbro has the composition of a high-Ti alkali basalt,  
716 actually a high-Ti hawaiiite (sodic trachybasalt), with 46% SiO<sub>2</sub>, 4% MgO, %Na<sub>2</sub>O-2>%K<sub>2</sub>O,  
717 absence of chromite and 5% TiO<sub>2</sub>. As a whole, this chemistry is close to that of the Mauna  
718 Loa volcano in Hawaii (Wanless et al., 2006), in agreement with an enriched mantle origin  
719 needed for such a high-Ti/low-Si composition. This peculiar chemistry is similar to several  
720 intraplate continental provinces such as the Trias rift-related from SW China, the high-Ti  
721 basalts from the Emeishan LIP (Zhang et al., 2013), the Ordovician Suordakh and Cambrian  
722 Kharaulakh mafic rocks from Siberia (Khudoley et al., 2013) or the 1.5 Ga Yangtse mafic  
723 province (Fan et al., 2013). It is also similar to some lithologies belonging to the lower  
724 Devonian Air ring complexes province in Niger, such as the Abontorok ring complex (Brown  
725 et al., 1989). The peculiarity of these compositions supports so a lower Devonian age for the  
726 Arrikine sill, as is proposed in the paleomagnetic section. The Arrikine gabbro composition  
727 points to a liquidus temperature of 1280°C and a temperature of 1080°C for the magma when  
728 it crystallized close to the surface.

729 Geodynamically, the Arrikine sill emplaced in Silurian sediments belonging to the  
730 western boundary of the Murzuq craton, actually its metacratonic margin (Fezaa et al., 2010;  
731 Liégeois et al., 2013). Its emplacement is related to the reactivation of this metacratonic  
732 margin due to a low stress that propagated in this remote intracontinental region. This was a  
733 consequence of “Caledonian” events that also generated in the area the so-called Caledonian  
734 unconformity, witness of the absence of lower Devonian sediment deposition (e.g. Ghienne et  
735 al., 2013). A link with the magmatic rocks at the origin of the sand injections at the Murzuq  
736 craton margins (Moreau et al., 2012) is suggested. The Air ring complexes were emplaced at

737 the same period, also along a major lithospheric discontinuity, the Raghane shear zone  
738 (Moreau et al., 1994), which is the western boundary of the Saharan metacraton (Abdelsalam  
739 et al., 2002).

740 The Mid-Carboniferous rejuvenation dated by the K-Ar age ( $325.6 \pm 7.7$  Ma) and  
741 marked by a aplitic fluid phase can be correlated with the erasure of the primary  
742 magnetization in the older sedimentary rocks of the Saharan platform (Henry et al., 2004). It  
743 corresponds to the second main Phanerozoic tectonic event in the area marked by the so-  
744 called Hercynian unconformity, responsible for important transpressive vertical movements  
745 such as the uplift of the Tiririne High (NW Murzuq basin).

746 Emplacement (c. 410 Ma) and rejuvenation (c. 326 Ma) of the Arrikine sill are thus  
747 attributed to intense tectonic reactivations that occurred around the Murzuq craton, able to  
748 generate magmas of mantle origin through adiabatic pressure release of uprising  
749 asthenosphere along mega-shear zones. The effects of these two events being known all over  
750 the Saharan platform, it is concluded that they resulted from major events that occurred at  
751 plate boundaries. We suggest that the extensional event in the Arrikine area at 410 Ma is  
752 linked to the continental breakup when the Hun terranes separated from the northern  
753 Gondwana margin (Stampfli and Borel, 2002, Blackey, 2008 and references therein) and, that  
754 the compressional event at c. 325 Ma reflects the far-field effects of the collision between  
755 Africa (belonging to Gondwana) and southern Europe (Schulmann et al., 2014).

756

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1078

1079 **Figure caption**

1080 Figure 1: Geological map of the southwestern Murzuq basin margin: A) Location in the North  
 1081 African context based on Lelubre et al., 1962; Liégeois et al., 2003, 2005; Fezaa et al.,  
 1082 2010. White rectangles indicate the circular structures discussed in the text: Mt T.=  
 1083 Mount Telout; I.E.= In-Ezzane; El Meh.= El Meherschema. B) Geological map of the  
 1084 studied area (modified after Le Caignec et al, 1957, Fezaa et al., 2010 and Geological  
 1085 Map of Libya, 1985, SW sheet).

1086 Figure 2: Detailed map of the Arrikine sill, from image analysis through Landsat 8 satellite  
 1087 images and from field observations and measurements: a) Names on the map concern  
 1088 geographical location of the two main “oueds” in this area. A color composite  
 1089 analysis, using TM6, TM3 and TM1 bands with contrast enhancement, highlights a  
 1090 NW-SE dark band corresponding to magmatism observed and sampled on the field.  
 1091 Contrasting spectral responses allow differentiating the Silurian clays in purple and the  
 1092 Devonian sandstones in brown. b) Names on the map concern geological formations.  
 1093 The relative homogeneity of responses within these different areas, leads us to try to  
 1094 specify the extension and edges of the sill. Four classes have been defined (green  
 1095 clays, marls and limestones; red dark rocks; brown sandstones; white alluvial  
 1096 material). The result is a classified image by maximum likelihood method. The  
 1097 Cambro-Ordovician levels are not differentiated by this classification and the  
 1098 magmatism in red is well underlined with its NW-SE orientation. However, the  
 1099 northern part of the red area shows a surprising lateral eastward extension.  
 1100 Comparison with the figure 2a rather suggests drainage of dark material by the Oued  
 1101 Arrikine. The paleomagnetic sampling sites are indicated with numbers by yellow  
 1102 squares (gabbro) and black dots (sedimentary formations). Cambro-Ordovician (ks),  
 1103 Silurian (s) with its two formations (Oued In Djerane and Acacus) and Devonian (d).  
 1104 Yellow circles correspond to injectites.

- 1105 Figure 3: U-Pb concordia diagram.
- 1106 Figure 4: Thermal demagnetization curves for representative samples of the dark facies.
- 1107 Figure 5: Typical thermomagnetic curves (normalized low field susceptibility  $K/K_0$  as a  
1108 function of the temperature  $T$  - in  $^{\circ}\text{C}$ ) for representative samples of the gray (IZ716)  
1109 and dark (IZ696) facies of the gabbro.
- 1110 Figure 6: a, b): Hysteresis loops (a: sample IZ725 and b: sample IZ678, both for dark facies);  
1111  $H_{cr}$  is the remanent coercive force, Field  $H$  in Tesla. c) Day plot (Day et al. 1977) of  
1112 the hysteresis ratios; PSD (pseudo-single domain) and MD (multidomain) areas for  
1113 magnetite.
- 1114 Figure 7: Orthogonal vector plots (filled circles: horizontal plane, crosses: vertical plane), in  
1115 stratigraphic coordinates for AF (a, c and e), thermal (b, d, f) and mixed (g) treatments  
1116 for samples IZ659A (a), IZ664A (b), IZ709A (c), IZ739A (d), IZ711A and B (e, f) and  
1117 IZ660A (g).
- 1118 Figure 8: Equal-area plot after dip correction (a) of ChRMs  $C$  and (b) of the mean-site  
1119 paleomagnetic directions with associated 95% confidence zone (open circles: negative  
1120 inclinations).
- 1121 Figure 9: In North-West Africa coordinates, comparison of the Arrikine pole with Lower  
1122 Devonian poles (see text), with the 560-210 Ma Gondwana APWP (in red 560-250 Ma  
1123 - Amenna et al., 2014, Henry 2015 – in orange 240-210 Ma - Domeier et al., 2012),  
1124 the Africa APWP (in green 200-0 Ma - Besse and Courtillot, 2002) and in blue the  
1125 430-380 Ma Gondwana APWP of Torsvick et al. (2012). Paleomagnetic poles with  
1126 associated confidence zone of Mt. Leyshon Devonian dykes (Mt. LDD) in Australia  
1127 (Clarck, 1996), Snowy River Volcanics (SRV) in Australia (Schmidt et al., 1987),  
1128 Herrada Member, Sierra Grande (HM) in Argentina (Rapalini and Vilas, 1991), Aïr  
1129 intrusives (Aïr) in Niger (Hargraves et al., 1987), Tin Serririne (TS – Derder et al.,  
1130 2006) and Arrikine (ARK) in Algeria (this study). Confidence zones are also presented  
1131 for the 250 or 500 Ma poles of the APWPs.
- 1132 Figure 10: Arrikine sill representative sample AF-ALG/Dj-Site58 compared with rocks  
1133 having also a high-Ti basaltic composition (Aïr, Niger (Abontorok ring-complex):  
1134 Brown et al., 1989; Siberia: Khudoley et al., 2013, Virunga: Rogers et al., 1998,  
1135 Hawaii: Wanless et al., 2006, Yangtse: Fan et al., 2013, SW China: Zhang et al.,  
1136 2013). (A) REE normalized to chondrites (Taylor and McLennan, 1985), (B) Dj 58  
1137 spectrum for incompatible elements normalized to MORB (Sun, 1980 and Pearce,



1138 1982) compared with high-Ti basaltic composition of rocks of Siberia, Virunga,  
1139 Hawaii, Yangtse and SW China (see references above)

1140 Figure 11: Arrikine gabbro AF-ALG/Dj-Site58 spectrum for incompatible elements  
1141 normalized to the Primitive Mantle (Sun and McDonough, 1989) compared with  
1142 Siberian high-Ti basalts (Suordakh event, Eastern margin of the Siberian craton;  
1143 Khudoley et al., 2013).

1144

1145 Table 1: Microprobe analysis

1146 Table 2: K-Ar dating of the Arrikine sill.

1147 Table 3: U-Th-Pb laser ablation ICP-MS analyses for zircons extracted from the Arrikine  
1148 gabbro.

1149 Table 4: Mean direction: Site, number of ChRM N, Declination D and Inclination I before  
1150 and after dip correction, Fisher (1953) parameters  $k$  and  $\alpha_{95}$ , and VGP or  
1151 paleomagnetic pole: Lat. ( $^{\circ}$ S), Long ( $^{\circ}$ E), with for the latter corresponding Fisher  
1152 parameters  $K$  and  $A_{95}$ .

1153 Table 5: Major and trace elements for the Arrikine sill measured on a Horiba-Jobin-Yvon  
1154 Ultima 2 ICP-AES at the IUEM, Pôle de Spectrométrie Océan, Brest, France, by  
1155 Céline Liorzou. CIPW norm and other parameters have been calculated using Norm4  
1156 Excel spreadsheet from Kurt Hollocher:  
1157 ([http://minerva.union.edu/hollochk/c\\_petrology/norms.htm](http://minerva.union.edu/hollochk/c_petrology/norms.htm)).

Mineral		SiO <sub>2</sub>	TiO <sub>2</sub>	Al <sub>2</sub> O <sub>3</sub>	FeO	MnO	MgO	CaO	Na <sub>2</sub> O	K <sub>2</sub> O	P <sub>2</sub> O <sub>5</sub>	Cr <sub>2</sub> O <sub>3</sub>	NiO	BaO	Total
<b>Magmatic minerals</b>															
Plagioclase	58 A-8	54.43	0.150	28.51	0.316	0.000	0.056	11.33	4.933	0.328	0.060	0.000	0.000	0.081	100.19
Pyroxène	58 B-2	50.76	1.027	1.252	12.818	0.246	11.16	21.99	0.506	0.014	0.000	0.002	0.000	0.025	99.81
Biotite	58 B-1	34.86	6.895	12.86	24.52	0.162	7.943	0.000	0.441	9.051	0.011	0.000	0.000	0.000	96.74
Biotite	58 A-5	34.99	7.247	13.53	21.42	0.087	9.491	0.040	0.688	9.068	0.000	0.000	0.000	0.000	96.57
Apatite	58 A-4	0.473	0.000	0.068	0.496	0.048	0.112	54.80	0.003	0.000	42.49	0.000	0.000	0.043	98.53
Opaque	58 A-6	0.005	52.51	0.069	45.22	0.684	1.636	0.051	0.000	0.000	0.000	0.000	0.002	0.000	100.17
Opaque	58 A-7	0.070	26.62	1.269	67.61	1.398	0.014	0.049	0.008	0.006	0.010	0.000	0.001	0.000	97.06
<b>Aplitic secondary minerals</b>															
Hyper-K spot	58 B-1	64.33	0.000	18.42	0.039	0.003	0.051	0.097	0.837	15.57	0.005	0.000	0.008	0.260	99.61
Hyper -K-Ba spot	58 B-1	61.97	0.073	20.09	0.201	0.000	0.003	0.832	2.837	9.821	0.000	0.000	0.000	3.481	99.30
High-K spot	58 B-1	65.10	0.125	21.12	0.172	0.015	0.003	2.206	8.490	2.801	0.000	0.000	0.000	0.098	100.13
<b>Altered Fe-Mg minerals</b>															
Altered spot	58 A-1	37.97	0.146	2.455	23.85	0.195	6.631	10.80	0.139	0.206	3.912	0.000	0.000	0.000	86.30
Altered spot (brown)	58 A-2	31.03	0.025	12.26	29.37	0.179	14.40	0.340	0.029	0.029	0.000	0.000	0.013	0.000	87.67
Altered spot (green)	58 A-3	31.86	0.057	11.26	30.80	0.129	13.52	0.447	0.105	0.045	0.000	0.000	0.000	0.034	88.25

Sample code	Lab. anal. ref.	Age (Ma)	±	error at ± 1 $\sigma$	K <sub>2</sub> O wt. %	<sup>40</sup> Ar <sub>R</sub> (10 <sup>-5</sup> cm <sup>3</sup> g <sup>-1</sup> )	<sup>40</sup> Ar <sub>R</sub> (%)	Fused weight (g)	<sup>36</sup> Ar <sub>exp</sub> (10 <sup>-10</sup> cm <sup>3</sup> )
AF-ALG/Dj-Ste58	B 7398-1	327.2	±	7.6	1.40	1.619	85.3	0.4042	38.2
	B 7399-2	314.2	±	7.3		1.549	86.8	0.3008	23.4
	B 7437-8	333.5	±	7.7		1.654	86.7	0.0580	5.00
	B 7459-11	327.9	±	7.6		1.623	83.4	0.0729	7.96
	<b>Mean age</b>	<b>325.6</b>	<b>±</b>	<b>7.7</b>					

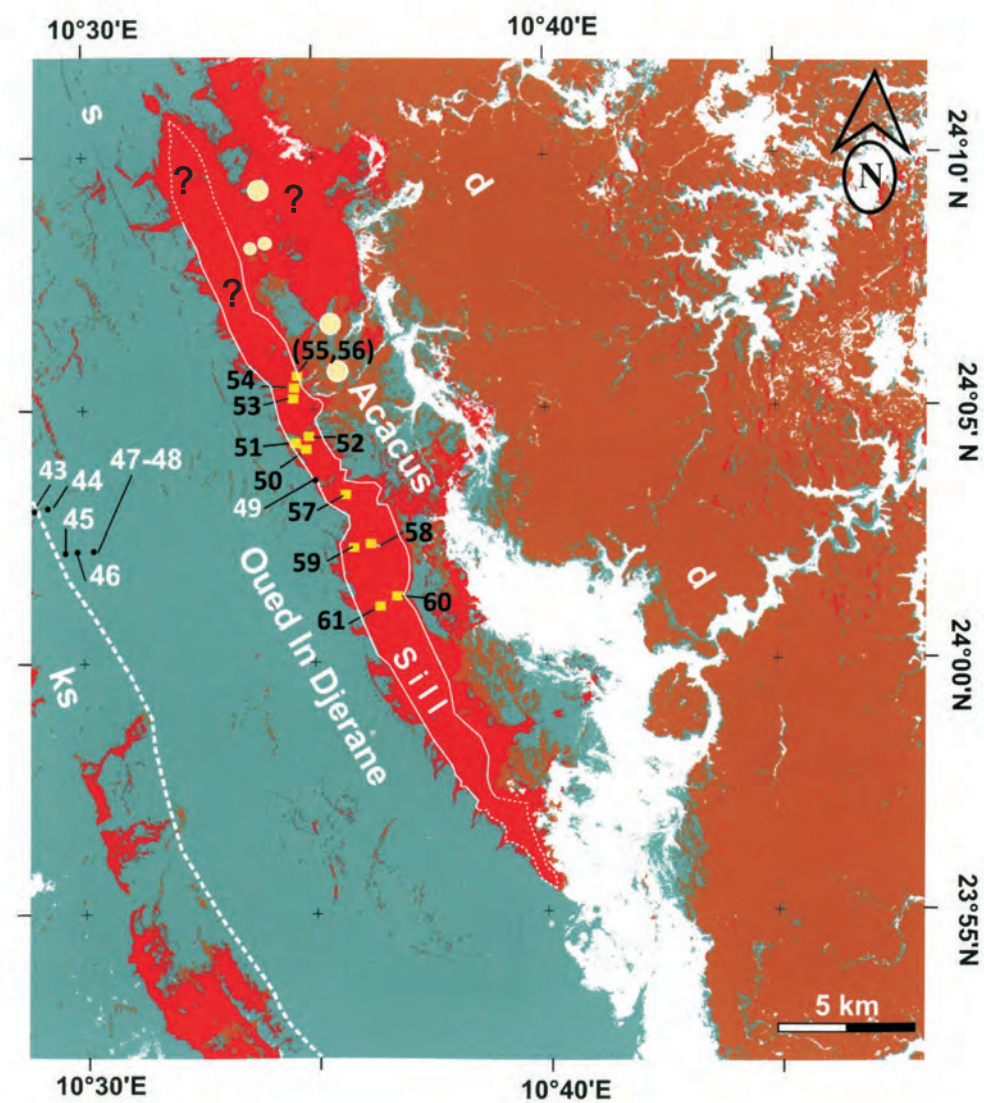
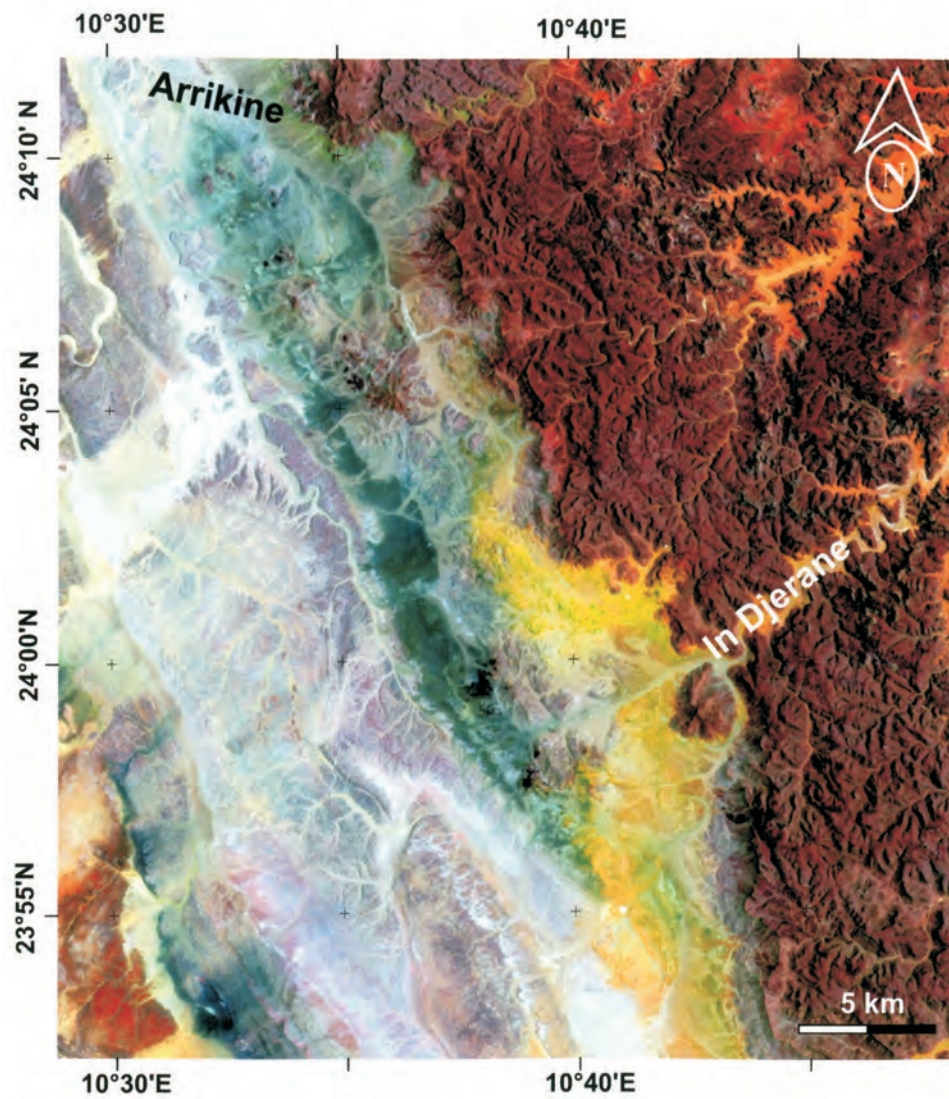
Sample	Pb* (ppm)	Th (ppm)	U (ppm)	Th/ U	<sup>208</sup> Pb/ <sup>206</sup> Pb	<sup>207</sup> Pb/ <sup>206</sup> Pb	± (1σ)	<sup>207</sup> Pb/ <sup>235</sup> U	± (1σ)	<sup>206</sup> Pb/ <sup>238</sup> U	± (1σ)	Rho	Apparent age (Ma)				Conc (%)
													<sup>206</sup> Pb/ <sup>238</sup> U	± (1σ)	<sup>207</sup> Pb/ <sup>206</sup> Pb	± (1σ)	
Arrikine gabbro																	
#1-1	87	82	142	0.58	0.16	0.18473	0.00207	13.247	0.112	0.5197	0.0041	0.94	2698	18	2696	18	100.1
#2-1	44	17	421	0.04	0.02	0.06024	0.00094	0.856	0.011	0.1030	0.0009	0.64	632	5	612	34	103.2
#2-2	25	7	247	0.03	0.02	0.12609	0.00177	6.777	0.078	0.3895	0.0037	0.82	2121	17	2044	25	103.7
#3-1	219	112	523	0.21	0.07	0.18420	0.00221	13.218	0.124	0.5201	0.0045	0.92	2700	19	2691	20	100.3
#4-1	134	151	213	0.71	0.19	0.12365	0.00147	6.635	0.061	0.3889	0.0032	0.88	2118	15	2010	21	105.4
#5-1	664	113	1037	0.11	0.02	0.18601	0.00217	13.412	0.121	0.5227	0.0044	0.93	2711	19	2707	19	100.1
#6-1	96	118	155	0.77	0.22	0.12506	0.00137	6.572	0.054	0.3809	0.0029	0.93	2081	13	2030	19	102.5
#7-1	308	84	804	0.10	0.04	0.06124	0.00189	1.066	0.030	0.1262	0.0017	0.46	766	10	648	65	118.3
#8-1	8	19	62	0.31	0.10	0.12485	0.00138	6.397	0.053	0.3715	0.0028	0.92	2036	13	2027	20	100.5
#9-1	12	56	103	0.55	0.16	0.06410	0.00108	0.946	0.014	0.1070	0.0010	0.63	656	6	745	35	88.0
#10-1	16	102	130	0.78	0.25	0.06221	0.00149	0.945	0.021	0.1102	0.0012	0.50	674	7	681	50	98.9
#11-1	10	38	91	0.41	0.14	0.06207	0.00129	0.955	0.018	0.1116	0.0011	0.54	682	6	676	44	100.9
#12-1	22	143	177	0.81	0.25	0.12459	0.00135	6.088	0.049	0.3543	0.0027	0.94	1955	13	2023	19	96.7
#13-1	648	127	1628	0.08	0.02	0.06548	0.00159	1.004	0.022	0.1112	0.0012	0.50	680	7	790	50	86.1
#13-2	338	72	1007	0.07	0.02	0.06236	0.00116	0.970	0.016	0.1129	0.0011	0.57	690	6	686	39	100.5
#14-1	7	47	68	0.69	0.21	0.05945	0.00195	0.796	0.024	0.0972	0.0012	0.42	598	7	584	70	102.4
#15-1	106	58	327	0.18	0.06	0.13819	0.00152	6.015	0.049	0.3158	0.0024	0.93	1769	12	2205	19	80.2
#16-1	466	204	1162	0.18	0.05	0.12585	0.00135	6.459	0.051	0.3724	0.0028	0.94	2041	13	2041	19	100.0
#17-1	567	210	1533	0.14	0.05	0.12481	0.00132	6.358	0.049	0.3696	0.0027	0.95	2027	13	2026	19	100.1

Table 3: U-Th-Pb laser ablation ICP-MS analyses for zircons extracted from the Arrikine gabbro

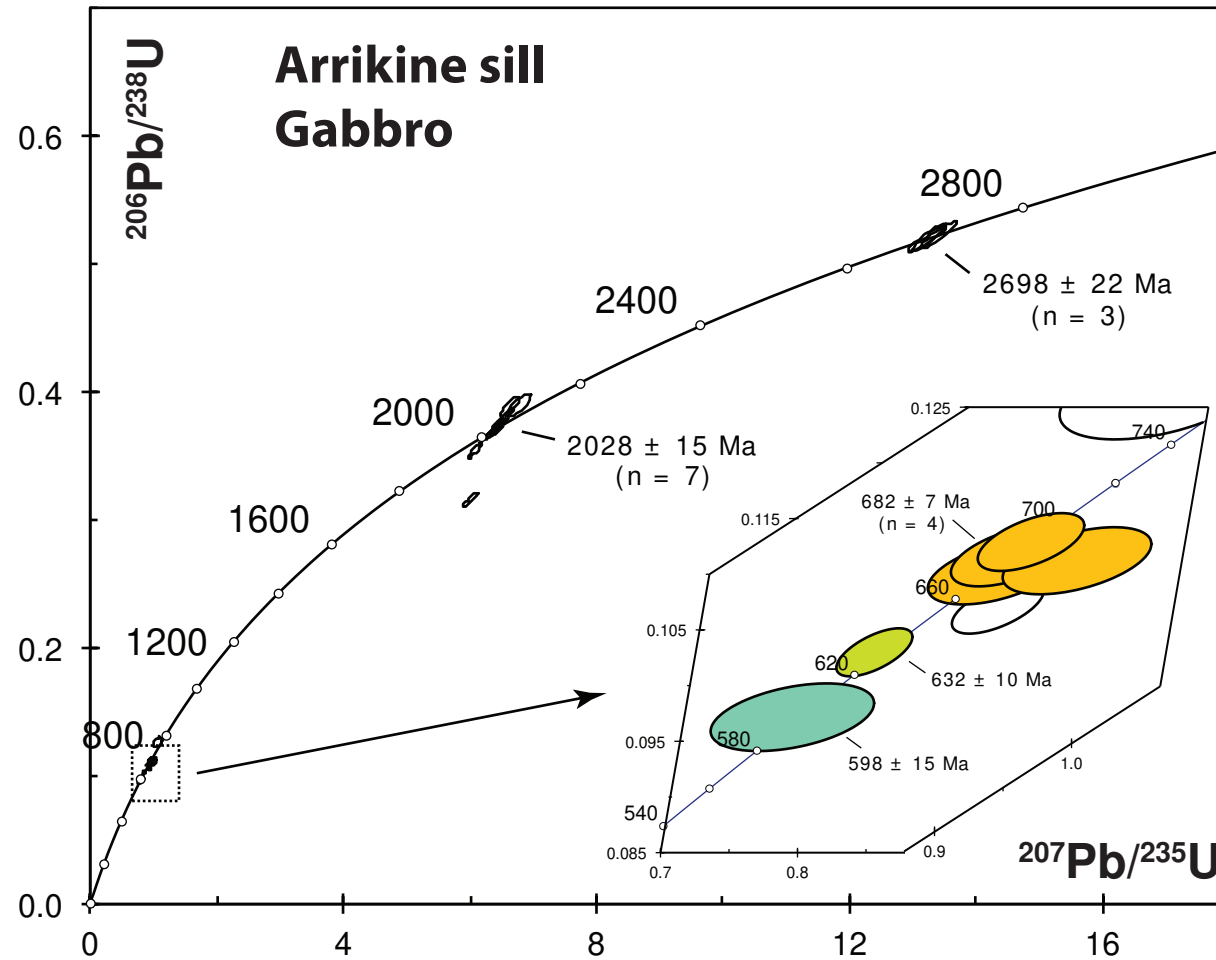
		Before and after dip correction						<b><i>PGVs and Paleomagnetic pole</i></b>			
Site	N	D(°)	I(°)	D(°)	I(°)	k	$\alpha_{95}$ (°)	Lat(°S)	Long(°E)	K	A <sub>95</sub> (°)
50	6	357.4	-24.7	354.6	-25.2	40	10.8	<b>52.4</b>	<b>19.1</b>		
51	3	342.6	-47.4	336.3	-46.3	416	6.1	<b>33.5</b>	<b>35.8</b>		
52	6	10.7	-45.7	4.6	-47.4	27	13.1	<b>37.3</b>	<b>5.4</b>		
53	9	357.8	-37.1	353.2	-37.5	109	4.9	<b>44.5</b>	<b>19.4</b>		
54	9	349.8	-36.2	345.4	-35.8	118	4.8	<b>43.9</b>	<b>29.7</b>		
55	2	-	-	-	-	-	-	-	-		
56	1	-	-	-	-	-	-	-	-		
57	10	355.8	-36.1	352.9	-36.5	66	6.0	<b>45.2</b>	<b>20.0</b>		
58	6	347.9	-35.2	343.7	-34.7	72	7.9	<b>44.1</b>	<b>32.2</b>		
59	7	2.7	-45.5	356.6	-46.4	44	9.2	<b>38.2</b>	<b>14.3</b>		
60	12	353.7	-36.0	349.3	-36.0	115	4.1	<b>44.8</b>	<b>24.7</b>		
61	13	351.1	-41.6	345.8	-41.3	21	9.4	<b>40.3</b>	<b>27.6</b>		
<b>Mean (specimens)</b>	84	355.4	-38.4	350.7	-38.6	38	2.5				
<b>Mean (Sites)</b>	10	355.5	-38.8	350.6	-38.9	86	5.2	<b>42.8</b>	<b>22.9</b>	<b>88</b>	<b>4.7</b>

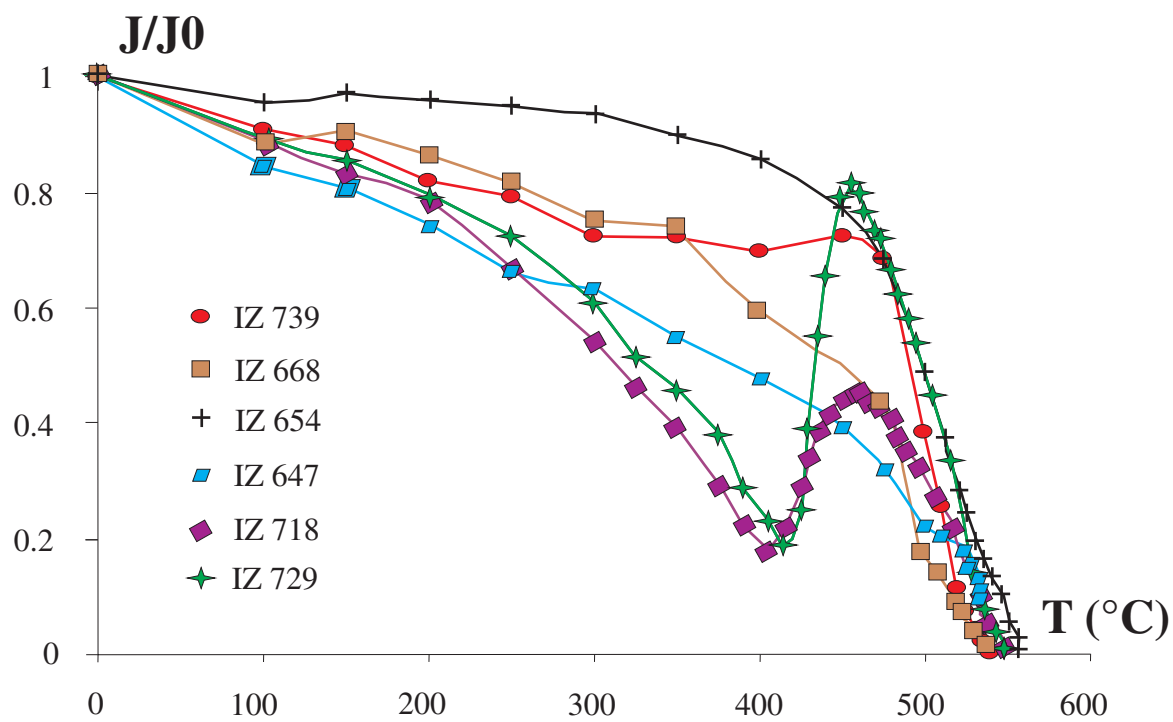
<b>Sample AF-ALG / Dj 58</b>						
<b>Oxide</b>	<b>a %</b>	<b>b %</b>	<b>Element</b>	<b>ppm</b>	<b>Normative minerals</b>	<b>Vol %</b>
SiO <sub>2</sub>	43.80	45.99	Rb	27.2	Plagioclase	60.47
TiO <sub>2</sub>	4.77	5.01	Sr	442	Orthoclase	11.05
Al <sub>2</sub> O <sub>3</sub>	13.97	14.66	Ba	1247	Diopside	8.16
Fe <sub>2</sub> O <sub>3</sub>	15.94	16.73	Sc	19.3	Hypersthene	1.13
MnO	0.19	0.20	V	171	Olivine	5.64
MgO	3.90	4.10	Cr	3	Ilmenite	6.13
CaO	6.28	6.60	Co	46	Magnetite	5.36
Na <sub>2</sub> O	4.12	4.32	Ni	39	Apatite	2.02
K <sub>2</sub> O	1.40	1.47	Y	38.6	Zircon	0.04
P <sub>2</sub> O <sub>5</sub>	0.87	0.91	Zr	320	Plagioclase An content	29.15
			Nb	32.7		
LOI	1.84	1.84	La	37.7	Differentiation Index	62.31
Total	97.11	97.11	Ce	84.0		
			Nd	48.9		
			Sm	9.9	Calculated density, g/cc	3.03
<b>a</b> Analysis			Eu	3.3	Calculated liquid density, g/cc	2.70
<b>b</b> Analysis recalculated			Gd	9.6	Calculated viscosity, dry, Pas	0.17
to 100% on an anhydrous basis			Dy	7.7	Calculated viscosity, wet, Pas	0.16
			Er	3.8	Estimated liquidus temp., °C	1280
			Yb	3.3	Estimated H <sub>2</sub> O content, wt, %	0.20
			Th	2.3		

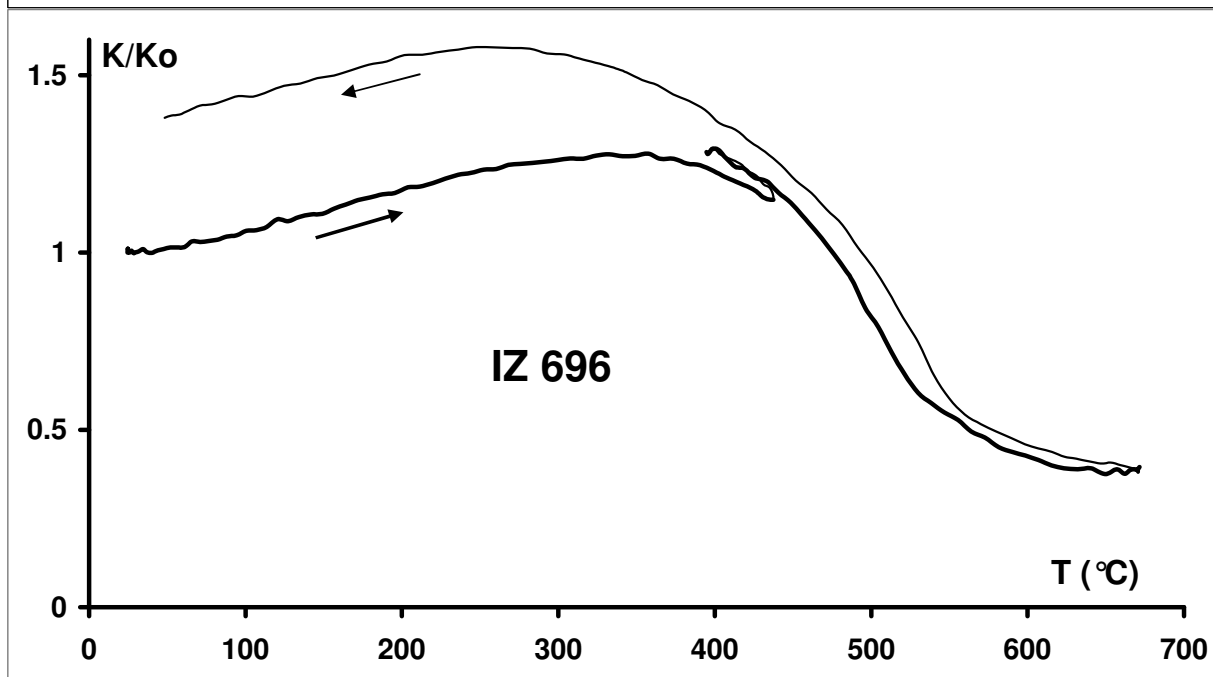
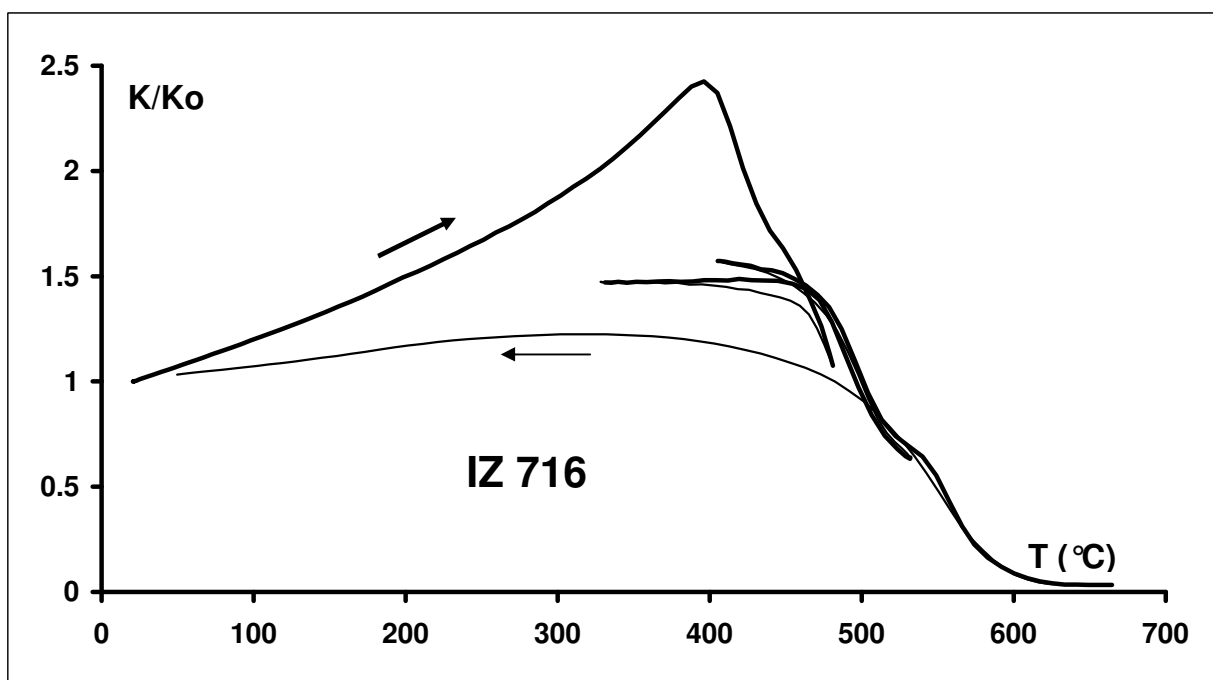


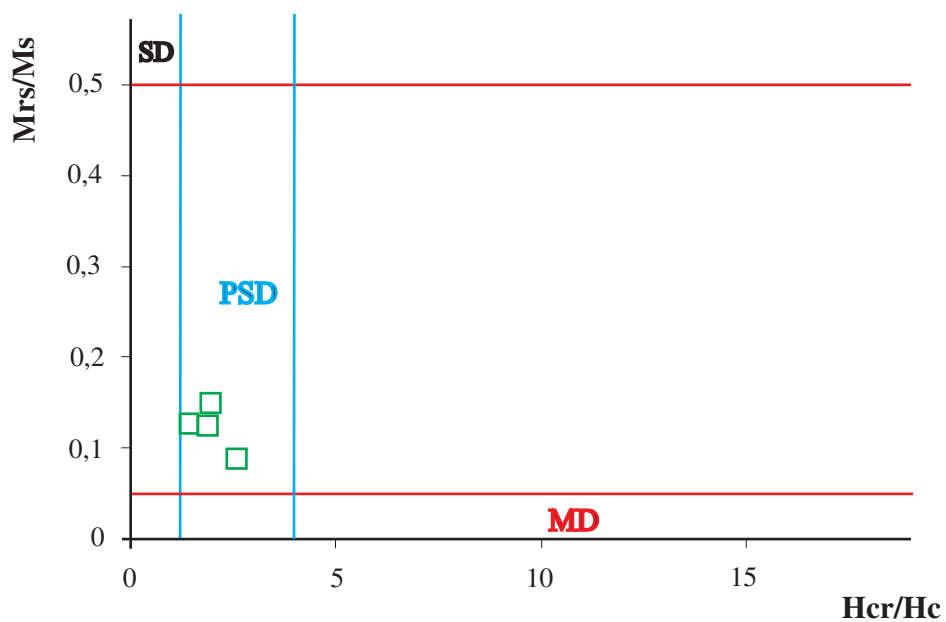
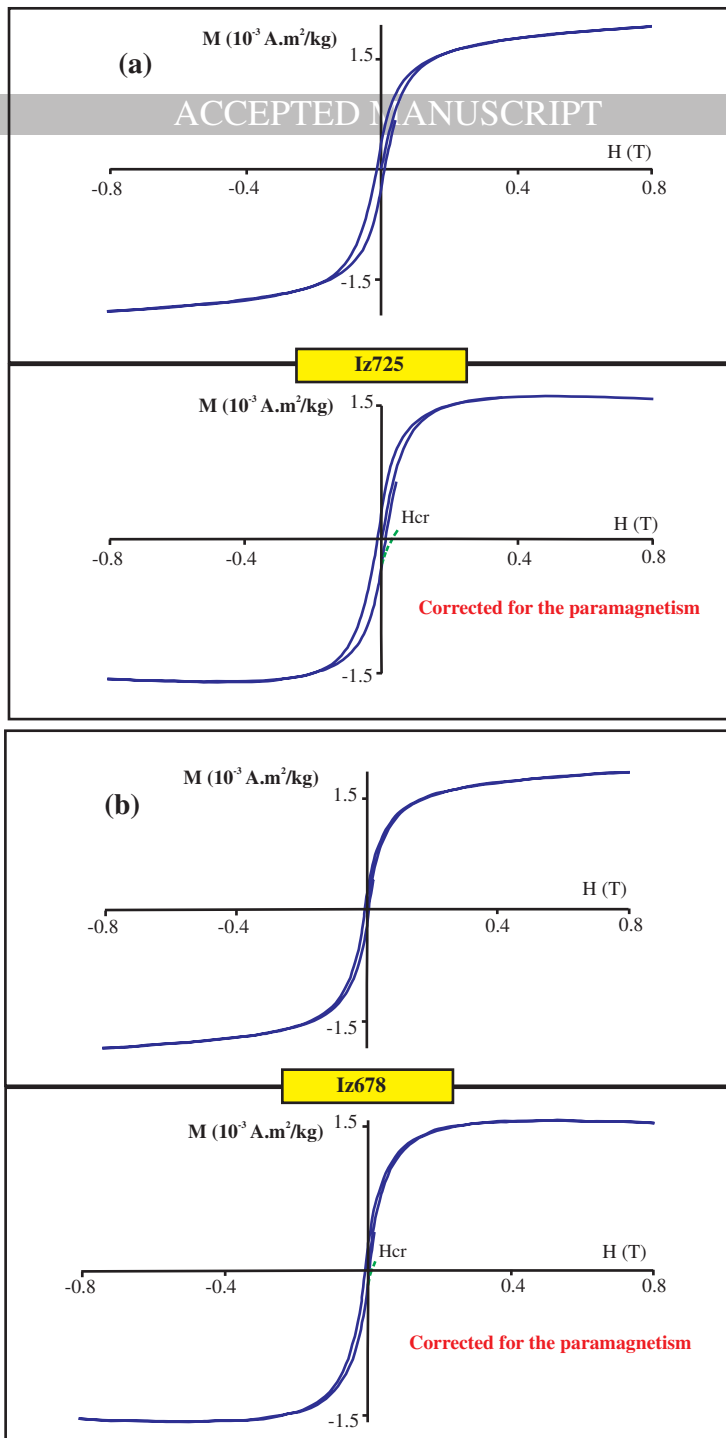


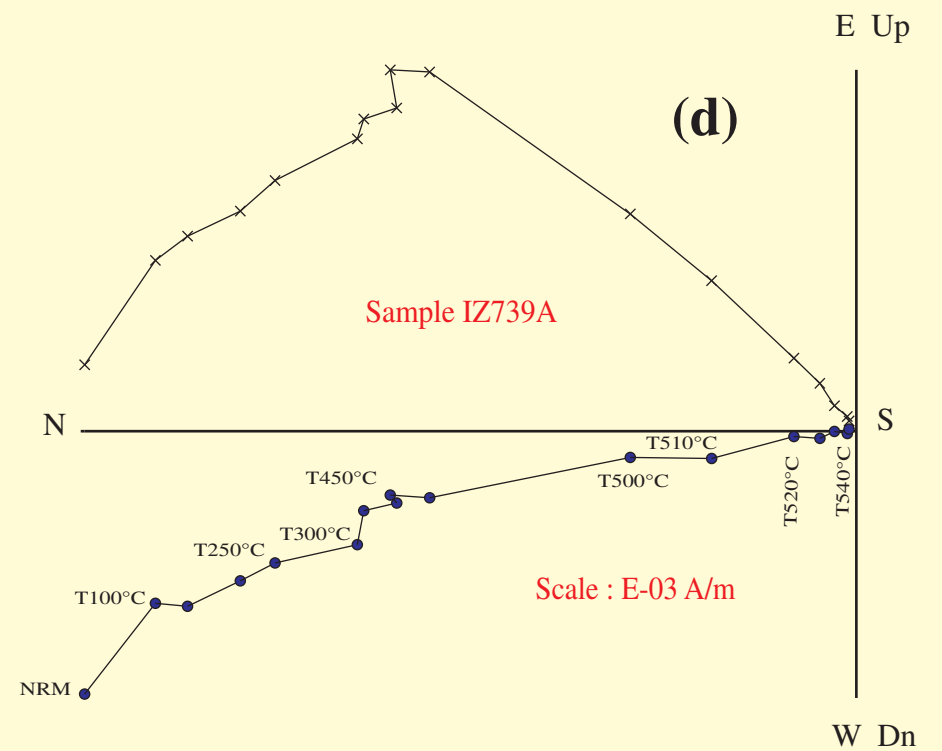
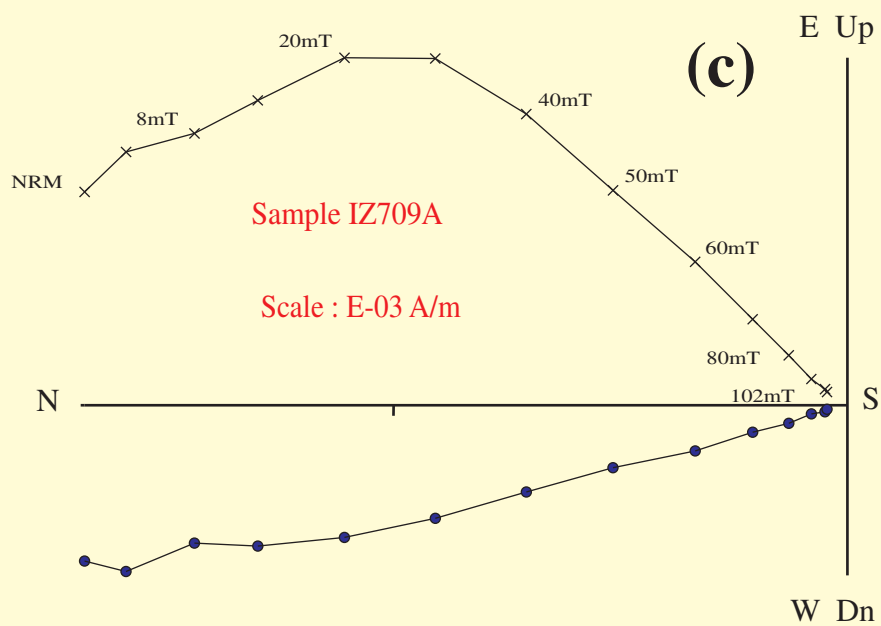
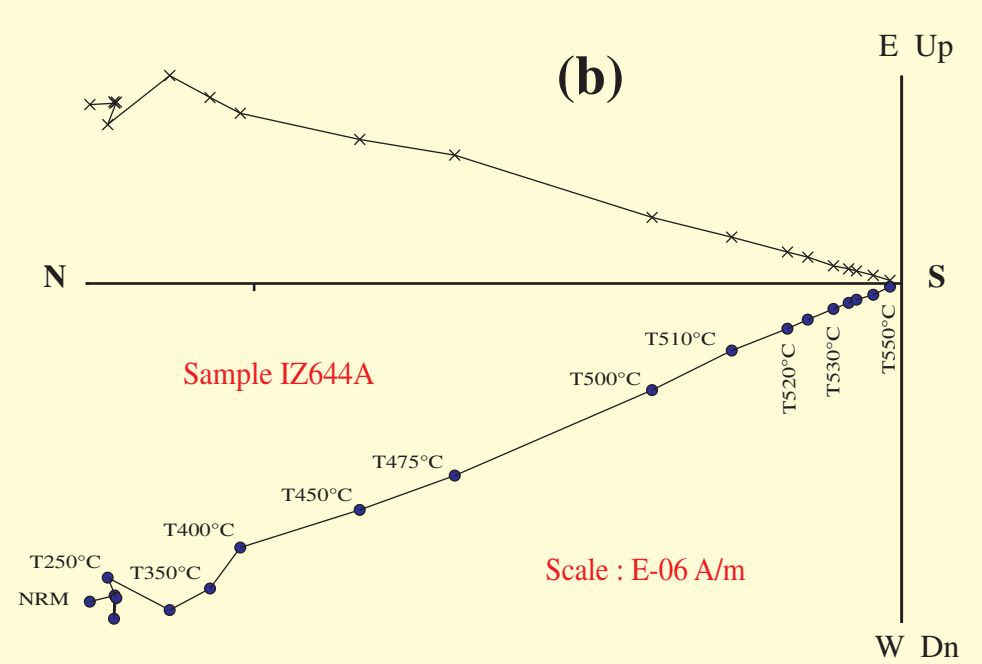
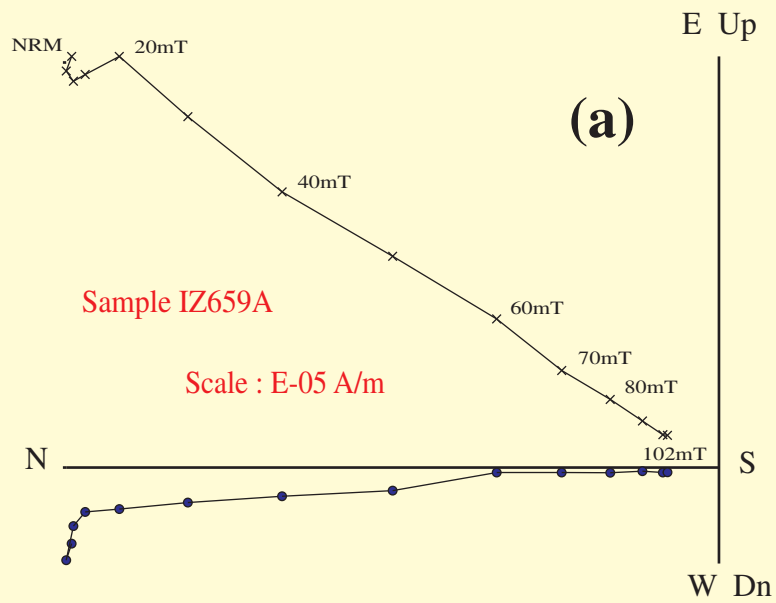


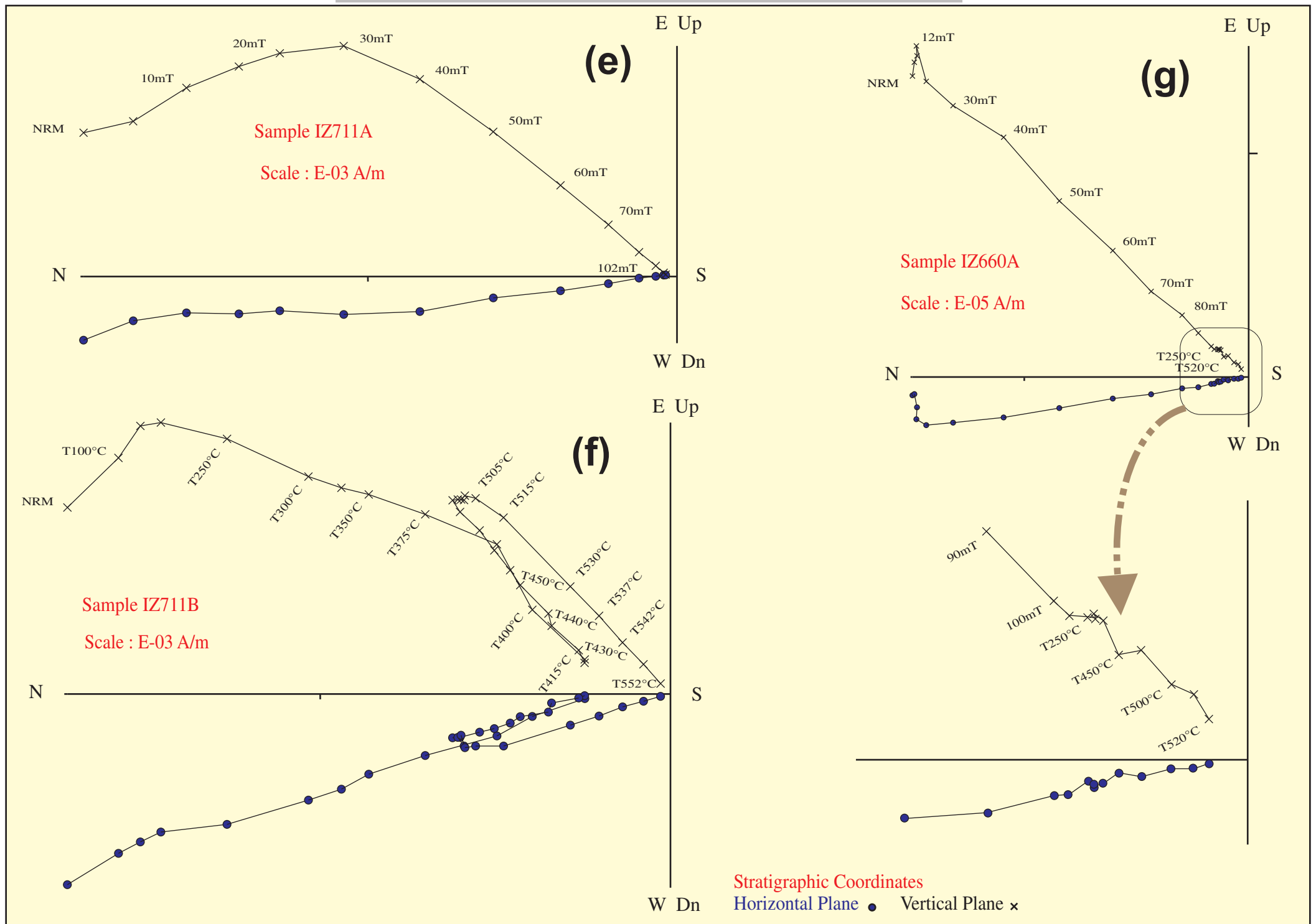


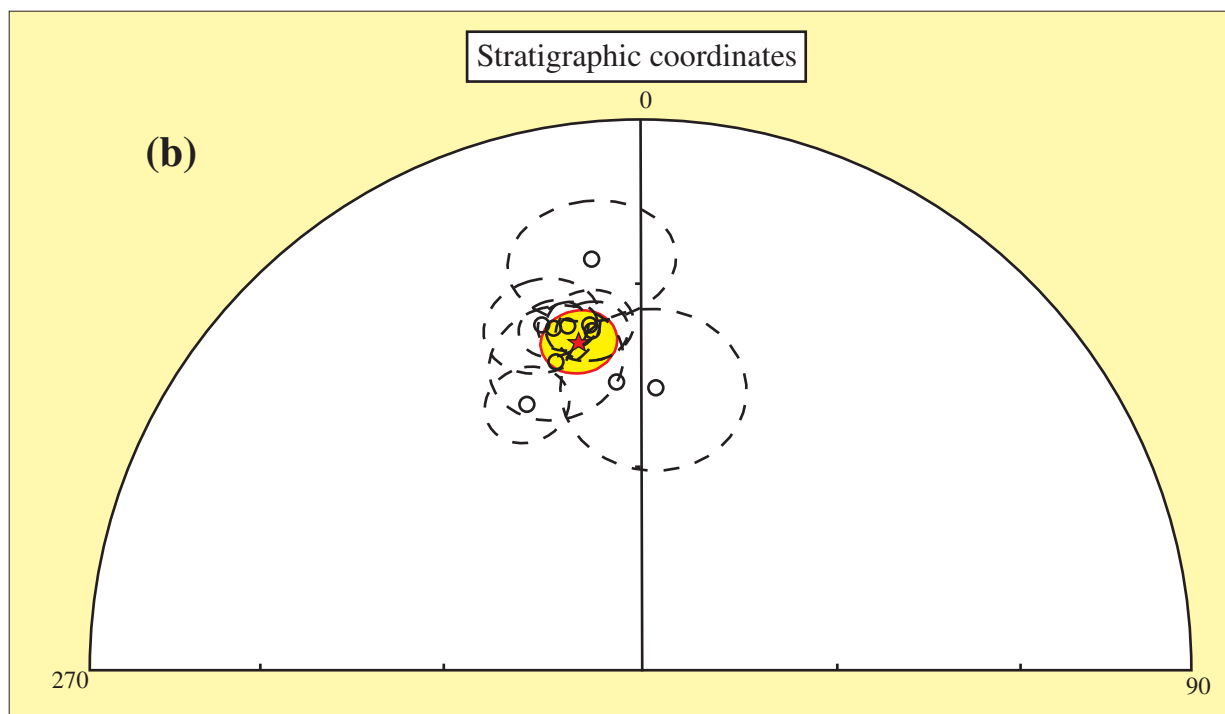
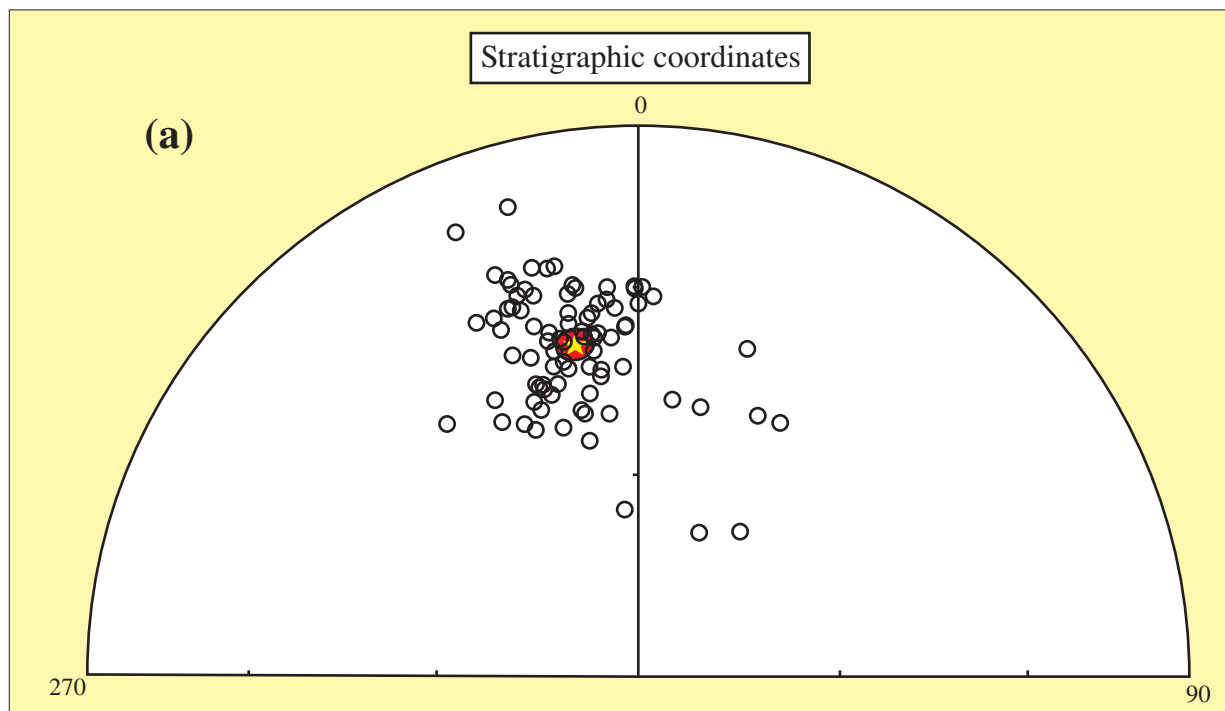






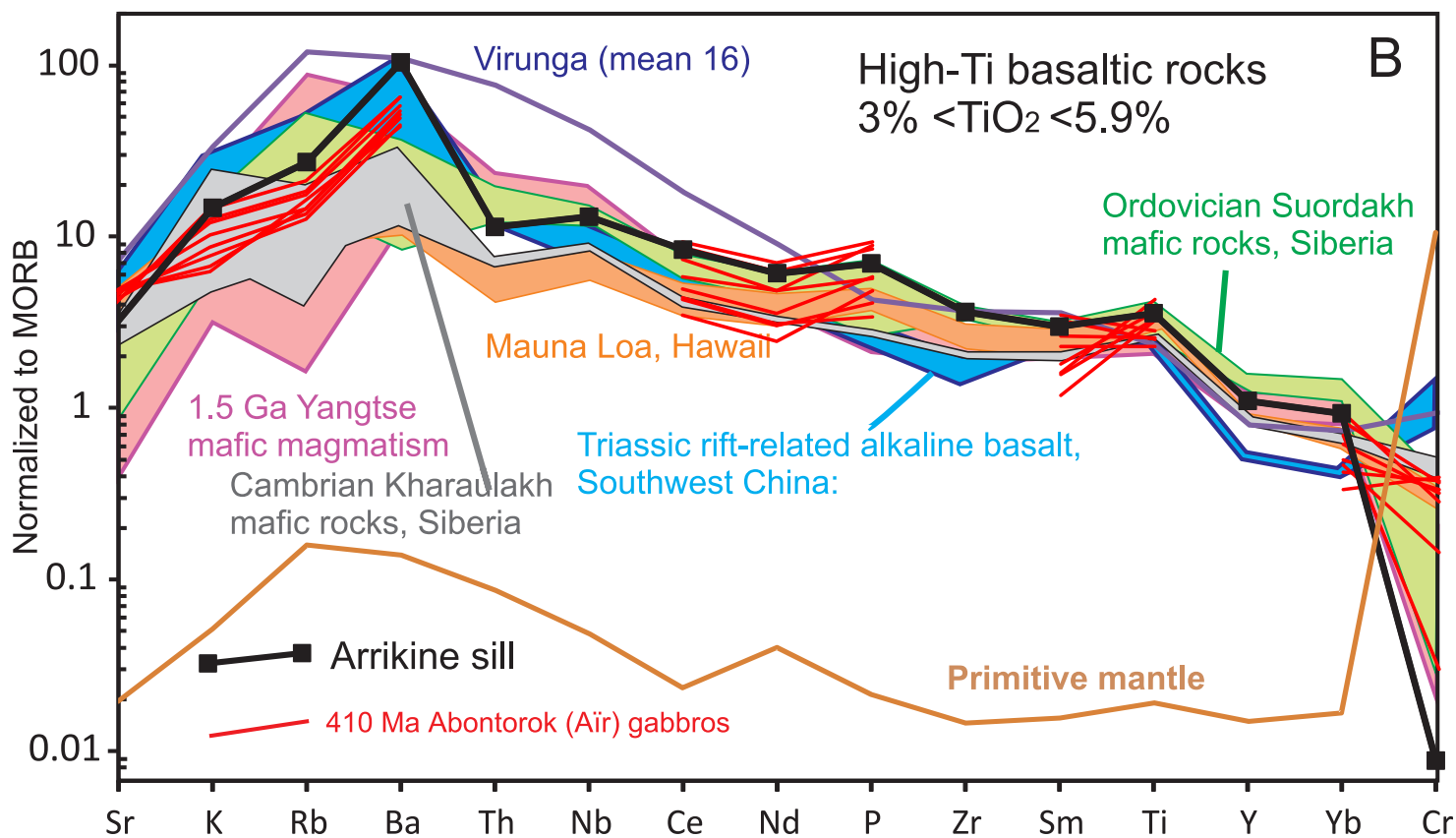
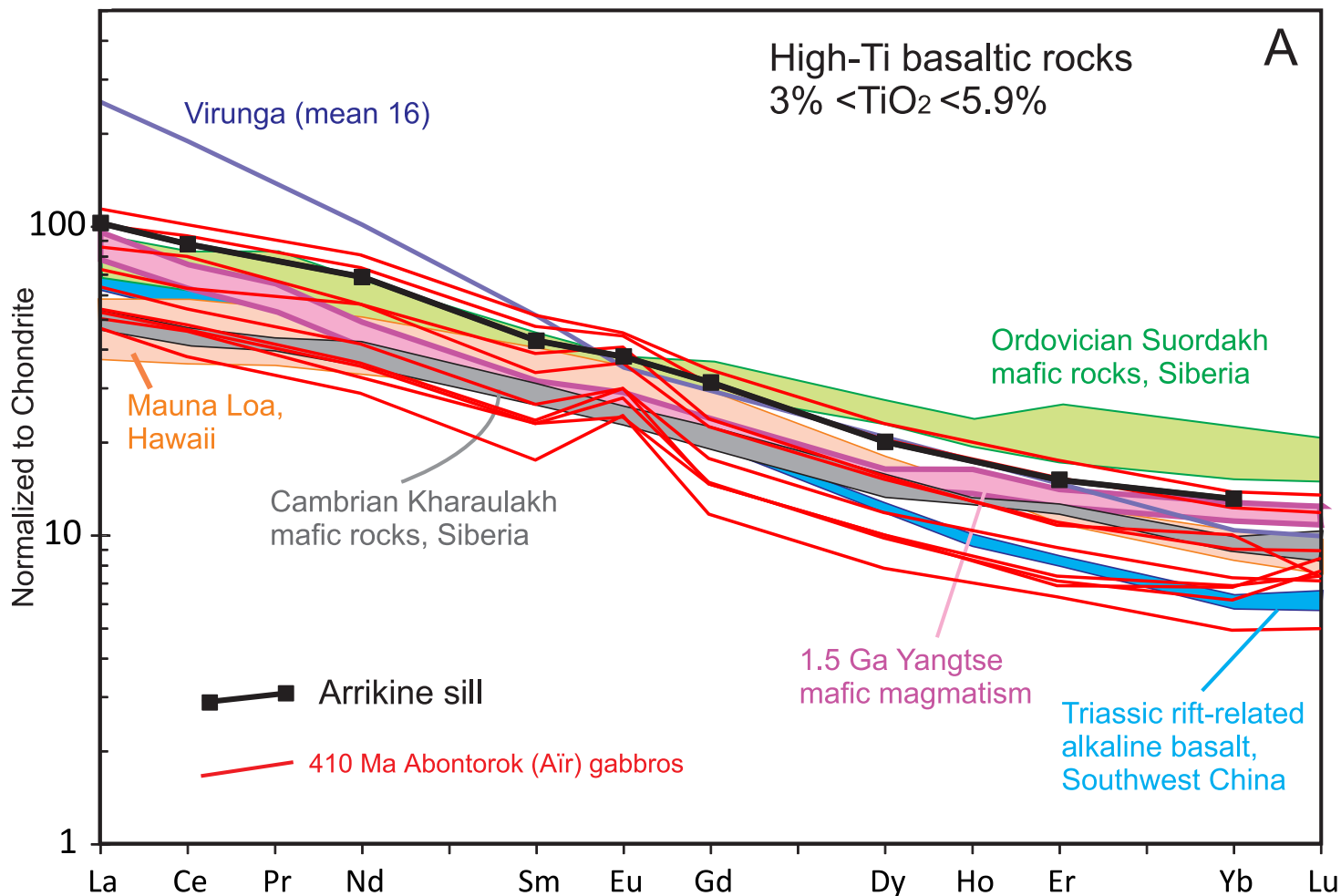


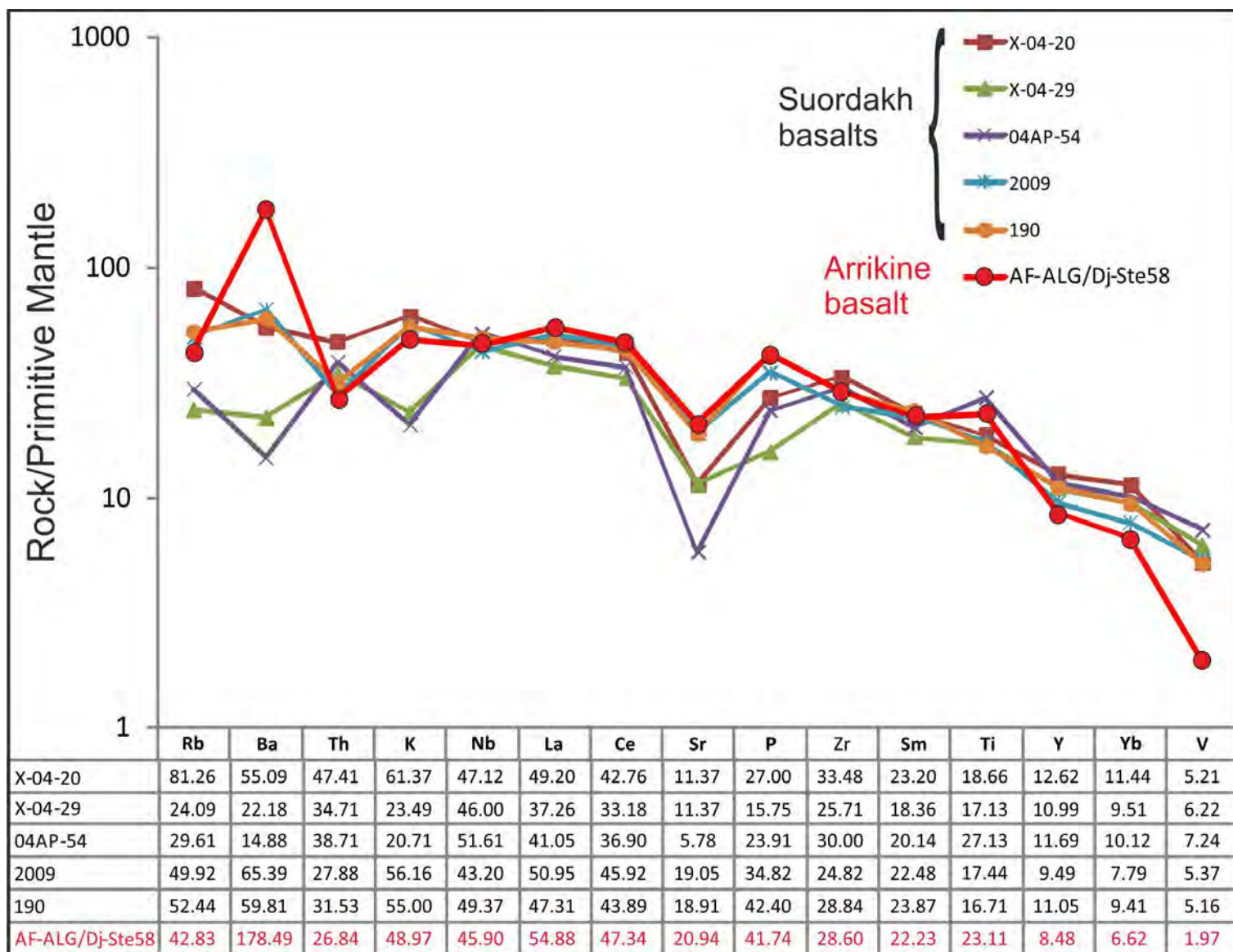


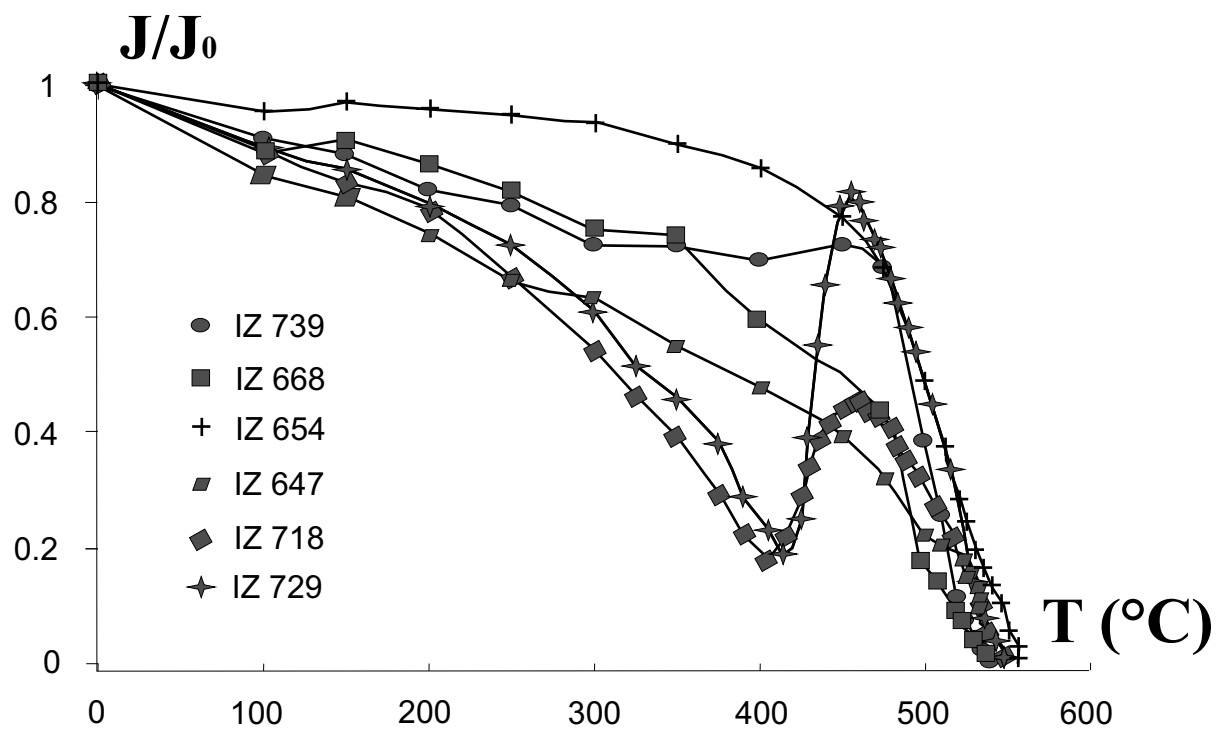


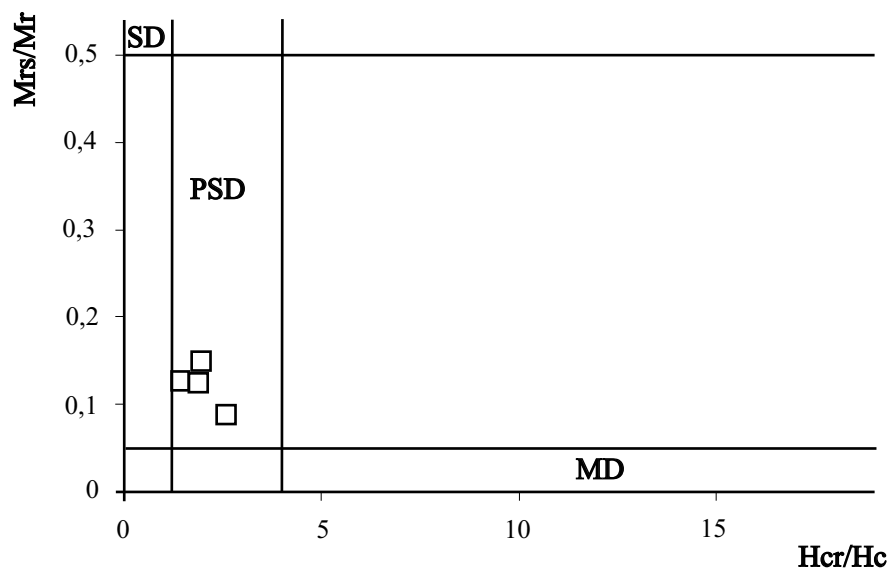
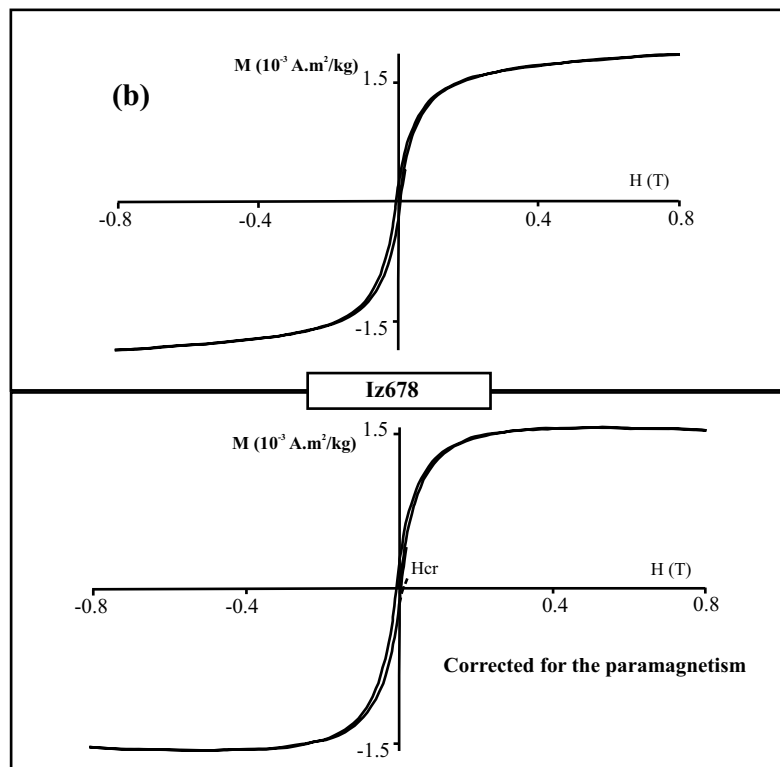
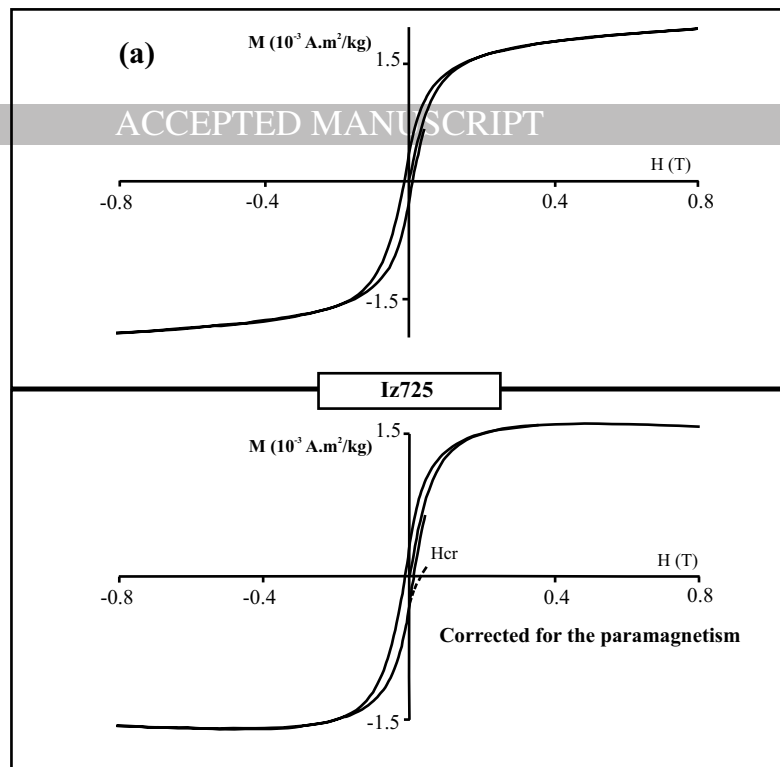


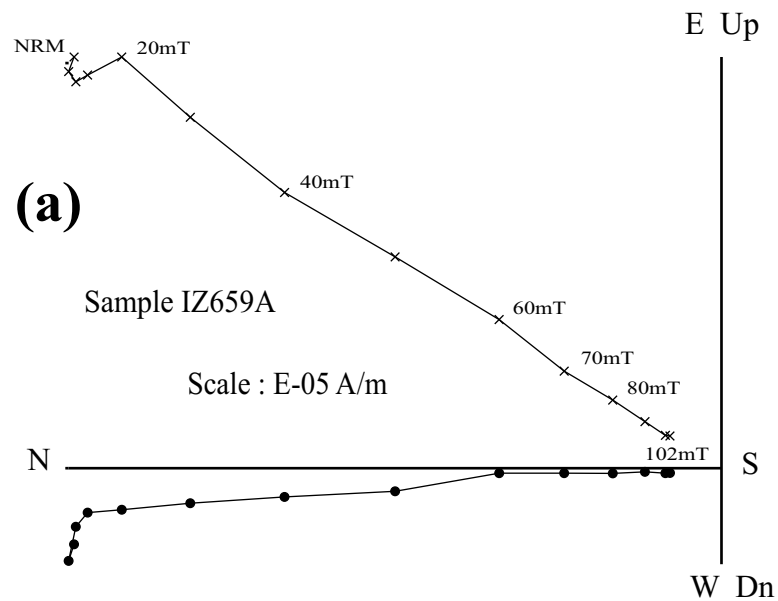




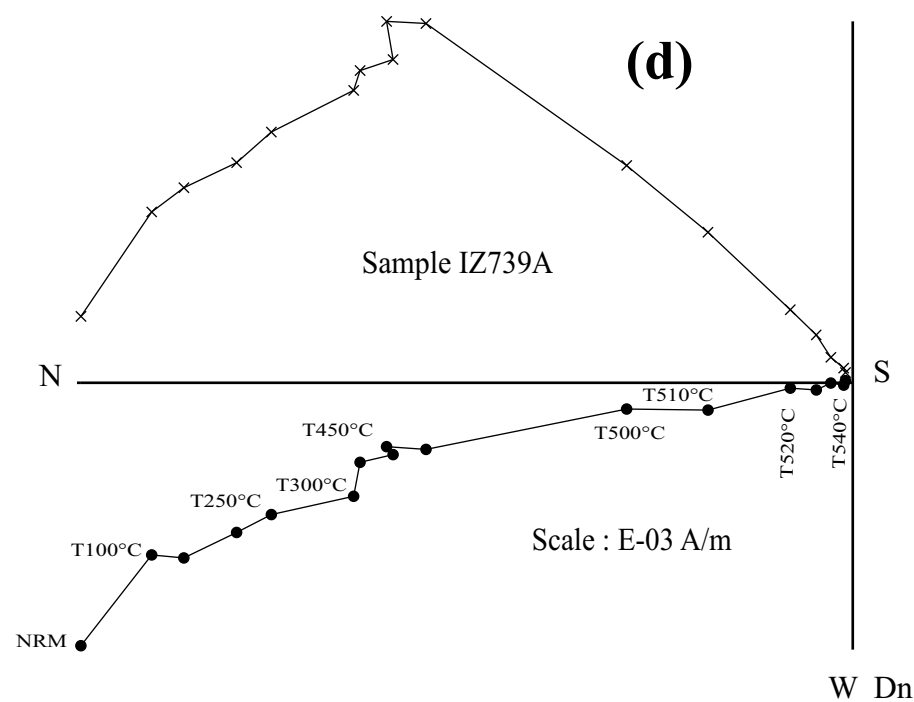
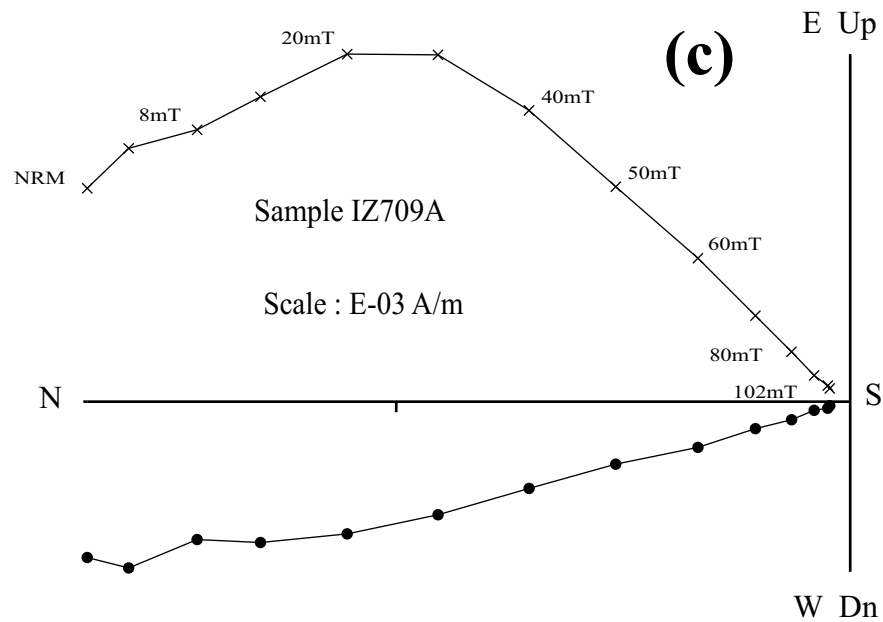
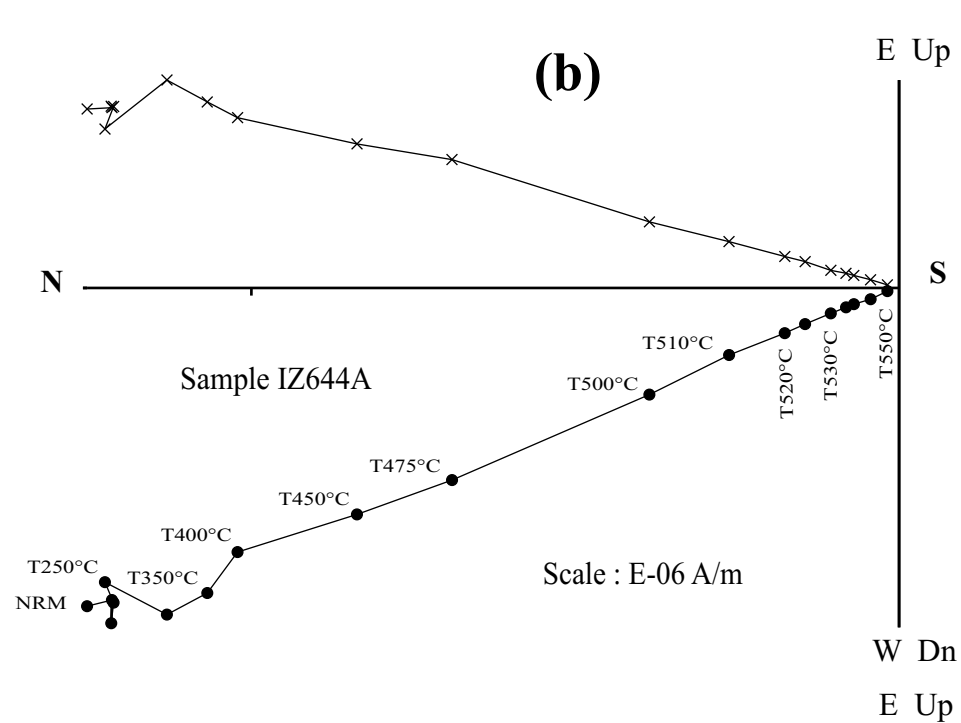


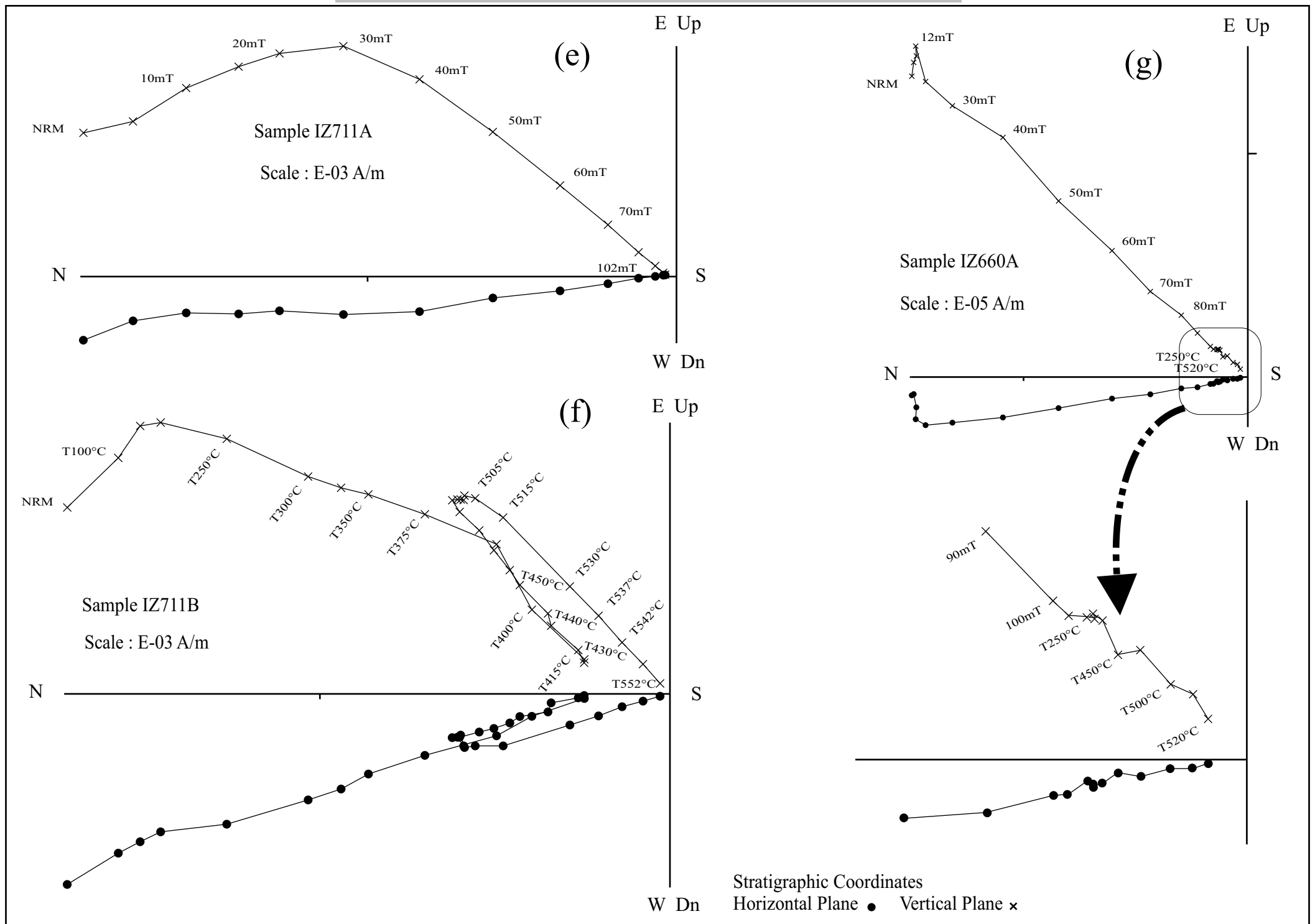


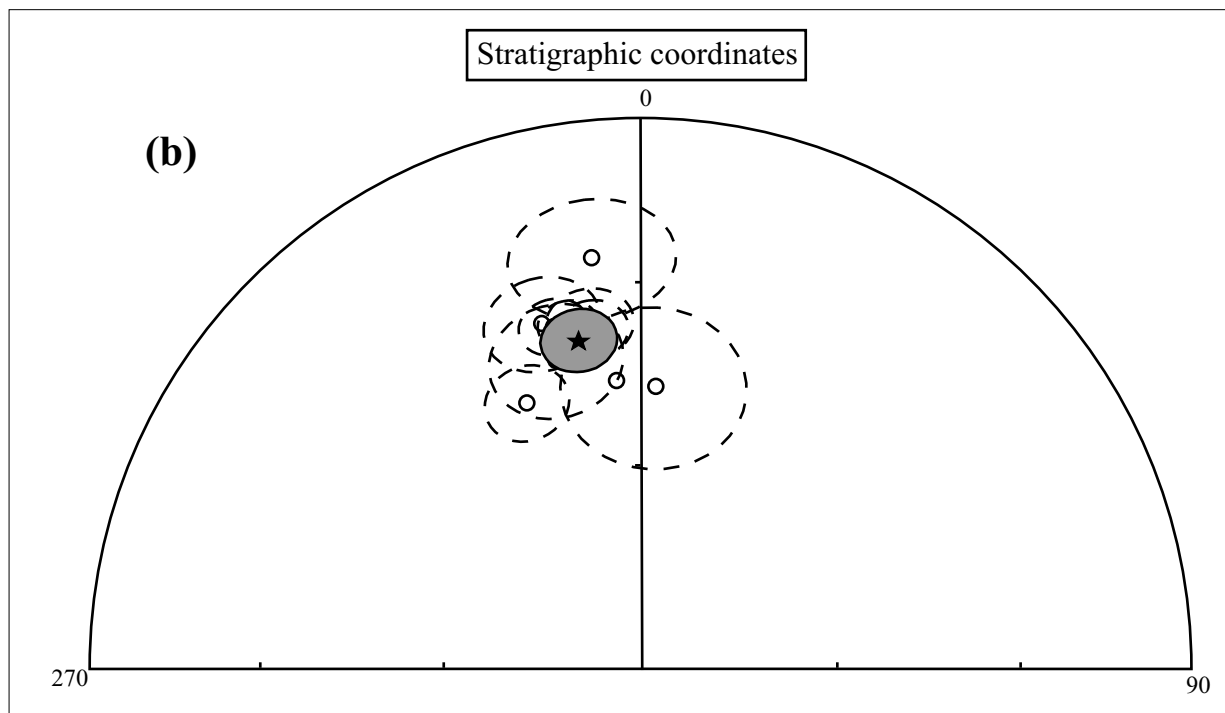
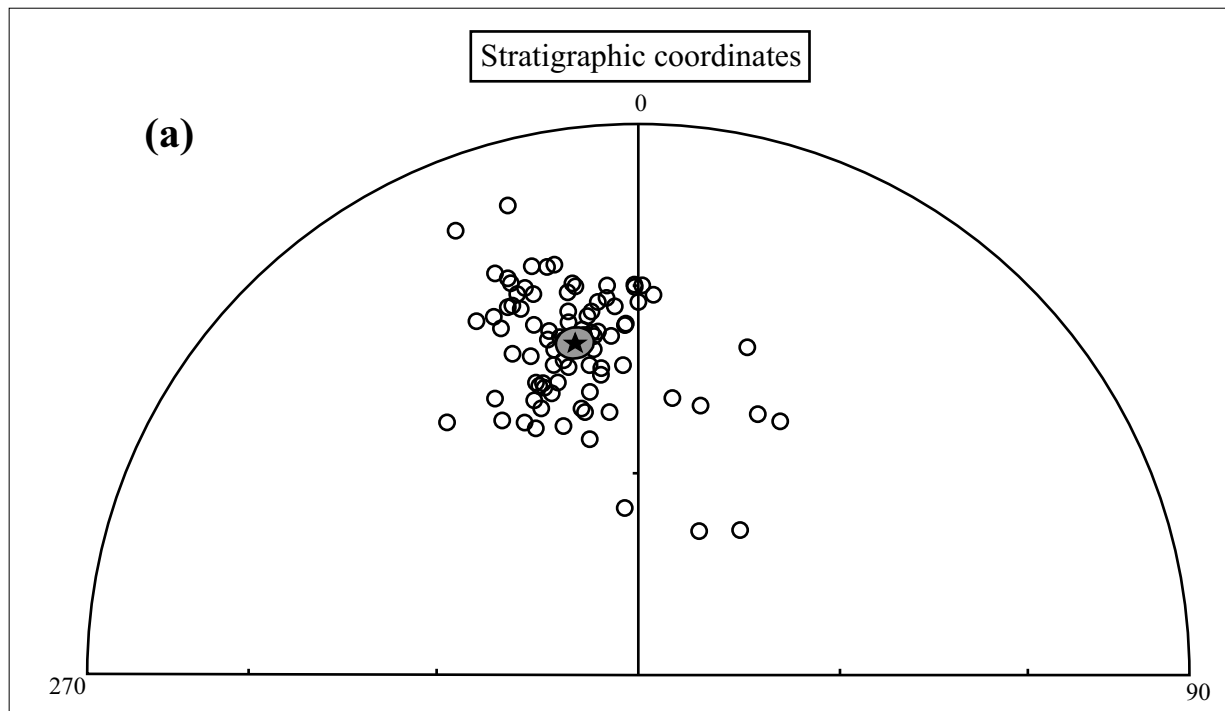


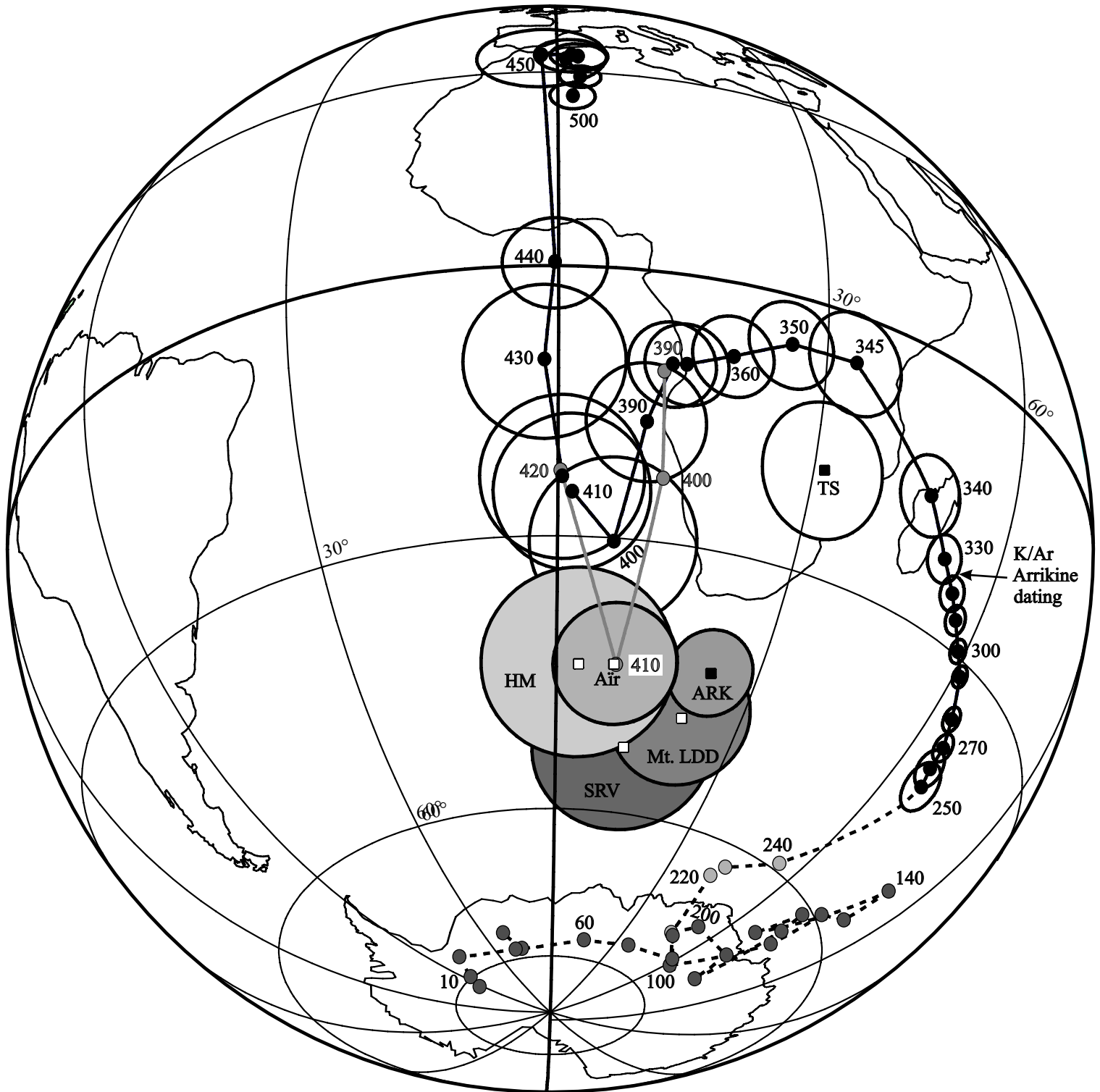


Stratigraphic Coordinates  
Horizontal Plane ● Vertical Plane x











1 A large mafic sill has been discovered in western border of the Murzuq basin  
2 Emplacement age around 410Ma has been obtained by paleomagnetic dating  
3 Emplacement is related to reactivation of mega-shear zones at cratons boundaries  
4 Obtained K/Ar age is that of a rejuvenation attributed to aplitic fluid percolation  
5 These ages correspond to that of the Caledonian and Hercynian unconformities  
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