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# Journal Pre-proof

Revised stratigraphic framework for the lower Anti-Atlas supergroup based on U–Pb geochronology of magmatic and detrital zircons (Zenaga and Bou Azzer-El Graara inliers, Anti-Atlas Belt, Morocco)

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1 **Revised stratigraphic framework for the lower Anti-Atlas Supergroup**  
2 **based on U-Pb geochronology of magmatic and detrital zircons (Zenaga and Bou**  
3 **Azzer-El Graara inliers, Anti-Atlas Belt, Morocco)**

4  
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51 **Abstract**

52 U-Pb geochronology of magmatic and detrital zircons (Zenaga and Bou Azzer-El Graara inliers,  
53 Anti-Atlas Belt, Morocco) and a reassessment of the published constraints suggest a revised  
54 stratigraphic framework for the lower Anti-Atlas Supergroup. Five, major unconformity-bounded  
55 lithostratigraphic packages are here distinguished: the two lower units of Paleoproterozoic age are  
56 named the Tasserda-Taghatine Group (2030-1706 Ma) and the Oumoula (Mimount) Formation (ca.  
57 1745-1650 Ma); the third unit of Paleoproterozoic to Neoproterozoic age (ca. 1650 to > 883 Ma) is  
58 the Tizi n'Taghatine Group; the fourth and fifth units of Neoproterozoic age are the ca. 883 Ma  
59 Tachdamt and the ca. 700 Ma Bleida formations. Implications of this revised stratigraphic  
60 framework include: 1) the Tasserda-Taghatine Group might be linked to the post-orogenic collapse  
61 after the Eburnean Orogeny; 2) the Tizi n'Taghatine Group might be ca. 1.1 Ga in age based on  
62 proposed correlation with the Taoudeni Basin succession in Mauritania; 3) the Bleida Formation  
63 likely reflects deposition in the foreland basin at the early stage of the Pan-African Orogeny; 4) the  
64 Oumoula (Mimount) Formation, Tizi n'Taghatine Group, and Tachdamt Formation potentially  
65 record extensional events within the Nuna/Columbia and Rodinia supercontinents; 5) the  
66 provenance of the lower Anti-Atlas Supergroup (based on our new detrital zircon dating) is mainly  
67 from the West African craton along with possible contributions from other cratons such as  
68 Amazonia and the Sahara Metacraton; 6) the flood basalt sequence of the Tachdamt Formation  
69 likely belongs to the ca. 885-883 Ma intraplate Iguerda-Taïfast Large Igneous Province (LIP) event  
70 defined by previously dated dykes in the Iguerda and Taïfast inliers; and 7) the 1650 Ma Zenaga  
71 LIP can be potentially linked with LIP magmatism in Baltica and Laurentia.

72 **Key Words:** Morocco; Anti-Atlas Belt; Proterozoic; lower Anti-Atlas Supergroup; U-Pb  
73 geochronology

74

75 **Highlights**

76 ► The Taghdout mafic sill of the Ifzwane Suite yielded an age of  $1676 \pm 37$  Ma

- 77 ► The maximum depositional age of the lower Anti-Atlas Supergroup ranges from 2049 Ma  
78 (bottom) to 700 Ma (top) and the provenance is mainly from the WAC
- 79 ► A new 5-part lithostratigraphic framework for the Proterozoic lower Anti-Atlas Supergroup of  
80 the Anti-Atlas Belt is proposed
- 81 ► The 1650-1640 Ma Zenaga and 885-883 Ma Iguerda-Taïfast LIPs provide important stratigraphic  
82 constraints
- 83 ► Correlation between the lower Anti-Atlas Supergroup and the Taoudeni Basin succession in  
84 Mauritania
- 85

Journal Pre-proof

## 86 **1. Introduction**

87 Substantial advances in the understanding of the geological evolution of several major Precambrian  
88 accretionary orogens of North America and Baltica (Hoffman, 1988; Chacko et al., 2000; Corrigan  
89 et al., 2005, 2009; Pehrsson et al., 2013; Bogdanova et al., 2008; 2016) have been made in recent  
90 years. However, there is still a fundamental gap in the knowledge of Precambrian orogenic belts in  
91 Africa. Much of the uncertainty about the accretionary history of the Precambrian African orogens  
92 stems from the limited database of reliable radiometric ages on structurally important units,  
93 combined with a lack of detailed studies using modern kinematic analysis in these deformed regions  
94 as well as of cutting-edge research using sedimentology and sequence stratigraphy.

95 One of the crucial Neoproterozoic African orogens is exposed in the Anti-Atlas Belt of Morocco  
96 located in the northern part of the West African Craton (WAC, Fig. 1). Extensive outcrops of the  
97 Neoproterozoic Pan-African Belt and underlying Paleoproterozoic foreland (Michard et al., 2008;  
98 Gasquet et al., 2008; Soulaïmani et al., 2018 and references therein) are well exposed and have been  
99 studied for a long time (e.g. Choubert, 1963). Traditionally, the Precambrian of the Anti-Atlas Belt  
100 has been divided into three broad lithosomes (e.g. Choubert, 1963): the P I (Archean to  
101 Paleoproterozoic), P II (early and middle Neoproterozoic), and P III (late Neoproterozoic) (Fig. 2).  
102 This lithostratigraphic subdivision has recently been challenged (Walsh et al., 2002; 2012; Thomas  
103 et al., 2002; 2004; Michard et al., 2008; Gasquet et al., 2008; Youbi et al., 2013). In the  
104 conventional interpretation, the oldest, Paleoproterozoic rocks (PI) of the Anti-Atlas Belt are  
105 subdivided into a series of complexes (e.g. the Zenaga and Kerdous complexes in the Siroua and  
106 Kerdous basement inliers, respectively). The rocks of the P II lithosome deposited on this basement  
107 have collectively been called the *Anti-Atlas Supergroup*, within which there are four groups of  
108 volcano-sedimentary units (e.g. Lkest-Taghdout, Bou Azzer, Iriri, and Saghro groups) and they are  
109 conventionally inferred to be Neoproterozoic in age. Various intrusive suites are also recognized  
110 (e.g. Ifzwane and Toudma Suites). These volcano-sedimentary units have been related to the earliest  
111 passive margin, oceanic, and island-arc settings with inferred ages between ca. 800 and 660-640

112 Ma. The earliest Pan-African deformation at 660-640 Ma resulted from closure of the ocean basin,  
113 SW-directed thrusting, and accretion of the island-arc remnants.

114 Recent studies provided new structural, geochronological, and geochemical data, however many  
115 critical questions about the accretionary history of the Anti-Atlas Belt are still debated (e.g.  
116 Soullaimani et al., 2018 and references therein). Such questions include the age of obducted oceanic  
117 crust segments (Siroua and Bou Azzer El Graara ophiolites), relationship between arc terranes to  
118 the north with ophiolites and an associated blueschist facies belt to the south, the age and nature of  
119 the basement in these terranes, the degree of interaction of Neoproterozoic magmas with the  
120 basement, and the timing of accretion of these terranes to the West African Craton. Finally,  
121 discrimination between the post-Eburnean sedimentary succession, recently dated to be late  
122 Paleoproterozoic in age (Ikenne et al., 2017; Soullaimani et al., 2019; Youbi et al., 2019; this work),  
123 and the Neoproterozoic sedimentary sequence (Leblanc and Moussine-Pouchkine, 1994; Bouougri  
124 and Saquaque, 2004; Blein et al., 2014a,b; Soullaimani et al., 2018; Youbi et al., 2018) remains  
125 uncertain.

126 The lower Anti-Atlas Supergroup of the Anti-Atlas Belt (Lkest-Taghdout Group *sensu* Thomas et  
127 al., 2004; Figs. 1D, 2 and 3) is one of the most important Proterozoic lithostratigraphic units on the  
128 WAC, yet with poor age constraints. It consists essentially of basalts, quartzites, and  
129 stromatolitic/oolitic carbonates, intruded by dolerite dyke swarms and sill complexes (Ifzwane and  
130 Toudma Suites). The lower Anti-Atlas Supergroup was previously thought to be ca. 1000-800 Ma  
131 in age on the basis of (1) stromatolites characteristic for the Neoproterozoic carbonate successions  
132 (Choubert, 1962), and (2) the Rb/Sr age of the contact-metamorphosed host rocks associated with  
133 mafic dykes ( $789 \pm 10$  Ma) in the Tachdamt Subinlier of the Bou Azzer El Graara Inlier (Clauer,  
134 1976; Cahen et al., 1984).

135 The present study is based on field observations and new U-Pb ages for igneous zircons from  
136 dolerite sill intruding carbonates from the lower Anti-Atlas Supergroup in the Zenaga Inlier and  
137 detrital zircon grains from four samples from the lower, middle, and upper parts of the lower Anti-



138 Atlas Supergroup in the Zenaga and Bou Azzer El Graara inliers. We propose a new  
139 lithostratigraphy for the Proterozoic lower Anti-Atlas Supergroup of the Anti-Atlas Belt based on  
140 a discussion of our data in combination with the previously published data (Abati et al., 2010;  
141 Walsh et al., 2012; Letsch, 2018; Bouougri et al., 2020) and constrain the provenance for the five,  
142 major unconformity-bounded packages of the supergroup.

## 143 **2. Geological background**

### 144 **2.1. The West African Craton**

145 The West African Craton (WAC, Fig. 1) extends across northwestern Africa, and consists of two  
146 Archean basement shields, the Reguibat Shield to the north and the Leo-Man Shield to the south.  
147 The Archean nuclei (3.5-2.7 Ga in age) of these shields are juxtaposed against an array of  
148 Paleoproterozoic domains (2.35-2.00 Ga in age) made up of greenstone belts and regions of  
149 extensive tonalite-trondhjemite-granodiorite (TTG) plutons, which are overlain by Meso- to  
150 Neoproterozoic and younger sedimentary basins. Smaller basement inliers are composed of  
151 Paleoproterozoic rocks: the Kayes and Kedougou-Kenieba inliers to the west, and those of the Anti-  
152 Atlas Belt to the north (Ennih and Liégeois, 2008; Abati et al., 2012). The WAC remained largely  
153 stable since 2 Ga (Schofield et al., 2006; Jessell et al., 2016). The basement of the WAC has  
154 evolved through several major orogenic cycles: the Paleoproterozoic Leonian cycle (several episodes  
155 between 3.40 and 3.10 Ga with uncertain chronology, related to continental accretion; e.g.  
156 Thiéblemont et al., 2004; Koffi et al., 2020), the Liberian cycle (2.85-2.70 Ga; Key et al., 2008;  
157 Koffi et al., 2020), the Paleoproterozoic Eburnean and Birimian cycles (2.15-2.07 Ga and 2.27-1.96  
158 Ga, respectively; Abouchami et al., 1990; Boher et al., 1992; Egal et al., 2002; Schofield et al.,  
159 2006; Baratoux et al., 2011; Grenholm et al., 2019; McFarlane et al., 2019) and the late  
160 Neoproterozoic Pan-African orogenic event (885-550 Ma; Thomas et al., 2002; 2004; Gasquet et  
161 al., 2005; 2008; Walsh et al., 2012; Youbi et al., 2018). The WAC hosts four major intracratonic  
162 basins: the large central Taoudeni Basin and the smaller Tindouf and Volta basins, to the north and  
163 southwest, respectively, and the Gourma aulacogen to the east (Moussine-Pouchkine and Bertrand-

164 Sarfati, 1978; Villeneuve, 2005; Deynoux et al., 2006; Lottaroli et al., 2009). These basins comprise  
165 mostly Mesoproterozoic-Neoproterozoic to Devonian-Carboniferous sediments. The WAC is  
166 surrounded by Pan-African belts, including the Trans-Saharan and Anti-Atlas belts (Ennih and  
167 Liégeois, 2008; Liégeois et al., 2013; Toummite et al., 2013; Brahimy et al., 2018).

168 Early compilations of WAC lithostratigraphy suggested an absence of Mesoproterozoic events or  
169 rocks, indicating a tectonically quiescent period between 1.7 and 1.0 Ga (e.g. Ennih and Liégeois,  
170 2008 and references therein). The lack of detrital zircons of that age in the pre-Cryogenian strata is  
171 considered characteristic of sediments sourced from the WAC (e.g. Abati et al., 2010; Gärtner et al.,  
172 2013; 2017; 2018). However, recent studies in the northern WAC (El Bahat et al., 2013; Kouyaté et  
173 al., 2013; Söderlund et al., 2013; Youbi et al., 2013; Ikenne et al., 2017) demonstrated important  
174 intraplate magmatic events based on dating of mafic dykes by the U-Pb method on baddeleyite and  
175 zircon. Six distinct events are distinguished throughout the inliers: ca. 2040 Ma Tagragra of Tata,  
176 1750 Ma Tagragra of Akka, 1650 Ma Zenaga, 1416–1380 Ma Bas Drâa, and 885-883 Ma Iguerda-  
177 Taïfast events, and 612-600 Ma pulse of the Central Iapetus Magmatic Province (CIMP; Walsh et  
178 al., 2002; Gasquet et al., 2004; El Bahat et al., 2013; Kouyaté et al., 2013; Söderlund et al., 2013;  
179 Youbi et al., 2013; El Bahat et al., 2017; Ikenne et al., 2017; Bouougri et al., 2020 ; Youbi et al.,  
180 2020). Re-Os geochronology of the sedimentary succession of the Atar Group in the Taoudeni  
181 Basin of Mauritania indicated depositional ages ranging from  $1109 \pm 22$  to  $1105 \pm 37$  Ma (Rooney  
182 et al., 2010). These Mesoproterozoic ages are confirmed by the study of microfossils (Beghin et al.,  
183 2017) and chemostratigraphy (Kah et al., 2012). Recently, Baratoux et al. (2019) identified four  
184 new magmatic events 1.79-1.76, 1.58, 1.53, and 0.91-0.87 Ga in the southern WAC that add  
185 additional evidence for Mesoproterozoic mafic intrusive activity on this craton. Ar-Ar ages of ca.  
186 1.24 Ga for the 1575 and 1520 Ma dyke swarms are considered to record a thermal overprint  
187 (Tapsoba et al., 2018).

## 188 **2.2. The Anti-Atlas Belt**

189 The ENE-WSW trending Anti-Atlas Belt (800 km long and 200 km wide) of Morocco is the most  
190 important segment of the Neoproterozoic Pan-African Orogen in the northern part of the WAC (Fig.  
191 1D). The Precambrian basement crops out in several erosional inliers (“boutonniers” in French),  
192 containing Late Ediacaran and younger units (Bas Drâa, Ifni, Kerdous, Tagragra of Akka, Tagragra  
193 of Tata, Igherm, Siroua, Zenaga, Bou Azzer, Saghro, and Ougnat; Fig. 1D) and distributed along  
194 two major fault systems (South Atlas Fault, SAF and Anti-Atlas Major Fault, AAMF; Fig. 1D).

195 The Anti-Atlas Belt contains the Paleoproterozoic basement (affected by the Eburnean Orogeny)  
196 unconformably overlain by the lower Anti-Atlas Supergroup (= Lkest-Taghdout Group *sensu*  
197 Thomas et al., 2004; = Tizi n’Taghatine Group *sensu* Bouougri and Saquaque, 2004), which is in  
198 turn unconformably overlain by the Ouarzazate Supergroup related to the Pan-African Orogeny.  
199 The two major periods of tectonothermal activity, associated with crustal accretion, have been  
200 recognized during the Proterozoic (Fig. 2) in the Anti-Atlas Belt and are described below.

201 (i) *The Paleoproterozoic Eburnean Orogeny (2.20-2.07 Ga*; Abouchami et al., 1990; Boher et al.,  
202 1992; Egal et al., 2002; Schofield et al., 2006; Baratoux et al., 2011; McFarlane et al., 2019). The  
203 Paleoproterozoic basement rocks are found exclusively SW of the AAMF (Choubert, 1963), but  
204 they may also constitute the basement to the Pan-African Orogen to the NE (Ennih and Liégeois,  
205 2001; Liégeois et al., 2006; Gasquet et al., 2008). The Paleoproterozoic basement is composed of  
206 metasedimentary rocks, granites, paragneisses, and migmatites, with U-Pb zircon ages ranging from  
207 2200 to 2030 Ma (e.g. Aït Malek et al., 1998; Walsh et al., 2002; Thomas et al., 2002; Gasquet et  
208 al., 2004, 2008; O’Connor, 2010).

209 (ii) *The Neoproterozoic Pan-African Orogeny (885–550 Ma)*. The Neoproterozoic sequences of the  
210 Anti-Atlas Belt are composed of units that were involved in the Pan-African Orogeny and the  
211 unconformably overlying post-collisional volcano-sedimentary rocks of the Ouarzazate Supergroup,  
212 which are a part of the Central Iapetus Magmatic Province (CIMP; Doblus et al., 2002; Maloof et  
213 al., 2005; Gasquet et al., 2005, 2008; Ernst and Youbi, 2017; Tuduri et al., 2018; Youbi et al.,  
214 2020). The Neoproterozoic rocks underlying the Ouarzazate Supergroup are subdivided into the

215 lower units (Lkest-Taghdout, Bou-Azzer, and Irii groups), affected by all Pan-African orogenic  
216 events, and the upper units (Saghro and Bou Salda groups), only affected by the latest stage of the  
217 Pan-African Orogeny (e.g. Walsh et al., 2002; Thomas et al., 2004; Gasquet et al., 2005; 2008;  
218 Alvaro et al., 2014).

219 The lower Anti-Atlas Supergroup consists essentially of basalt, quartzite, and stromatolitic/oolitic  
220 carbonate, cut by undated dolerite dyke swarms and sill complexes (Thomas et al., 2004; Bouougri  
221 and Saquaque, 2004). It marks the onset of a period of rifting at the northern portion of the WAC  
222 during the late Neoproterozoic. The rifting culminated with the opening of an oceanic basin on the  
223 northern edge of the WAC. The relics of the oceanic crust (Bou-Azzer Group) are preserved in the  
224 Bou-Azzer and Siroua inliers as highly sheared allochthonous ophiolite complexes (El Hadi et al.,  
225 2010; Blein et al., 2014a and references therein). The ophiolitic assemblage is mainly made up of  
226 ultramafic rocks (serpentinites with few chromite pods), but also includes mafic meta-cumulates,  
227 meta-basaltic sheeted dykes, and, to a smaller extent, pillow lavas (Leblanc, 1975, 1981; Admou  
228 and Juteau, 1998). Their geochemical signatures point to emplacement in a supra-subduction zone  
229 (Bodinier et al., 1984; Naidoo et al., 1991; Ahmed et al., 2005; Hodel et al., 2017; Triantafyllou et  
230 al., 2019; 2020). Dating of plagiogranite intrusions in the Siroua Inlier indicated formation of an  
231 oceanic crust at ca. 760 Ma (U-Pb zircon age; Samson et al., 2004). Given the absence of reliable  
232 radiometric data on the age of the lower Anti-Atlas Supergroup, ca. 760 Ma was considered as the  
233 minimum age of its formation. The rifting might be associated with the late Neoproterozoic  
234 extension within the supercontinent Rodinia (Gasquet et al., 2008). Following the opening of the  
235 ocean basin, a subduction phase was initiated with island-arc development (the Irii Group in the  
236 Siroua Inlier; Thomas et al., 2002; Triantafyllou, 2015; 2016). The cores and overgrowths of zircon  
237 grains from the Irii Group migmatite yielded two U–Pb SHRIMP ages at  $743 \pm 14$  and  $663 \pm 13$   
238 Ma, respectively (Thomas et al., 2002). The older age obtained from the cores is interpreted as the  
239 age of crystallization and emplacement of the migmatite protolith in an island–arc setting, whereas  
240 the younger age is thought to represent the timing of ophiolite obduction and arc accretion to the

241 northern edge of the WAC in association with the regional metamorphism and the main orogenic  
242 event in the Anti-Atlas Belt (Thomas et al., 2002).

243 The development of the Anti-Atlas Belt oceanic arc complexes occurred during three magmatic  
244 flare-ups at ca. 750, 700, and 650 Ma that were interspersed with an early major tectonic episode  
245 between 730-700 Ma (Triantafyllou et al., 2018; 2020). A garnet Sm-Nd date of  $647.2 \pm 1.7$  Ma  
246 was recently obtained for the Tasriwine ophiolite complex in the Siroua Inlier (Inglis et al., 2016).  
247 This age is more than 15 million years younger than the previous age estimate by Thomas et al.  
248 (2002). North of the AAMF in the eastern Anti-Atlas Belt (Figs. 1D and 2), the oldest rocks are  
249 thick deposits of basalt and turbidite of the Saghro Group, interpreted as either a distal-slope  
250 equivalent of the upper part of the lower Anti-Atlas Supergroup, deposited between the WAC to the  
251 south and an island arc (Saquaque et al., 1989; Thomas et al., 2002), or slivers of the stretched  
252 Eburnean crust (Hindermeyer, 1953; Ouguir et al., 1996; Fekkak et al., 2001; 2002; Massironi et al.,  
253 2007; Schiavo et al., 2007; Michard et al., 2017).

254 Preliminary detrital zircon data reported by Liégeois et al. (2006) and summarized in Gasquet et al.  
255 (2008) suggest that the sediments of the Saghro Group in the Kelaa Mgouna section were derived  
256 from sources ranging in age from about 880 to 615 Ma. The later post-collisional, extensional stage  
257 (605–542 Ma) is characterized by molasse deposition and magmatism of the Ouarzazate  
258 Supergroup, which was followed by the foreland development in the Saharan cratonic basin to the  
259 south (Taroudant and Tata groups; e.g. Doblas et al., 2002; Maloof et al., 2005; Gasquet et al.,  
260 2005, 2008; Hefferan et al., 2014; Karaoui et al., 2014; 2015; Tuduri et al., 2018; Geyer and  
261 Landing, 2006; 2020; Youbi et al., 2020).

262

### 263 **2.3. The lower Anti-Atlas Supergroup**

264 Since our study is focused on the lower Anti-Atlas Supergroup, we will give here an updated review  
265 of its lithology, stratigraphy, and nomenclature. The lower Anti-Atlas Supergroup, known in the  
266 Anti-Atlas literature as the Taghdout (Lkest/Tizi n'Taghatine) Group, contains a sedimentary

267 succession deposited along the northern margin of the WAC (e.g. Thomas et al., 2002, 2004;  
268 Bouougri and Saquaque, 2004; Alvaro et al., 2014). The lower part of the lower Anti-Atlas  
269 Supergroup shows a progressive change of sedimentary facies from SW to NE representing a broad  
270 proximal to distal trend: conglomerates, sandstones, and quartz sandstones predominate in the Ifni  
271 and Kerdous inliers, while quartz sandstones, carbonates, shales, turbidites, and basaltic lava flows  
272 (the so called “Série des Calcaires et Quartzites”; see Choubert, 1962) developed in the Igherm,  
273 Siroua, Zenaga, and Bou Azzer El Graara inliers (Bouougri and Saquaque, 2004; Thomas et al.,  
274 2004; Alvaro et al., 2014). The scarcity of structural studies combined with modern sedimentology  
275 and sequence stratigraphy to test correlations between the groups in different inliers has led to a  
276 proliferation of synonymous names in the lithostratigraphic terminology: e.g. the Jbel Lkest Group  
277 in the Kerdous Inlier (Hassenforder, 1987; O’Connor, 2010), the Bleida-Tachdamt or Bleida Group  
278 in the Bou Azzer El Graara Inlier (Leblanc and Moussine-Pouchkine, 1994; Blein et al., 2014a), and  
279 the Taghdout Group (= Tizi n’Taghatine) in the Zenaga Inlier (Thomas et al., 2002, 2004; Bouougri  
280 and Saquaque, 2004). Correlation between these groups (recently combined into the Lkest-  
281 Taghdout Group, equivalent to the lower Anti-Atlas Supergroup; Gasquet et al., 2008; Álvaro et al.,  
282 2014) is both lithostratigraphic and structural. The only lithostratigraphic correlation linking the  
283 Siroua-Zenaga and Bou Azzer El Graara inliers was proposed by Bouougri and Saquaque (2004)  
284 who considered the section of the Tizi n’Taghatine mountain pass as a stratotype for the group  
285 named after this locality. Although their lithostratigraphy requires revision, we will refer to it in this  
286 work. Two reasons motivated this choice: (1) the dated samples of this study are from the same area  
287 that Bouougri and Saquaque (2004) focused on, and (2) the Siroua-Zenaga and Bou Azzer El  
288 Graara inliers are the only areas where modern sedimentology and sequence stratigraphy were  
289 applied (Fig. 1D). However, we will use the name of the lower Anti-Atlas Supergroup (an  
290 equivalent of the Lkest-Taghdout Group) because, although the latter group is widely known in the  
291 literature (Gasquet et al., 2008; Álvaro et al., 2014), it combines genetically unrelated sequences  
292 deposited in superimposed sedimentary basins and in different tectonic settings with long-lasting

293 hiatuses along unconformities. Five, major unconformity-bounded lithostratigraphic packages are  
294 further here distinguished: the lower Tasserda-Taghatine Group, the second Oumoula (Mimount)  
295 Formation, the third Tizi n'Taghatine Group, the fourth Tachdamt Formation, and the uppermost  
296 Bleida Formation.

297 For a long time, until 2013, available data on the lower Anti-Atlas Supergroup suggested that the  
298 continental passive margin developed during the Tonian between about 1000 Ma (based on  
299 stromatolite morphology pointing to a Neoproterozoic age; Choubert, 1963) and ca. 787 Ma (Rb-Sr  
300 isochron age of contact metamorphism of the host rocks associated with mafic dykes at Tachdamt;  
301 Clauer, 1976; Cahen et al., 1984). However, these age constraints are still debated (Gasquet et al.,  
302 2008; Alvaro et al., 2014; Ikenne et al., 2017; Soulaïmani et al., 2018; 2019; Youbi et al., 2019).  
303 There are too few reliable geochronologic age constraints to tightly constrain the age of the lower  
304 Anti-Atlas Supergroup. It unconformably overlies the Eburnean basement, including the post-  
305 tectonic  $2032 \pm 5$  Ma Azguemerzi granodiorite of the Zenaga Inlier, and in turn is unconformably  
306 overlain by the subhorizontal volcanic and sedimentary succession of the Late Ediacaran (ca. 606-  
307 541 Ma) Ouarzazate Supergroup (Figs. 6A, B). The lower Anti-Atlas Supergroup records a low-  
308 grade tectonometamorphic event related to the Pan-African Orogeny dated at  $663 \pm 13$  Ma (Thomas  
309 et al., 2002) and  $647.2 \pm 1.7$  Ma (Inglis et al., 2016) in the Siroua Inlier. Consequently, the  
310 depositional age of the lower Anti-Atlas Supergroup can be broadly bracketed between  $2032 \pm 5$   
311 Ma and  $663 \pm 13 / 647.2 \pm 1.7$  Ma.

312 The lower Anti-Atlas Supergroup sediments are intruded by abundant dolerite and gabbro dykes  
313 and sills, also found as dyke swarms and sill complexes in the Paleoproterozoic basement of the  
314 Zenaga (Fig. 4) and several other inliers (e.g. Igherm, Tagragra of Akka, Tagragra of Tata, Kerdous,  
315 and Bas Drâa). These rocks have been collectively called the Ifzwane Suite in the Zenaga and  
316 Siroua inliers (Thomas et al., 2004), and the Toudma Suite in the Kerdous Inlier (Thomas et al.,  
317 2004; O'Connor, 2010).

318 Letsch (2018) showed that the quartzites of the Mimount Formation (*sensu* Thomas et al., 2002)  
319 and the Oumoula Formation (*sensu* Bouougri and Saquaque, 2004), overlying the basal, carbonate-  
320 quartzite-metapelite units near the Town of Taghdout at the northern margin of the Zenaga Inlier  
321 (see Thomas et al., 2002), do not contain detrital zircons younger than ca. 1987 Ma (Benziane,  
322 2007; Walsh et al., 2012) and ca. 1809 Ma (Abati et al., 2010), respectively. Neoproterozoic detrital  
323 zircon grains only appear in the overlying formations of the Saghro and Bou Salda groups of the  
324 Siroua Inlier (Abati et al., 2010). Detrital zircons from the Mimount and Oumoula formations  
325 display exclusively Paleoproterozoic ages, suggesting a ca. 1890 Ma maximum depositional age for  
326 the upper parts of the formations. In the Ifni Inlier, detrital zircons from quartzites of the upper part  
327 of the Lkest Group (the basal group of the lower Anti-Atlas Supergroup) exhibit two main clusters  
328 in distribution of SHRIMP  $^{207}\text{Pb}/^{206}\text{Pb}$  ages, 2060-1890 and 1810-1640 Ma (Schulte et al., 2016).  
329 The maximum depositional age is given by a weighted average  $^{207}\text{Pb}/^{206}\text{Pb}$  date of  $1744 \pm 26$  Ma  
330 based on four youngest concordant grains.

331

### 332 **3. Sampling and Analytical Procedures**

#### 333 **3.1. Sampling**

334 In order to investigate the maximum depositional age of the lower Anti-Atlas Supergroup of the  
335 Anti-Atlas Belt, we collected five samples from two key inliers within the Anti-Atlas Orogen  
336 (Zenaga and Bou Azzer El Graara) for geochronological investigation. The Taghdout sill that  
337 intruded to the lower part of the lower Anti-Atlas Supergroup in the Zenaga Inlier was sampled for  
338 geochemistry and zircon U-Pb geochronology: sample TA1 was collected at the centre of the sill  
339 and herein dated at  $1676 \pm 37$  Ma and sample TA2 is from the margin of the sill. Four  
340 metasedimentary samples were collected for detrital zircon U-Pb geochronology: one sample from  
341 metasilstones of the Tasserda Formation of the Zenaga Inlier and three samples from the  
342 Agoumy, Tachdamt, and Bleida formations of the lower Anti-Atlas Supergroup in the Bou Azzer



343 Inlier (Figs. 3, 4, and 5). Sample site locations, a brief description of the samples, and their ages are  
344 provided in Table 1.

345

### 346 **3.2. Analytical procedures**

#### 347 **3.2.1. Whole-rock geochemistry**

348 Elemental concentrations of whole rocks were obtained at “Service d'Analyse des Roches et des  
349 Minéraux” at the Geochemical and Petrographical Research Center in Nancy (SARM laboratory-  
350 CRPG-CNRS, Nancy, France). To avoid contamination, crushing and pulverizing were undertaken  
351 using agate mills. Powders were treated with  $\text{LiBO}_2$ , dissolved with  $\text{HNO}_3$ , and analyzed and  
352 calibrated with international geostandards (Basalt BE-N, Basalt BR, Dolerite WS-E, Microgabbro  
353 PM-S). Major and trace elements were analyzed using inductively coupled plasma atomic emission  
354 spectroscopy (ICP-AES) and Inductively Coupled Mass Spectrometry (ICP-MS), respectively. The  
355 analytical procedures are described in detail at <http://helium.crgp.cnrs-nancy.fr/SARM/>.

356

#### 357 **3.2.2. U-Pb Geochronology**

358 Zircon grains from all studied samples were separated from 3-5 kg of crushed rocks using  
359 conventional heavy liquid and magnetic separation techniques (jaw crusher, disk grinder, Wilfley  
360 table, Frantz isodynamic magnetic separator, and density separation using bromoform and  
361 methylene iodide) at the Centro de Pesquisas Geocronológicas, Universidade de São Paulo (CPGeo-  
362 USP), Brazil. About 20 to 100 zircon grains from each sample were mounted in epoxy resin,  
363 polished to approximately half of mean grain thickness for further imaging with transmitted light  
364 and cathodoluminescence to resolve internal grain texture. After coating with Au,  
365 cathodoluminescence (CL) images of zircon grains were obtained using a Quanta 250 FEG  
366 scanning electron microscope equipped with Mono CL3 + cathodoluminescence spectroscope  
367 (Centaurus) at the Centro de Pesquisas Geocronológicas of the University of São Paulo (CPGeo-  
368 USP), Brazil. The conditions used in CL analysis were as follows: 60  $\mu\text{A}$  emission current, 15.0 kV

369 accelerating voltage, 7  $\mu\text{m}$  beam diameter, 200  $\mu\text{s}$  acquisition time, and a resolution of  $1024 \times 884$   
370 pixels.

371

### 372 **3.2.2.1 Sensitive High Resolution Ion Microprobe (SHRIMP II) analysis**

373 Igneous zircon grains from the dolerite sample TA 1 and detrital zircon grains from the quartzite  
374 samples BA21 and BA24 were analyzed by the U-Pb isotopic technique using a SHRIMP II at  
375 CPGeo-USP following the analytical procedures of Williams (1998). The crystals were mounted,  
376 together with the TEMORA-2 reference zircon (Black et al., 2004), in epoxy and polished to expose  
377 the interior of the grains. Correction for common Pb was made based on  $^{204}\text{Pb}$  measured, and the  
378 typical error for the  $^{206}\text{Pb}/^{238}\text{U}$  ratio was less than 2%; uranium abundance and U/Pb ratios were  
379 calibrated against the TEMORA-2 reference material, and the ages were calculated using Isoplot®  
380 version 3.0 software application (Ludwig, 2003). Errors are reported as  $1\sigma$  deviations and ages have  
381 been calculated at the 95% confidence level.

### 382 **3.2.2.2 Laser Ablation Inductively Coupled Plasma Mass Spectrometry (LA-ICP-MS) analysis**

383 U-Pb analyses of detrital zircon grains from the samples TG-15-1 and ALBL29 were performed  
384 using a Neptune multicollector inductively coupled plasma mass spectrometer (MC-ICP-MS) and  
385 an Analyte G2 excimer laser ablation (LA) system at CPGeo-USP, Brazil. The U-Pb analysis was  
386 performed in the following order: two blanks, two 612 NIST standard glasses, three external  
387 standards, 13 unknown samples, two external standards, and two blanks. An interpolation of the  
388 four blank measurement sequences, using the bracketing method with two blanks before and two  
389 after the samples, was subtracted from all the seven measured masses (202, 204, 206, 207, 208, 232,  
390 and 238). The  $^{235}\text{U}$  isotope was not measured, but calculated using the relative abundance  
391  $^{238}\text{U}/^{235}\text{U}=137.88$ . The  $^{204}\text{Hg}$  interference on  $^{204}\text{Pb}$  was corrected using  $^{202}\text{Hg}$  and assuming  
392  $^{204}\text{Hg}/^{202}\text{Hg} = 4.350360$ . Using the partially corrected  $^{207}\text{Pb}/^{206}\text{Pb}$  ratio as an estimate for the age  
393 and the equations of Stacey and Kramers (1975), the  $^{206}\text{Pb}/^{204}\text{Pb}$ ,  $^{207}\text{Pb}/^{204}\text{Pb}$ , and  $^{208}\text{Pb}/^{204}\text{Pb}$   
394 relative abundances were calculated and corrected for common (non-radiogenic) Pb fraction.

395 Bracketing measurement of unknowns with NIST-612 standard glasses (two measurement  
396 sequences before and after) was used to convert the total Pb, Th, and U signals to ppm  
397 concentrations.

398 A final bracketing (with three measurements before and two after unknown) with the GJ-1 reference  
399 zircon (Jackson et al., 2004) was used to correct the effect of the instrumental fractionation on four  
400 ratios ( $^{206}\text{Pb}/^{238}\text{U}$ ,  $^{207}\text{Pb}/^{235}\text{U}$ ,  $^{207}\text{Pb}/^{206}\text{Pb}$ , and  $^{208}\text{Pb}/^{232}\text{Th}$ ) before they were used to calculate the  
401 respective ages. Errors in all arithmetic operations were algebraically propagated. Least squares  
402 fitting was used for the best line fit and all the data are reported at the  $1\sigma$  level calculated with  
403 Isoplot 3.70 (Ludwig, 2008). A U-Pb analysis is concordant when it overlaps with the Concordia  
404 within uncertainty. So, it seems to be appropriate to exclude results with a low level of concordance  
405 ( $^{206}\text{Pb}/^{238}\text{U}$  age /  $^{207}\text{Pb}/^{206}\text{Pb}$  age  $\times 100$ ), but very large errors that overlap with the Concordia from  
406 interpretation. Thus, an interpretation with respect to the obtained ages was done for all grains  
407 within the concordance interval of 90-110 % ( $^{206}\text{Pb}/^{238}\text{U}$  age /  $^{207}\text{Pb}/^{206}\text{Pb}$  age  $\times 100$ ) which is  
408 often used (e.g. Spencer et al., 2016). Discordant analyses were generally interpreted with caution.  
409 Analyses with discordance of more than 10% were discarded. For zircons younger than 1300 Ma,  
410  $^{206}\text{Pb}/^{238}\text{U}$  ages were preferred, while for older zircons the  $^{207}\text{Pb}/^{206}\text{Pb}$  ages were used. The detrital  
411 zircon age distribution by probability density was performed using the AgeDisplay application (an  
412 EXCEL workbook to evaluate and display univariate geochronological data using binned frequency  
413 histograms and probability density distributions) (Sircombe, 2004).

414

## 415 **4. Results**

### 416 **4.1. The Great Mafic Sill of the Ifzwane Suite at Taghdout Town**

#### 417 **4.1.1. Field observations**

418 The Taghdout sill was collected near Taghdout Town, ca. 9 km NW of Tazenakht City (Table 1 and  
419 Figs. 3 and 4), from a 227 m thick, well-preserved, massive sill belonging to the Ifzwane Suite  
420 (dolerite dyke swarm and sill complex of the Zenaga Inlier, of likely multiple ages as discussed in

421 section 2.1). The sill intruded into the lower part of the lower Anti-Atlas Supergroup (Fig. 6C),  
422 which mainly consists of carbonate and mixed siliciclastic-carbonate sequence of the Ifrane-n-  
423 Taghatine Formation (Bouougri and Saquaque, 2004). The sill can be traced in outcrops for more  
424 than one km in length and has thick chilled margins of marble grading away from the intrusion into  
425 carbonate host rock (Figs. 3, 4, and 6D). The following new samples were used in this study: TA1  
426 (centre of the sill) and TA2 (border of the sill) from the Taghdout sill of the Zenaga Inlier.

#### 427 **4.1.2. Petrographic features**

428 Fresh samples of the Taghdout sill are dark-grey, but due to widespread hematitic alteration the  
429 rock appears reddish. However, the original texture and primary igneous mineralogy are usually  
430 well preserved. The texture varies from a coarse-grained doleritic texture in the centre of the sill to a  
431 fine-grained, microlitic porphyric texture at the border of the sill (Fig. 7).

432 The primary igneous minerals consist of clinopyroxene (augite and locally pigeonite), very rare  
433 orthopyroxene, plagioclase, and oxides. Apatite and zircon are present as accessory minerals;  
434 amphibole and, more rarely, biotite are found around pyroxene crystals or as irregular patches in the  
435 groundmass. Although tholeiitic mafic rocks are generally characterized by low alkaline content  
436 ( $\text{Na}_2\text{O}+\text{K}_2\text{O}$ ), granophyric quartz/K-feldspar mesostasis was observed in coarse-grained portions of  
437 the sill, consistent with the examples previously described in literature (e.g. Fodor, 2009; Block et  
438 al., 2015). The secondary mineral assemblage includes albite, chlorite, actinolite, epidote, titanite,  
439 calcite, and quartz, which is characteristic for the greenschist facies of metamorphism.

440

#### 441 **4.1.3. Geochronology: U-Pb dating of the Great Taghdout sill**

442 The zircon grains from the sample TA1 range in length from 120 to 150  $\mu\text{m}$ . Cathodoluminescence  
443 (CL) images (Figure S1 of supplementary data) show that most zircons are stubby sub-euhedral to  
444 euhedral, with a typical oscillatory zoning within a large uniform central zone, followed by much  
445 finer, poorly resolvable oscillatory zoned bands. Some grains lack any zoning (Figure S1 of  
446 supplementary data). We analyzed 18 spots on 13 zircon grains (Table S1 of supplementary data).

447 High Th/U ratios (between 1.18 and 3.38) support a magmatic origin for the zircons (Fig. 8) (cf.  
448 Möller et al., 2003; Hoskin and Schaltegger 2003; Linnemann et al., 2007). Most of the analyzed  
449 spots with high U and Th concentrations show evidence for significant Pb loss. On the Wetherill  
450 diagram, these thirteen analyses are aligned along a discordia that shows two intercepts with the  
451 concordia (Fig. 9A): an upper intercept with an age of  $1639 \pm 34$  Ma and a lower intercept with an  
452 age of  $688 \pm 65$  Ma. The age of  $1639 \pm 34$  Ma is interpreted to reflect the time of magmatic  
453 emplacement of the sill, whereas the  $688 \pm 65$  Ma age is thought to correspond to the time of a  
454 metamorphic and /or thermal overprint event. However, these intercept ages have a very high Mean  
455 Square Weighted Deviation (MSWD = 6.8), making this age poorly constrained. In contrast,  
456 analyses with low Th and U concentrations (Th <1000 ppm, spots 2.1, 7.1, and 13.1) produced  
457 nearly concordant  $^{207}\text{Pb}/^{206}\text{Pb}$  ages. We therefore calculated the weighted mean average  $^{207}\text{Pb}/^{206}\text{Pb}$   
458 age for those spots characterized by low Th and U concentrations, obtaining a mean  $^{207}\text{Pb}/^{206}\text{Pb}$  age  
459 of  $1676 \pm 37$  Ma (n = 3, MSWD = 3.1; Fig. 9B).

460

#### 461 4.1.4. Geochemical characteristics

462 The major and trace element composition of previously studied and new samples (TA1 and TA2)  
463 are listed in Table 2. On the total alkali versus silica diagram (Fig. 10A), the studied samples TA1  
464 and TA2 and the ca. 1.6 Ga doleritic dykes and sills previously analyzed by Youbi et al. (2013) plot  
465 in the subalkaline basalt field. These additional dolerites are: DZ 36 (dated at  $1640 \pm 2$  Ma and  
466  $1641 \pm 3$  Ma; Kouyaté et al., 2013) and DZ 40 (dated at  $1656 \pm 9$  Ma; Kouyaté et al., 2013) of the  
467 Zenaga Inlier (close to Tazenakht City), and doleritic dyke (AMHA069 dated at  $1654 \pm 16$  Ma;  
468 Kouyaté et al., 2013) from the Agadir Melloul Inlier. This subalkaline affinity is also evident from  
469 Nb/Y ratios that are below 0.67 (cf. Pearce and Cann, 1973; Winchester and Floyd, 1977).  
470 Additionally, their non-orogenic tholeiitic character is shown in the Ti vs. V diagram of Shervais  
471 (1982) (Fig. 10B) with the data plotting exclusively within the Continental Flood Basalt (CFB) and  
472 Mid-Ocean Ridge Basalt (MORB) fields ( $20 < V/\text{Ti} < 50$ ). When plotted on a multi-element

473 variation diagram (spidergram; Fig. 11), the dated dykes display patterns similar to some  
474 continental tholeiites such as those of the Central Atlantic Magmatic Province (CAMP) with a  
475 moderate enrichment in large-ion lithophile elements (LILE) and distinct positive Pb and negative  
476 Nb anomalies (Bertrand et al., 1982; Youbi et al., 2003; Marzoli et al., 2004, 2019; Tegner et al.,  
477 2019). Nb anomalies have been related to crustal contamination or the influence of a subduction  
478 zone, which was either active during magma generation or earlier metasomatized lithospheric  
479 mantle with which asthenospheric magma interacted during ascent (e.g. Thompson et al., 1982;  
480 Dupuy and Dostal, 1984; Coish and Sinton, 1992; Dorais et al., 2005). Furthermore, the negative  
481 Nb anomaly is accompanied by negative P anomaly; this pattern is similar to some low-Ti CFB  
482 tholeiitic magmas contaminated by continental crust. This and the relatively high Ti/V ratios (see  
483 Fig. 10B) point to a non-orogenic origin of the mafic magmas, which were contaminated with  
484 continental crust during magma ascent/emplacement (Ikenne, 1997; Hafid, 1999; Hafid et al., 2001;  
485 El Aouli et al., 2004; 2010; El Bahat et al., 2013; Youbi et al., 2013; El Bahat et al., 2017).

## 486 **4.2. Geochronology: Detrital Zircon Study of the lower Anti-Atlas Supergroup**

### 487 **4.2.1 Metasiltstone of the Tasserda Formation: sample TG-15-1**

488 The sample TG-15-1 (Table 1) is purple, coarse-grained quartz-rich metasiltstone collected from the  
489 Tasserda Formation near the village of Taghadout (Figs. 3 and 4). The Tasserda Formation consists  
490 of thick- to medium-bedded quartz arenites with few channels filled with conglomerates. The main  
491 mineral components of siltstone are quartz and feldspar embedded in a sericite-rich matrix (Fig.  
492 12A).

493 The zircon grains of this sample exhibit mean lengths and widths of 120 and 60  $\mu\text{m}$ , respectively. In  
494 most cases, they are sub-euhedral to euhedral in shape. The internal texture of zircon grains shows  
495 oscillatory concentric, sector, and banded zoning (Figure S2 of supplementary data). The Th/U ratio  
496 ranges from 0.03 to 1.15 (Fig. 8). About 83 % of the analyzed zircons have Th/U ratios larger than  
497 0.1, potentially indicating that they were derived from intermediate to felsic igneous rocks (cf.  
498 Möller et al., 2003; Hoskin and Schaltegger 2003; Linnemann et al., 2007). The other 17 % have a

499 Th/U ratio below 0.1 and could be of metamorphic origin. From the total of 99 measured spots, 74  
500 yielded less than 10 % discordant ages (Table S2 of supplementary data). About 99 % of the zircon  
501 population is of Paleoproterozoic age (26 % are Rhyacian and 73 % are Orosirian), ranging from  
502  $2294 \pm 32$  Ma to  $1831 \pm 35$  Ma, with a prominent peak at around 2022 Ma and three minor peaks at  
503 about 2294, 2142, and 1870 Ma. One remaining analysis, representing 1% of the concordant ages of  
504 this sample, yielded an age of  $2653 \pm 32$  Ma (Fig. 13A, B). The dominance of the Paleoproterozoic  
505 ages indicates that almost all of the detrital zircon grains were derived from the Paleoproterozoic  
506 (Eburnean) crust. The seven youngest zircon grains yielded a concordia age of  $2049.3 \pm 4.6$  Ma,  
507 which constrains the maximum depositional age of the Tasserda Formation (Fig. 13A, B).

#### 508 **4.2.2. Quartzites of the Agoumy Formation: sample BA21**

509 The sample BA21 (Table 1) is a medium-grained quartzite collected from the Agoumy Formation  
510 at the Tachdamt section (Figs. 3 and 5). The Agoumy Formation consists of thick- to medium-  
511 bedded quartzites with few conglomerates nested in channels.

512 The quartzites are dominated by quartz that is mainly subrounded. Some quartz grains show  
513 overgrowth rims. The rock is moderately sorted with a matrix that consists of silt-sized quartz and  
514 mica minerals such as sericite. Intergranular voids are filled with secondary Fe-Ti oxides (Fig.  
515 12B).

516 Zircons from this sample have a mean length and width of 110 and 70  $\mu\text{m}$ , respectively. Zircon  
517 grains are mainly sub-rounded to rounded in shape with some euhedral crystals (Figure S3 of  
518 supplementary data). Most of them show oscillatory zoning, suggesting that they are of igneous  
519 origin. This is corroborated by their Th/U ratios ranging from 0.19 to 1.50 (Fig. 8). The internal  
520 textures include oscillatory zoning, planar zoning, banded zoning, and a minor portion with sector  
521 zoning. A total of 55 spots have been analysed (Table S3 of supplementary data), of which 50  
522 yielded 90 to 110% concordant ages. About 84% are Paleoproterozoic (Statherian - 4%, Orosirian -  
523 24%, Rhyacian - 54%, and Siderian - 2%), 12% of all ages are Neoproterozoic and 4% are  
524 Mesoarchean ( $2869 \pm 8$  Ma and  $2857 \pm 32$  Ma). The Neoproterozoic ages range from  $2735 \pm 11$  to

525 2531 ± 18 Ma, while the Paleoproterozoic ages range from 2498 ± 14 to 1786 ± 20 Ma. The  
526 probability-density plot is dominated by peaks at ca. 2205, 2080, 2035, and 1812 Ma. The six  
527 youngest grains, ranging from 1834 ± 24 Ma to 1786 ± 20 Ma, combined yield a concordia age of  
528 1818 ± 6.4 Ma (MSWD = 1.5; probability = 0.22), which likely represents the maximum  
529 depositional age for the Agoumy Formation (Fig. 13C, D).

#### 530 **4.2.3. Quartzites of the Tachdamt Formation: sample ALBL29**

531 The sample ALBL29 (Table 1) is a medium- to coarse-grained, poorly sorted quartz arenite from  
532 the lowermost part of the Tachdamt Formation at the Bleida section (Figs. 3 and 5). This formation  
533 consists of basaltic lava flows and interbedded volcanic breccias, lapilli tuffs, and volcanoclastic  
534 sediments. The mineral assemblage mainly consists of quartz with a small amount of feldspar and a  
535 matrix of silt-sized quartz and mica minerals (Fig. 12C).  
536 Zircon grains from this sample exhibit a mean length and width of 160 and 90 μm, respectively.  
537 The grains are sub-rounded and euhedral (Figure S4 of supplementary data). The internal textures  
538 predominantly show concentric zoning, with few grains lacking any zoning. The Th/U ratio ranges  
539 from 0.01 to 1.26, with a dominant population of likely igneous origin (98%; 0.1 to 1.26) and a  
540 minor population with a potentially metamorphic origin (2%; 0.01 to 0.1; Fig. 8). Out of 96  
541 measured spots (Table S4 of supplementary data), 92 yielded less than 10% discordant ages (Fig.  
542 14A, B). About 54% of the ages are Mesoproterozoic (Calymmian - 22%, Ectasian - 17%, and  
543 Stenian - 17%), 41.3% of the ages are Paleoproterozoic (Siderian - 4%, Rhyacian - 8%, Orosirian -  
544 18%, and Statherian - 11%), 2% of the ages are Neoproterozoic (Tonian), 1% of the ages are  
545 Mesoarchean, and 1% of the ages are Neoproterozoic. Neoproterozoic and Mesoarchean ages are each  
546 represented by one analysis at 2577 ± 53 and 2969 ± 48 Ma, respectively. Paleoproterozoic ages  
547 range from 2482 ± 66 to 1603 ± 175 Ma with two prominent peaks at ca. 1978 and 1752 Ma. The  
548 Mesoproterozoic ages range from 1572 ± 55 to 1043 ± 12 Ma with four prominent peaks at ca.  
549 1525, 1280, 1208, and 1060 Ma. The Neoproterozoic age at 822 ± 6 Ma is only 92% concordant  
550 and has a very low Th/U ratio of 0.01 indicating some overprint of the zircon grain. This grain is



551 therefore not used to calculate a maximum depositional age. The inferred maximum age of  
552 deposition is set by a concordia age derived from the youngest population of concordant grains at  
553  $1060 \pm 5.5$  Ma (four grains), and  $930 \pm 12$  Ma, a 100% concordant single grain analysis with a  
554 Th/U ratio of 1.10 (Fig.14A, B).

#### 555 **4.2.4. Metasandstone of the Bleida Formation: sample BA24**

556 The sample BA24 (Table 1) is a fine-grained, moderately sorted metasandstone from the middle  
557 part of the Bleida Formation (Figs. 3 and 5) at the Bleida section. The Bleida Formation consists of  
558 up to 600 m thick siliciclastic deposits interbedded with volcanoclastics. The sample contains  
559 quartz, plagioclase, and lithic fragments with matrix bearing clays, sericite, muscovite, and chlorite.  
560 The grains are subangular to subrounded indicating that the sediment is immature (Fig. 12D).

561 The zircon grains of this sample exhibit a mean length and width of 125 and 85  $\mu\text{m}$ , respectively,  
562 and most of them are sub-rounded to sub-angular in shape. The internal texture of the zircons  
563 revealed by CL images comprise concentric and sector zoning, and some unzoned grains (Figure S5  
564 of supplementary data). The Th/U ratio ranges from 0.18 to 1.31, indicating a magmatic origin of  
565 the zircon grains (Fig. 8). Of 44 measured spots (Table S5 of supplementary data), 40 yielded less  
566 than 10% discordant ages. A total of 93% exhibits Paleoproterozoic ages (Siderian - 7%, Rhyacian -  
567 49%, and Orosirian - 37%), Mesoarchean, Mesoproterozoic, and Neoproterozoic populations are  
568 each 2%. The Mesoarchean is represented by one age of  $2852 \pm 7$  Ma, while the Paleoproterozoic  
569 ages range from  $2487 \pm 6$  to  $1841 \pm 12$  Ma with one prominent peak at ca. 2075 Ma and three minor  
570 peaks at ca. 2158, 1960, and 1852 Ma. The only Mesoproterozoic age is  $1276 \pm 14$  Ma and the  
571 Neoproterozoic age is  $977 \pm 10$  Ma. Two youngest grains, overlapping within errors, give a  
572 concordia age of  $1855 \pm 6.5$  Ma (Fig. 14C, D).

573 In summary, the inferred maximum depositional ages based on concordia diagrams for the  
574 sedimentary succession of the lower Anti-Atlas Supergroup are  $2049.3 \pm 4.6$  Ma for the  
575 Tasserda Formation,  $1818 \pm 6.4$  Ma for the Agoumy Formation,  $1060 \pm 5.5$  Ma for the Tachdant

576 Formation, and  $1855 \pm 6.5$  Ma for the Bleida Formation (Table 1). However, the latter formations  
577 could be younger (see Section 5.3)

578

## 579 **5. Discussion**

### 580 **5.1 The 1.68-1.64 Ga Zenaga Event**

581 The SHRIMP U-Pb geochronologic data for the Great Taghdout sill presented here shows a  
582 weighted mean average  $^{207}\text{Pb}/^{206}\text{Pb}$  age of  $1676 \pm 37$  Ma interpreted to date the magmatic  
583 emplacement of the sill (Fig. 9B). This age is similar within error to those obtained for other  
584 dolerite dykes and sills in the studied area (Zenaga Inlier), as well as in other inliers of the Anti-  
585 Atlas Belt (e.g. Agadir Melloul Inlier). Kouyaté et al. (2013) obtained ages of  $1640 \pm 2$  Ma and  
586  $1641 \pm 3$  Ma (U-Pb baddeleyite) on the sill (DZ 36), which intrudes the Eburnean basement in the  
587 Zenaga Inlier (Fig. 4). They also obtained an age of  $1656 \pm 9$  Ma (U-Pb baddeleyite) on another sill  
588 (DZ40), which is connected to a sill intruding the lowermost part of the lower Anti-Atlas  
589 Supergroup of the Zenaga Inlier (the Tasserda-Taghatine Group) and an age of  $1654 \pm 16$  Ma (U-Pb  
590 baddeleyite) on a large NE-trending dyke, which crosscuts the Eburnean basement of the Agadir  
591 Melloul Inlier. More recently, Youbi et al. (2019) dated two sills intruding quartzites of the Lkest  
592 Group (equivalent of the Oumoula Formation) in the Kerdous Inlier. One of the sills yielded an  
593 upper intercept date of ca. 1640 Ma, constrained with ID-TIMS data for three zircon grains. LA-  
594 ICP-MS data for the same sample yielded a weighted mean average  $^{207}\text{Pb}/^{206}\text{Pb}$  age of ca. 1650 Ma,  
595 identical within analytical uncertainties to the ID-TIMS date. Only a single baddeleyite grain has  
596 been analyzed from the second sill, producing a similar crystallization age of ca. 1650 Ma. All of  
597 these ages overlap within error with the newly obtained age from this study.

598 These ca. 1.68-1.64 Ga ages match the Rb-Sr dates, which were obtained for granitic rocks from the  
599 Anti-Atlas Belt; the Timrharbine granite (Zenaga Inlier), Tasserhert granite (Kerdous Inlier), and  
600 Oued Chaiba granite (Bas Drâa Inlier) dated at  $1650 \pm 50$ ,  $1680 \pm 35$ , and  $1640 \pm 32$  Ma,  
601 respectively (Charlot, 1976; 1982). This suggests an event of bimodal magmatism during the late

602 Paleoproterozoic, post-dating the ca. 2.04 Ga Tagragra of Tata LIP extensional event (Kouyaté et  
603 al., 2013; Youbi et al., 2013), and confirms the polycyclic character of magmatism in the Anti-Atlas  
604 Belt, previously inferred in Kerdous and Tagragra of Akka inliers (Gasquet et al., 2004; Barbey et  
605 al., 2004). These ca. 1650 Ma ages are interpreted to belong to a single LIP event, which is termed  
606 the Zenaga Event (Kouyaté et al., 2013; Youbi et al., 2013; Ernst, 2014).

607 Globally, magmatism at ca. 1640-1656 Ma is also observed on other cratons (Fig. 15). In eastern  
608 and southwestern Greenland, the Melville Bugt dyke swarm extends for about 1000 km and is  
609 precisely dated at 1622-1635 Ma by the U-Pb method ( $1622 \pm 3$ ,  $1629 \pm 1$ ,  $1632 \pm 1$ , and  $1635 \pm 3$   
610 Ma; Halls et al., 2011). Following upon previous work on the ca. 1630 Ma Melville Bugt dyke  
611 swarm along the NW coast of Greenland, Klausen and Nilsson (2019) confirmed its proposed  
612 continuation to the SE Greenland, indicating an extraordinary >2000 km long Trans-Greenlandic  
613 dyke swarm of Laurentia. This correlation was not only based on the swarm's continuous pattern,  
614 but also on a matching  $1630 \pm 4$  Ma U-Pb baddeleyite age determination (Klausen and Nilsson,  
615 2019) and similar whole-rock compositions as well as distinctive outcrop characteristics. In  
616 Laurentia, the bimodal Narakay Volcanic Complex in the Dease Arm of the Great Bear Lake,  
617 associated with deposition in the intracontinental Hornby Bay Basin, NWT, Canada, yielded TIMS  
618 U-Pb zircon age of  $1663 \pm 8$  Ma (Bowring and Ross, 1985).

619 Corresponding magmatism on Baltica includes the ca. 1.64 Ga NW-trending dyke swarms of Häme,  
620 Sipoo, and Suomenniemi and the related rapakivi granite complexes exposed in the Svecofennian  
621 Domain of the Fennoscandian Shield in southeast Finland (Vaasjoki et al., 1991; Salminen et al.,  
622 2017). The most extensive of these dyke swarms is the  $1642 \pm 2$  Ma Häme diabase dyke swarm (U-  
623 Pb ID-TIMS baddeleyite age; Salminen et al., 2017).

624 In Southern Siberia, the Nersa Sill Complex from the Biryusa Block of the Irkutsk Promontary has  
625 a  $1641 \pm 8$  Ma U-Pb baddeleyite age (Ernst et al., 2016). Other potential matches are rejected. The  
626 Khibelen dykes of the Baikal Uplift, Siberia have a  $1674 \pm 29$  Ma Sm-Nd pyroxene, plagioclase,  
627 and whole-rock isochron age (Gladkochub et al., 2010). However, the Khibelen dykes are on a

628 trend from the SW-trending Chaya dyke subswarm, which has a  $1752 \pm 3$  Ma U-Pb baddeleyite age  
629 (Gladkochub et al. 2010; Ernst et al. 2016), and so the Chaya and Khibelen dykes likely represent a  
630 single ca. 1750 Ma swarm. The type section for the Early Riphean, a major Russian stratigraphical  
631 division, in the South Urals Mountains, southeast Baltica, contains poorly-dated trachybasalts of the  
632 Navysh Subformation of the Ai Formation, which were originally thought to be of similar age.  
633 However, based on more recent high-precision SHRIMP U-Pb zircon dating these volcanics are  
634  $1752 \pm 11$  Ma in age (Krasnobaev et al., 2013a, b; Puchkov et al. 2014).

## 635 **5.2. Stratigraphic Implications Arising From the U-Pb Dating of the Great Taghdout Sill**

636 Based on the previously published data, the age of the lower Anti-Atlas Supergroup is bracketed  
637 between ca. 2032 and 647 Ma (see section 2.3.). The new U-Pb SHRIMP zircon age of  $1676 \pm 37$   
638 Ma obtained for the herein named Taghdout sill provides a new constraint on the minimum age for  
639 the lowermost part of the lower Anti-Atlas Supergroup in the Zenaga Inlier. The Taghdout sill  
640 intrudes into the lower Anti-Atlas Supergroup and is consequently younger than the intruded  
641 sedimentary rocks (that is the Tasserda and Ifrane-n-Taghatine formations) and most likely older  
642 than the unconformably overlying Oumoula Formation. The new age indicates that the lowermost  
643 part of the lower Anti-Atlas Supergroup was likely deposited during the late Paleoproterozoic (2.03-  
644 1.64 Ga). It could potentially be related to extension during the post-Eburnean orogenic collapse  
645 and if so should be close in age to ca. 2.0 Ga. In contrast, the much younger ca. 1.67 Ga Taghdout  
646 sill, likely related to the ca. 1.65 Ga Zenaga LIP, is a part of the mantle plume event/s that occurred  
647 at the late stage of the Nuna / Columbia supercontinent assembly (Fig. 15).

648 The new  $1676 \pm 37$  Ma age for the sill in the Zenga Inlier is equivalent within uncertainty to the ca.  
649  $1706 \pm 7$  Ma age obtained for mafic intrusions crosscutting basal quartzites in the Igherm Inlier  
650 (Ikenne et al., 2017) and the ca. 1650 Ma age of the thick sills intruding the basal part of the lower  
651 Anti-Atlas Supergroup in the Kerdous Inlier (Youbi et al., 2019). Together, these ages indicate  
652 associated extension. Therefore, the post-Eburnean hiatus of one billion years on the WAC is no  
653 longer valid and the pre-Pan-African sediments and magmatic events are preserved along the

654 northern margin of the craton. The 1640-1680 Ma (this study; Youbi et al., 2019) and the ca. 1706  
655 Ma (Ikenne et al., 2017) magmatic events indicate that extension predated deposition of the thick  
656 quartzites of the Oumoula Formation, but postdated accumulation of the Tasserda and Ifrane-n-  
657 Taghatine formations. Further geochronological work is required to determine whether the ca. 1650  
658 Ma, ca. 1676 Ma, and ca. 1706 Ma intrusions represent separate intraplate events on the WAC or  
659 belong to a single long-lived 1.70-1.65 Ga LIP event across the WAC and potentially representing  
660 an event extending into Baltica and Laurentia (Fig. 15). Regardless, all these ages for mafic sills are  
661 additional evidence that the lower part of the lower Anti-Atlas Supergroup has an upper  
662 Paleoproterozoic age and, therefore, nearly one billion years older than previously thought.

663

### 664 **5.3. Constraints From Detrital Zircons on the Depositional Age of the Lower Anti-Atlas**

#### 665 **Supergroup**

666 The youngest detrital zircons from the sample TG-15-1 (the Tasserda Formation) are ca. 2049 Ma  
667 old (Fig. 13A, B), providing a maximum depositional age for the formation. This is in agreement  
668 with the geological context of the lower Anti-Atlas Supergroup: (i) the age of the succession is  
669 bracketed between  $2032 \pm 5$  Ma and  $647 \pm 2$  Ma; (ii) the  $1676 \pm 37$  Ma Taghdout sill indicates that  
670 the intruded sedimentary rocks (that is the Tasserda and Ifrane-n-Taghatine formations) are older  
671 than  $1676 \pm 37$  Ma, consistent with the minimum age constraints of ca. 1650 Ma in the Kerdous  
672 Inlier (Youbi et al., 2019) and ca. 1710 Ma in the Igherm Inlier (Ikenne et al., 2017).

673 The six youngest zircon grains in the sample BA21 (the Agoumy Formation) yielded a late  
674 Paleoproterozoic concordia age of  $1818 \pm 6.4$  Ma (Fig. 13C, D), providing the maximum age for  
675 deposition of this formation. The U-Pb detrital zircon data obtained from this sample are consistent  
676 with the U-Pb zircon data for quartzites of the Mimount Formation in the Siroua Inlier (*sensu*  
677 Thomas et al., 2002), which is the stratigraphic equivalent of the Oumoula Formation (*sensu*  
678 Bouougri and Saquaque, 2004) of the lower Anti-Atlas Supergroup. For the Mimount Formation,  
679 Abati et al. (2010) obtained LA-ICP-MS  $^{207}\text{Pb}/^{206}\text{Pb}$  zircon ages from ca. 2936 to ca. 1809 Ma, with

680 major peaks at ca. 2200 and 2086 Ma and a number of minor peaks (ca. 2005, 1950, and 1818 Ma).  
681 The youngest zircon in the Oumoula Formation gave a concordant age of  $1809 \pm 15$  Ma (Abati et  
682 al., 2010), while the youngest population age is ca. 1818 Ma, providing a maximum age for  
683 deposition of the Oumoula Formation. For the Mimount Formation, Walsh et al. (2012) obtained a  
684 range of SHRIMP  $^{207}\text{Pb}/^{206}\text{Pb}$  zircon ages from about 2182 to 1987 Ma, with a major peak at ca.  
685 2050 Ma. The similarity between detrital zircon age distribution in our sample BA21 from the  
686 Agoumy Formation and that of the Mimount Formation in Abati et al. (2010) and Walsh et al.  
687 (2012) indicates a similar provenance for the Mimount (= Oumoula) and Agoumy formations.  
688 The lack of ca. 1.67 Ga zircon grains in samples from the Mimount and Agoumy formations is  
689 worth a note given the prominence of the 1650 Ma Zenaga event (see Section 5.1). However, mafic  
690 melts are usually poor in zircons; this may explain their relative absence in these formations.  
691 The youngest detrital zircon grains from the Tachdamt Formation which overlap within errors are  
692 ca. 1060 Ma old (Fig. 14A, B). However, there are two younger single zircon grains with ages at  
693  $930 \pm 12$  and  $822 \pm 6$  Ma. The latter age has a concordance of only 92% and a very low Th/U ratio,  
694 therefore the maximum age of sedimentation is constrained by ca. 1060 Ma and ca. 930 dates,  
695 consistent with ca. 883 Ma U-Pb SHRIMP zircon ages for two tuff horizons interbedded in the  
696 basal part of the Tachdamt Formation in the Siroua Inlier (Youbi et al., 2018, Bouougri et al.,  
697 2020). The youngest concordant zircon dates that overlap within errors in the sample BA24 from  
698 the Bleida Formation yield a concordia age of  $1855 \pm 6.5$  Ma. The two youngest zircon grains  
699 provide dates at  $1276 \pm 14$  Ma and  $977 \pm 10$  Ma (Fig. 14C, D). However, Letsch (2018) reported a  
700 pronounced bimodal age distribution for a greywacke from the Bleida Formation in the Bou Azzer  
701 Inlier with a major Paleoproterozoic population (peaked at ca. 2060 Ma) and a minor  
702 Neoproterozoic one (with a prominent mode at ca. 690 Ma); the latter defines, along with the  $697.3$   
703  $\pm 2.2$  Ma maximum depositional age of the metasilstone from the most basal part of the Bleida  
704 Formation (Bouougri et al., 2020), a Cryogenian maximum depositional age for this formation

705 (although the minimum depositional age is younger than 883 Ma). These data imply a significant  
706 hiatus (697 to <883 Ma) in time of deposition between the Tachdamt and Bleida formations.

707 Both the Bleida and Tachdamt formations contain Mesoproterozoic detrital zircon ages (Fig. 14),  
708 however these ages are less abundant in the Bleida Formation, suggesting a change in the  
709 provenance consistent with a major hiatus in the time of deposition for these units. Importantly,  
710 detrital zircon ages indicate that the provenance for the lower Anti-Atlas Supergroup was  
711 overwhelmingly Paleoproterozoic and Mesoproterozoic in age with a much smaller contribution from  
712 Archean and Neoproterozoic sources.

#### 713 **5.4. Interpretation of Detrital Zircon Age Clusters**

714 Four major detrital zircon age clusters have been obtained in this study: 3.0-2.4, 2.3-1.74, 1.7-1.4,  
715 and 1.25-0.95 Ga (Fig. 16). The few Archean zircon grains, forming a poorly defined cluster, were  
716 likely derived from the Reguibat and Leo-Man shields (Rocci et al., 1991; Thiéblemont et al., 2004;  
717 Key et al., 2008; Koffi et al., 2020). The second cluster (2.30-1.74 Ga) reflects contribution from  
718 mafic and felsic magmatic intrusions emplaced during the Eburnean Orogeny and its aftermath  
719 along the northern margin of the WAC (Abouchami et al., 1990; Boher et al., 1992; Egal et al.,  
720 2002; Schofield et al., 2006; Kouyaté et al., 2013; Youbi et al., 2013 Baratoux et al., 2011; 2019;  
721 McFarlane et al., 2019). The third cluster (1.7-1.4 Ga) is likely linked to the emplacement of dykes  
722 and sills accompanying the protracted rifting and the breakup of the Nuna / Columbia  
723 supercontinent (Youbi et al., 2013; Evans et al., 2013). Zircons of the fourth cluster (1.25-0.95 Ga)  
724 are common in the early Neoproterozoic sedimentary rocks. It is thought that these zircons were  
725 derived from the 1.25-0.95 Ga Grenvillian arcs (Rino et al., 2008) that underwent uplift and erosion  
726 during and in the aftermath of the amalgamation of the Rodinia supercontinent, and they may have  
727 been transported for thousands of kilometres inside the supercontinent by rivers (Rainbird et al.,  
728 1997). Interestingly, metamorphic and intrusive units with these ages are common in the Sunsás  
729 Orogen (e.g. Sadowski and Bettencourt, 1996; Santos et al., 2008), which borders the Andean side  
730 of the Amazonian Craton (Fig. 17), and the Sveconorwegian Orogen of the northwestern part of

731 Baltica (e.g. Andersson et al., 2007; Christoffel et al., 1999; Cosca et al., 1998) that were adjacent  
732 to the WAC in the Rodinia supercontinent configuration.

### 733 **5.5. Speculations on the Provenance and Paleogeographic Reconstruction**

734 The main feature of our detrital zircon populations is the general absence or scarcity of  
735 Mesoproterozoic ages (Figs. 13 and 14), which is consistent with the earlier studies that argued for a  
736 tectonomagmatically quiescent period between ca. 1.7 and 1.0 Ga (e.g. Ennih and Liégeois, 2008).

737 The absence or scarcity of Mesoproterozoic ages has been considered diagnostic in distinguishing  
738 the WAC provenance as a possible source (e.g. Nance and Murphy, 1994; Friedl et al., 2000;  
739 Linnemann et al., 2004; Abati et al., 2010; Bahlburg et al., 2010; Gärtner et al., 2013, 2016, 2017).

740 How could this be reconciled with the U-Pb geochronological evidence for major mafic magmatic  
741 events at ca. 1.75, 1.65, and 1.41-1.38 Ga (e.g. Youbi et al., 2013) ? The simplest explanation is that  
742 the mafic magmatism does not produce a significant detrital zircon population, and hence these  
743 major LIP events are not reflected in the detrital zircon record.

744 Samples from the upper part of the lower Anti-Atlas Supergroup (the Tachdamt and Bleida  
745 formations) contain Mesoproterozoic detrital zircons. Considering that in most reconstructions of  
746 the Nuna-Columbia and Rodinia supercontinents, the Amazonian Craton and Baltica were adjacent  
747 to the WAC (e.g. Hoffman, 1991; Dalziel, 1991, 1997; Weil et al., 1998; Evans, 2013; Li et al.,  
748 2013; Johansson, 2009; 2014; Merdith et al., 2017) and had their sources of Mesoproterozoic  
749 zircons far from their borders with the WAC (Fig. 15), both of these cratons could have been  
750 potential sources of detrital zircons as long as a long-distance fluvial transport was involved.

751 However, the Amazonian Craton (Tassinari and Macambira, 1999) is characterized by a continuous  
752 range of Mesoproterozoic ages since it undergone several late Paleoproterozoic to Mesoproterozoic  
753 orogenic events, and igneous and metamorphic units with the ages ranging from 2000 to 1000 Ma  
754 are common there (e.g. Santos et al., 2000; Tassinari et al., 2000). While the sample ALB29 is  
755 indeed characterized by a continuous range of Mesoproterozoic ages, the sample BA24 has only  
756 few Mesoproterozoic zircons. Tantalizingly, similar to the sample ALB29 range of



757 Mesoproterozoic detrital zircon ages was observed for the ca. 1.1 Ga Polissya Series in the western  
758 part of the Ukrainian Shield (SW part of Baltica (Sarmatia); Shumlyanskyy et al., 2015) that was  
759 positioned adjacent to the WAC in the configuration of Fig. 15, but equally lacks sources for the  
760 Mesoproterozoic detrital zircons. A long-distance transport from NW Baltica has thus been inferred  
761 for this mode in detrital zircon ages based on the combined ages and Hf isotope systematics.  
762 Mesoproterozoic detrital zircon ages were also found in the Neoproterozoic sedimentary units in the  
763 Mauritanides and in the Taoudeni Basin (Straathof, 2011; Nicoll et al., 2010; Bradley et al., 2015;  
764 see for compilation Gärtner et al., 2017), indicating long-distance fluvial transport of  
765 Mesoproterozoic zircons in the aftermath of the Grenville Orogeny (cf. Rainbird et al., 1997).

#### 766 **5.6. Refining Proterozoic Stratigraphy of the Anti-Atlas Belt**

767 Figure 18 shows a tentative correlation of the lower Anti-Atlas Supergroup (*sensu* Thomas et al.,  
768 2004) among the Anti-Atlas inliers. Based on our own geochronologic data and on published U-Pb  
769 geochronology, as well as on lithostratigraphy, we propose to abandon the Lkest-Taghdout Group  
770 name since it combines genetically unrelated successions deposited in different, superimposed  
771 basins recording different tectonic settings and bracketed by long-lasting hiatuses. Instead, we  
772 propose to use the lower Anti-Atlas Supergroup as the highest stratigraphic hierarchy unit for the  
773 Proterozoic, pre-Pan-African successions of the Anti-Atlas Belt that can be further subdivided in  
774 five unconformity-bounded lithostratigraphic units.

775 (1) The lower lithostratigraphic unit is Paleoproterozoic in age (ca. 2030-1706 Ma) and comprises  
776 the Tasserda and Ifrane-n-Taghatine formations, named here the Tasserda-Taghatine Group. Our  
777 maximum depositional age for the Tasserda Formation (<2049 Ma) agrees well with the only other  
778 detrital zircon study for the Tasserda and Ifrane-n-Taghatine formations by Letsch (2018), who  
779 inferred <2038-1975 Ma maximum depositional age for this succession. The lower group is herein  
780 related to the Eburnean Orogen collapse at ca. 2.0 Ga and deposition in extensional intracratonic  
781 basins.

782 (2) The second lithostratigraphic unit is also Paleoproterozoic in age (ca. 1745-1650 Ma) and  
783 comprises the Oumoula (= Mimount) Formation. In the Ifni Inlier, the stratigraphic equivalent of  
784 the lower Anti-Atlas Supergroup, the “Groupe des Quartzites” (Jeannette et al., 1981), consists of  
785 quartzites and metasilstones that have been subdivided into three units that unconformably overlie  
786 the Eburnean basement. U-Pb dating of the Eburnean basement in this inlier yielded  $2042 \pm 5$  Ma  
787 and  $2026 \pm 7$  Ma ages for the Alouzad and Sahel granites, respectively (Schulte et al., 2016). The  
788 detrital zircons from the Ifni Inlier quartzites yielded ages ranging between ca. 2020 and 1745 Ma  
789 with no evidence of the Mesoproterozoic ages (Benziane et al., 2016). We tentatively correlate the  
790 Ifni Inlier succession with the Oumoula Formation based on a similar detrital zircon distribution  
791 patterns.

792 In the Agadir Melloul and south Siroua inliers, the stratigraphic equivalents of the lower Anti-Atlas  
793 Supergroup are the Iguiguil-Tizi n’Taghatine and Ahmamad formations, that both unconformably  
794 overlie the Eburnean granites and diorites with ca. 2.04-2.03 Ga U-Pb zircon ages (Choubert et al.,  
795 1992; Blein et al., 2014b). The Iguiguil-Tizi n’Taghatine Formation (Fig. 18) contains the Jbel  
796 Iguiguil quartzites that are sedimentologically similar to the quartzites of the Oumoula Formation  
797 (*sensu* Bouougri and Saquaque, 2004) and Mimount Formation (*sensu* Thomas et al., 2002; 2004).  
798 The Ahmamad Formation (Fig. 18) consists of quartzites and locally developed carbonate rocks  
799 exposed along the strike-slip faults in the southeastern part of the Azaghar-Izazen Inlier (Blein et  
800 al., 2014b and references therein). Both the Iguiguil-Tizi n’Taghatine and Ahmamad formations  
801 appear to have a similar age, although spatial (and structural) relationships are not clear (Choubert  
802 et al., 1992; Blein et al., 2014b). The Ahmamad Formation is similar to the Oumoula Formation in  
803 Jbel Iguiguil. The Iguiguil-Tizi n’Taghatine Formation is intruded by  $570 \pm 7$  Ma dyke swarms and  
804 sill complexes of dolerites and gabbros (U-Pb zircon age; Blein et al., 2014b).

805 The Agadir Melloul Inlier has several ridges with nearly vertical conglomeratic quartzite beds  
806 sandwiched in Paleoproterozoic granites. Detrital zircon dates from the quartzites yielded the major

807 peak at ca. 2.0 Ga and the youngest ages at ca. 1.8-1.7 Ga (Soulaimani et al., 2019), similar to the  
808 distribution trend shown by detrital zircons from the the Oumoula Formation.

809 In the Kerdous Inlier, the stratigraphic equivalent of the lower Anti-Atlas Supergroup is the Jbel  
810 Lkest Group and includes the whole section of the “PII” quartzites in the Anzi District  
811 (Hassenforder, 1987; O’Connor et al., 2010). The thick, hematite-rich quartzite sequence is not  
812 subdivided because of its monotonous nature and the lack of preservation of original textures due to  
813 extensive recrystallization, however it seems to correspond to the same depositional facies as the  
814 Oumoula Formation (*sensu* Bouougri and Saquaque, 2004) and Mimount Formation (*sensu* Thomas  
815 et al., 2002; 2004) quartzites. The great thickness of this formation and its homogeneity imply a  
816 constant supply of sediment sustained for a long period of time balanced by a continuous  
817 subsidence. Letsch (2018) constrained the maximum depositional age for the quartzite to be 1780  
818 Ma based on U-Pb LA-ICP-MS detrital zircon data. As noted above (Section 5.2.), Youbi et al.,  
819 (2019) recently dated two mafic sills emplaced into the quartzite. One of the sills yielded an upper  
820 intercept date of ca. 1640 Ma, constrained by ID-TIMS data on three zircon fractions. LA-ICP-MS  
821 data for the same sample yielded a weighted mean average  $^{207}\text{Pb}/^{206}\text{Pb}$  date of ca. 1650 Ma,  
822 identical within analytical uncertainties to the ID-TIMS date. Only a single fraction of baddeleyite  
823 has been analyzed from the second sill, indicating a similar crystallization age of ca. 1650 Ma.

824 The unconformably bounded Oumoula Formation is thus bracketed in age between ca. 1745 and  
825 1650 Ma (Fig. 18). The data suggest that there was a pronounced change in the provenance across  
826 the unconformity between the Oumoula and the underlying Tasserda-Taghatine Group.  
827 Furthermore, the underlying units show a different structural pattern from the Oumoula Formation  
828 in the Zenaga Inlier (cf. Choubert et al., 1951; Brabers, 1988; Letsch, 2018), suggesting that they  
829 were deposited in different basins. Discontinuous distribution of the lower group formations and the  
830 Oumoula Formation, combined with geochemical composition of the 1.75-1.64 Ga mafic dykes and  
831 sills akin to continental flood basalts (Section 4.1.4.), indicate deposition of the Oumoula Formation  
832 at ca. 1.75-1.64 Ga in the compartmentalized extensional basins (rifts).

833 (3) The third lithostratigraphic unit is Paleoproterozoic to Neoproterozoic in age (< ca. 1706-1640  
834 Ma to > ca. 883 Ma). It incorporates the Ifarkhs-n-Tirsal, Wanimzi, Tamgarda, Agoumy, and Imi  
835 n-Tizi formations. This lithostratigraphic unit is named here the Tizi n'Taghatine Group. Our  
836 detrital zircon ages for the Agoumy Formation provide the first maximum depositional age for  
837 this succession (<1818 Ma), but unfortunately it does not improve independent age constraints for  
838 this group. Ca. 1650 Ma sills and dykes that intrude the underlying Oumoula Formation, but not the  
839 Tizi n'Taghatine Group provide the maximum depositional age and ca. 883 volcanics in the  
840 unconformably overlying Tachdamt Formation set the minimum depositional age for this  
841 succession. The succession was likely deposited in the intracratonic basin related to the rifting  
842 within the assembling Rodinia supercontinent.

843 (4) The fourth lithostratigraphic unit is the ca. 883 Ma Tachdamt Formation. The Tachdamt  
844 Formation consists of a succession of basaltic lava flows interbedded with subordinate pyroclastic  
845 rocks. Youbi et al. (2018) and Bouougri et al., (2018; 2020) dated two tuff horizons, interbedded in  
846 the basal part of the Tachdamt Formation in the Siroua Inlier, at  $885 \pm 5.9$  Ma and  $883 \pm 2.3$  Ma (U-  
847 Pb SHRIMP on zircon). The NE30°-trending Iguerda-Taïfast mafic dyke swarm in the Zenaga and  
848 Iguerda-Taïfast inliers and the coeval volcanics of the Tachdamt Formation are part of the ca. 885-  
849 883 Ma Iguerda-Taïfast Large Igneous Province (LIP) in the Anti-Atlas Belt of Morocco (Kouyaté  
850 et al., 2013; Söderlund et al., 2013; Youbi et al., 2013, 2018; Bouougri et al., 2020), related to the  
851 rifting within the Rodinia supercontinent. The formation was likely deposited in the rift basin at the  
852 late stage in the assembly of the supercontinent Rodinia, but it remains uncertain whether this basin  
853 evolved into a passive continental margin setting along the northern margin of the WAC. Similar-  
854 age LIP events have been recognized in the southern part of the WAC (Baratoux et al., 2019), and  
855 on the Tarim Terrane of North China (Wang et al., 2015; Zhang et al., 2019) and the Congo Craton  
856 (Armstrong et al., 2005; Johnson et al., 2007).

857 (5) The uppermost unit, the ca. 700 Ma Bleida Formation is characterized by fine-grained and thin-  
858 bedded siliciclastic turbidites, interbedded with minor basaltic lava flows, exhalative facies (jasper

859 and ironstone), and black shales. Depositional setting of the Bleida Formation and the minimum age  
860 of its detrital zircons point to deposition in a foreland basin developed in the lead of the ca. 660-680  
861 Ma Pan-African Orogeny (Thomas et al., 2004; Bouougri et al., 2020).

### 862 **5.7. Potential correlation with the Taoudeni Basin succession in Mauritania**

863 Our study indicates that the lower Anti-Atlas Supergroup records Mid-Proterozoic protracted and  
864 episodic deposition in intracontinental basins developed in association with extension and LIPs  
865 events on the WAC. Considering that these conditions, likely linked to far-distant tectonic stresses  
866 and emplacement of mantle plumes, could also have affected other parts of the WAC, we explore  
867 here correlation with another Mid-Proterozoic basin on the WAC, the Taoudeni Basin, located  
868 approximately 500 km to the south of the Anti-Atlas Belt (Fig. 1). It contains three unconformity-  
869 bounded successions in the ascending order: the Char, Atar, and Assabet el Hassiane groups.

870 The lower part of the sedimentary fill of the Taoudeni Basin, the Hodh Supergroup (Super-groupe  
871 du Hodh), is subdivided into the three unconformity-bound groups: the Char, Atar, and de l'Assabet  
872 el Hassiane (Moussine-Pouchkine and Bertrand-Sarfati, 1997; Benan and Deynoux, 1998). The  
873 mixed carbonate-siliclastic Char Group was deposited in a fault-bounded, compartmentalized basin  
874 developed on the deeply weathered Eburnean basement with aeolean and fluvial redbeds grading  
875 upsection to tidally influenced, marine siliclastic and carbonate deposits with some evidence  
876 (halite pseudomorphs, mudcracks and mudchips, dessication cracks) for warm and arid climate. The  
877 depositional and tectonic setting of the Char Group is similar to that of the Tasserda-Taghatine  
878 Group (lithostratigraphic unit 1 in the Anti-Atlas Belt). Similarly, both successions carry detrital  
879 zircons with youngest ages corresponding to the Eburnean basement with an additional contribution  
880 in case of the Zenaga Inlier of the Anti-Atlas Belt of slightly younger, ca. 2042-1975 Ma  
881 provenance (Bradley et al., 2015; Straathof, 2011; Letsch, 2018; this study). We provisionally  
882 correlate these successions and link the development of their corresponding basins to the post-  
883 Eburnean orogen collapse.

884 We recognize no correlative stratigraphic interval in the Taoudeni Basin to the ca. 1745-1650 Ma  
885 Oumoula Formation (lithostratigraphic unit 2 of the Anti-Atlas Belt), which we link to extension  
886 and LIP emplacement on the WAC at the final stage in the assembly of the Nuna supercontinent.  
887 Importantly, ca. 1750 and 1650 Ma mafic magmatism, linked with deposition of the Oumoula  
888 Formation, was not so far recognized in the Taoudeni Basin and the surrounding Reguibat Craton.  
889 The next potentially correlative stratigraphic intervals in the Anti-Atlas Belt and the Taoudeni Basin  
890 are the Tizi n'Tghatine (lithostratigraphic unit 3) and Atar groups, respectively. Both are mixed  
891 carbonate-siliciclastic successions deposited in intracratonic basins in warm and arid climate  
892 (Leblanc and Moussine-Pouchkine, 1994; Moussine-Pouchkine and Bertrand-Sarfati, 1997). Our  
893 youngest detrital zircon ages for the Agoumy Formation (ca. 1818 Ma) correspond well with the  
894 youngest detrital zircon ages reported for the Atar Group ( $1804 \pm 33$  Ma, Straathof, 2011; Nicoll et  
895 al., 2010; Bradley et al., 2015). Re-Os isochron ages for black shales from the Atar Group yielded  
896 the late Mesoproterozoic depositional age ( $1106 \pm 12$  Ma,  $1109 \pm 22$  Ma, and  $1105 \pm 37$  Ma;  
897 Rooney et al., 2010). These Mesoproterozoic ages are confirmed by the study of microfossils  
898 (Beghin et al., 2017) and chemostratigraphy (Kah et al., 2012). Carbonates of the the Tizi  
899 n'Tghatine Group have highly negative to near zero carbon isotope values (-5.0 to +1.5 ‰; Alvaro  
900 et al., 2014). In comparison, carbonates of the Atar Group show large carbon isotope oscillations  
901 from -6.0 to +4.0 ‰ (Fairchild et al., 1990; Shields et al., 2007; Kah et al., 2012) that are  
902 characteristic of the latest Mesoproterozoic to early Neoproterozoic seawater.

903 It is uncertain whether there is a stratigraphic equivalent of either the ca. 883 Ma Tachdamt  
904 Formation / Bleida formation (lithostratigraphic units 4 and 5) in the Taoudeni Basin, but the  
905 siliciclastic Assabel el Hassiane Group yielded youngest ages for detrital zircons of  $952 \pm 38$  Ma  
906 and  $965 \pm 29$  Ma (Straathof, 2011; Bradley et al., 2015), which are consistent with the youngest  
907 ages for detrital zircons from the Tachdamt Formation (see Fig. 18). More importantly, both the  
908 Tachdamt Formation and the Assabel el Hassiane Group show influx of Mesoproterozoic detrital  
909 zircons in the aftermath of the Grenville Orogeny. Similar pattern for the Laurentia has been related

910 to giant fluvial systems that delivered material from the Grenville Orogen inside the Rodinia  
911 supercontinent (Rainbird et al., 1997, 2012; Rainbird and Young, 2009). Kuznetsov et al. (2017)  
912 suggested the Rodinia supercontinent configuration ringed by continental arcs and with  
913 predominantly internal runoff that seems to fit well to the detrital zircon patterns of the WAC.

## 914 **6. Concluding remarks**

915 The main conclusions of our study are summarized below.

916 (1) The lower Anti-Atlas Supergroup was considered as a late Neoproterozoic passive continental  
917 margin sequence deposited during rifting and breakup on the northern margin of the WAC,  
918 culminating with opening of an oceanic basin between the northern edge of the WAC and an  
919 unknown terrane at ca. 760 Ma. Based on U-Pb geochronology data (this work and literature data)  
920 and lithostratigraphic constraints, we propose a new lithostratigraphic framework for the  
921 Proterozoic lower Anti-Atlas Supergroup of the Anti-Atlas Belt. U-Pb geochronologic data  
922 indicate that deposition of the lower Anti-Atlas Supergroup lasted from the middle Paleoproterozoic  
923 to late Neoproterozoic with significant hiatuses along unconformities. Five major, unconformity-  
924 bounded lithostratigraphic units are distinguished in the lower Anti-Atlas Supergroup. The lower  
925 unit is the ca. 2030-1706 Ma Tasserda-Taghatine Group and comprises the Tasserda and Ifrane-n-  
926 Taghatine formations. The second unit is the ca. 1745-1650 Ma Oumoula (Mimount) Formation.  
927 The third unit is Paleoproterozoic to Neoproterozoic in age (the ca. 1706-1640 Ma to > ca. 883 Ma  
928 Tizi 'n'Taghatine Group that incorporates the Ifarkhs-n-Tirsal, Wanimzi, Tamgarda, Agoumy,  
929 and Imi n-Tizi formations. The fourth unit is the ca. 883 Ma Tachdamt Formation and the  
930 uppermost unit is the ca. 700 Ma Bleida Formation.

931 Using this lithostratigraphic framework in combination with ages of LIPs on the WAC and herein  
932 developed correlation with the Taoudeni Basin succession highlights protracted and yet episodic  
933 mid-Proterozoic sedimentation in intracratonic basins on the WAC in association with extensional  
934 events and far-distance plate-boundary stresses that lasted from the early stage in the Nuna /  
935 Columbia assembly to the early stage in the breakup of the Rodinia supercontinent. Deposition of

936 the Tasserda-Taghatine Group is potentially linked to the post-Eburnean orogenic collapse. The  
937 Oumoula (Mimount) Formation might reflect the intracontinental extension within the Nuna  
938 /Columbia supercontinent in response to either ca. 1750 Ma or 1650 Ma LIP event. The Tizi  
939 'n'Taghatine Group might be correlative to the ca. 1.1 Ga Atar Group (Taoudeni Basin). The  
940 Tachdamt Formation records rifting within the Rodinia supercontinent, but it is uncertain whether  
941 this event led to the development of the passive continental margin along the northern boundary of  
942 the WAC. Finally, the Bleida Formation deposited in the foreland basin records the early stage in  
943 the assembly of the Gondwana supercontinent. The WAC formed the core of the Nuna / Columbia  
944 supercontinent and transited to the Rodinia supercontinent directly unaffected by the supercontinent  
945 breakup and subsequent assembly.

946 (2) The maximum depositional age and provenance of the lower Anti-Atlas Supergroup are  
947 constrained by new detrital zircon ages from the lower, middle, and upper parts of the sedimentary  
948 successions at the Zenaga and Bou Azzer El Graara inliers. The inferred maximum depositional  
949 ages are: ca. 2049 Ma for the Tasserda Formation, ca. 1818 Ma for the Agoumy Formation, ca.  
950 1060 Ma for the Tachdamt Formation, and ca. 700 Ma for the Bleida Formation. Four detrital  
951 zircon age clusters have been defined in this study: ca. 3.0-2.4, 2.3-1.74, 1.7-1.4, and 1.25-0.95 Ga.  
952 The provenance of detrital zircon was mainly on the WAC (that is the Reguibat and Leo-Man  
953 shields), but other cratons such as Amazonia, Baltica, and the Sahara Metacraton (Hoggar, Mali)  
954 cannot be excluded as possible subsidiary source areas. Importantly, we find that subsequent to the  
955 Grenville Orogeny, long-distance transport by fluvial systems delivered onto the WAC  
956 Mesoproterozoic zircons either from Fennoscandia of Baltica or Sunsas Belt of Amazonia (see Fig.  
957 15). Transport from either of these two areas requires giant fluvial systems directed inward of the  
958 Rodinia supercontinent.

959 (3) The Tachdamt Formation (lithostratigraphic unit 4) consists mainly of a succession of basaltic  
960 lava flows. A ca. 885 Ma mafic dyke swarm in the Zenaga and Iguerda-Taïfast inliers and the  
961 coeval volcanics of the Tachdamt Formation constitute a LIP in the Anti-Atlas Belt of Morocco,



962 named the Iguerda-Taïfast LIP. Similarly aged LIPs also occur on other cratons such as the Tarim  
963 Terrane (North China) and Congo Craton (Central Africa) and might reflect rifting at the late stage  
964 in the assembly of the Rodinia supercontinent.

965

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982

## 983 **Figure captions**

984 **Fig. 1.** (A) Location of Northwestern Africa; (B) Geological map of Northwestern Africa (after  
985 Thiéblemont, 2016 and Brahimi et al., 2018); (C) Distribution of main tectonic elements (after  
986 Brahimi et al., 2018); and (D) Geological map of the Anti-Atlas Belt in southern Morocco showing

987 the studied area in the Zenaga and Bou Azzer El Graara inliers (after Gasquet et al., 2008).

988 Abbreviations: G.- Groups; J.- Jbel; Paleoprot.- Paleoproterozoic; T.-Tizi

989

990 **Fig. 2.** Generalized lithostratigraphic column for the Anti-Atlas Pan-African orogen. After Thomas  
991 et al., (2004), modified by Gasquet et al., (2005; 2008) and Youbi et al., (2013), and taken from  
992 Youbi et al., (2013). The conventional viewpoint has been that there is a gap in magmatic activity  
993 between 1.7 and 1.0 Ga in the Anti-Atlas Belt (Mesoproterozoic gap). However, this interpretation  
994 was challenged with the results of Youbi et al. (2013), El Bahat et al. (2013), Kouyaté et al. (2013),  
995 Söderlund et al. (2013), which all indicated important intraplate magmatic events at ca. 1.75, 1.65,  
996 and 1.38-1.41 Ga. See the text for more explanation. Note that this figure is the outdated correlation  
997 scheme for the Anti-Atlas Belt since it uses outdated chronology of the Anti-Atlas cover  
998 successions. Abbreviations: PI (XI), PII (XII), PII-III (XII-III) and PIII (XIII) are the classical  
999 stratigraphic symbols used on Anti-Atlas geological maps (Choubert, 1963). Adoud.- Adoudounien  
1000 (or Adoudounian); Calc. Sup.- “Calcaires Supérieures” (Upper Limestones) ; Cambr.- Cambrian;  
1001 HKCA- High-K calc-alkaline (granitoids); L. d. Vin - Lie de Vin (wine-coloured) ; Paleoproteroz.-  
1002 Paleoproterozoic.

1003

1004 **Fig. 3.** Lithostratigraphic subdivision, sequence stratigraphy, and paleogeographic data for the  
1005 lower Anti-Atlas Supergroup (modified after Bouougri and Saquaque, 2004). We use the name of  
1006 the Ifrane-n-Taghatine Formation instead of the Taghdout Formation to avoid confusion with the  
1007 widely used name of the group - “the Taghdout Group” in the Zenaga Inlier. Abbreviations: systems  
1008 tracts consist of (LST) lowstand system tracts, (TST) transgressive system tracts, and (HST)  
1009 highstand system tracts. U1 to U5 are unconformities. DS1 to DS5 are depositional systems.

1010

1011 **Fig. 4.** Geology of the Zenaga Inlier (modified after Choubert, 1963; Ennih and Liégeois, 2001;  
1012 Thomas et al., 2002; Kouyaté et al., 2013; Youbi et al., 2013). The Taghdout Group corresponds to

1013 the lower Anti-Atlas Supergroup (Thomas et al., 2004). Inset shows a rose diagram that summarizes  
 1014 mafic dyke trends measured from the map and Google Earth; n is the number of measurements. The  
 1015 U-Pb ages obtained in this study are indicated with asterisks; other age constraints are from  
 1016 literature (Kouyaté et al., 2013; Youbi et al., 2013).

1017

1018 **Fig. 5.** Simplified geological map of the Bou Azer-El Graara Inlier showing the main lithotectonic  
 1019 units. The localities (Tachdamt and Bleida) mentioned in the text are indicated, as well as the  
 1020 location of samples used for U-Pb detrital geochronology (after Blein et al., 2014a). The Bleida  
 1021 Group corresponds to the lower Anti-Atlas Supergroup (Blein et al., 2014a)

1022

1023 **Fig. 6.** (A) General view of the unconformity between the Eburnean basement and the overlying  
 1024 quartzites of Oumoula Formation (O Fm.) of the lower Anti-Atlas Supergroup at Tizi n'Taghatine  
 1025 locality. The Eburnean basement (EB: gneiss and schists) is crosscut by pegmatite dyke (PE). The  
 1026 arrow indicates the road from Taliwine to Tazenakht. Geologists are Fida Medina, Maria Giuditta  
 1027 Fellin, Claudio Faccenna, and Riccardo Lanari; (B) View of the unconformity between the schists  
 1028 of the Bleida Formation (B Fm.) and the overlying volcanic rocks of the Ouarzazate Supergroup  
 1029 (OS); (C) View of the Taghdout sill (TS) crosscutting the mixed siliciclastic-carbonate succession of  
 1030 the Ifrane-n-Taghatine Formation (IT Fm.) in the Zenaga Inlier near the Taghdout Town. Note  
 1031 hornfels (h) at the baked margin of the sill. The Ifrane-n-Taghatine Formation is overlain by thick  
 1032 quartzites of Oumoula Formation (O Fm.); and (D) Detailed view of the upper contact between the  
 1033 Taghdout dolerite sill (TS) and limestones of the Ifrane-n-Taghatine Formation (IT Fm.)  
 1034 transformed into hornfels.

1035

1036 **Fig. 7.** Thin-section photomicrographs of the Great Taghdout sill. (A) coarse-grained dolerite in the  
 1037 centre of the sill with plagioclase (Pl) and clinopyroxene (Cpx) minerals and oxides (Ox), crossed  
 1038 nicols; (B) fine-grained, microlitic porphyric texture at the border of the sill with microlites (tiny

1039 crystals) of plagioclase (Pl) and clinopyroxene (Cpx) minerals and oxides (Ox). Epidote (Ep) is  
1040 filling vesicles; crossed nicols.

1041

1042 **Fig. 8.** Th/U ratios versus concordant U-Pb ages of igneous and detrital zircon grains of the lower  
1043 Anti-Atlas Supergroup.

1044

1045 **Fig. 9.** Zircon U-Pb SHRIMP geochronology: (A) Wetherill concordia plots of U-Pb zircon data for  
1046 the Taghdout sill (sample TA) of the Zenaga Inlier. (B) Weighted mean average  $^{207}\text{Pb}/^{206}\text{Pb}$  age of  
1047 analysis with low Th and U concentrations (spots 2.1, 7.1, and 13.1).

1048

1049 **Fig. 10.** (A) Total alkali vs. silica (TAS) diagram for the chemical classification and nomenclature  
1050 (after Le Bas et al., 1986). Subalkaline vs. alkaline series boundary (dotted line) is from Irvine and  
1051 Baragar (1971). (B) Ti vs. V variations in samples from the Zenaga, Agadir Melloul, Iguerda-  
1052 Taifast, and Tagragra of Akka inliers including four datasets reported by Mortaji (1989,  
1053 unpublished thesis), Hafid (1999, unpublished thesis), Hafid et al., (2001) and El Aouli et al.,  
1054 (2010); diagram is after Shervais (1982). The studied samples TA1 (centre of the sill dated at  $1676$   
1055  $\pm 37$  Ma; this work) and TA2 (border of the sill) and the 1.6 Ga dolerite dykes and sills from the  
1056 Zenaga and Agadir Melloul inliers collected by Kouyaté et al., (2013) are also shown on this  
1057 diagram.

1058

1059 **Fig. 11.** Primitive mantle-normalized trace element 'spidergram' for the dated mafic sill and dykes  
1060 of the Zenaga and Agadir Melloul inliers. Primitive mantle values used for normalization are from  
1061 Sun and McDonough (1989).

1062

1063 **Fig; 12.** Thin section photomicrographs of the sedimentary samples. (A) Metasiltstones of the  
1064 Tasserda Formation: sample TG-15-1; purple, coarse-grained and quartz-rich metasiltstone with

1065 quartz (Qz) and feldspar (Pl) embedded in a sericite-rich matrix (Mx); crossed nicols. (B) Quartzites  
1066 of the Agoumy Formation: sample BA21, with quartz grains that are mostly subrounded. (C)  
1067 Quartzites of the Tachdamt Formation: sample ALBL29, with a mineral assemblage comprised  
1068 mainly of quartz (Qz) associated to a small amount of feldspars (Pl) with a matrix (Mx) of silt-sized  
1069 quartz and mica minerals. (D) Metasandstone of the Bleida Formation: sample BA24, fine-grained,  
1070 moderately sorted metasandstone with quartz (Qz), plagioclase (Pl), and lithic fragments in mica  
1071 matrix (Mx).

1072

1073 **Fig. 13.** U-Pb ages for detrital zircon grains from samples TG-15-1 (metasiltstones of the Tasserda  
1074 Formation) and BA21 (quartzites of the Agoumy Formation) of the lower Anti-Atlas Supergroup.  
1075 (A) and (C) are concordia diagrams; (B) and (D) are combined binned frequency and probability  
1076 density plots of detrital zircon grains. Abbreviations: Conc.-Concordance; Decay-const. errs.-  
1077 Decay-constant errors

1078

1079 **Fig. 14.** U-Pb ages for detrital zircon grains from samples ALBL29 (quartzites of the Tachdamt  
1080 Formation) and BA24 (metasandstone of the Bleida Formation) of the lower Anti-Atlas  
1081 Supergroup. (A) and (C) are concordia diagrams; (B) and (D) are combined binned frequency and  
1082 probability density plots of detrital zircon grains. Abbreviations: Conc.-Concordance; Decay-const.  
1083 errs.- Decay-constant errors

1084

1085 **Fig. 15.** Paleomagnetically and geologically informed reconstruction of WAC, Amazonia, and  
1086 Baltica cratons (SAMBA connection) along with Laurentia and Siberia in the core of the Nuna  
1087 supercontinent (modified after Evans and Mitchell, 2011); additional ca. 1640-1680 Ma dyke  
1088 swarms include the  $1641 \pm 8$  Ma Nersa sill complex of Southern Siberia (U-Pb baddeleyite TIMS  
1089 age; Metelkin et al, 2011; Ernst et al., 2016),  $1635.0 \pm 2.7$  to  $1622.1 \pm 3.2$  Ma Melville Bugt dykes  
1090 of western Greenland (U-Pb baddeleyite TIMS age; Halls et al., 2011; Klausen and Nilsson, 2019),

1091 1663 ± 8 Ma bimodal Narakay Volcanic Complex of Canada (U-Pb zircon TIMS age; Bowring and  
1092 Ross, 1985), and 1642 ± 2 Ma dyke swarms of Häme, Sipoo, and Suomenniemi of the  
1093 Fennoscandian Shield in southeast Finland (U-Pb ID-TIMS baddelleyite age; Vaasjoki et al., 1991;  
1094 Salminen et al., 2017). Star locates mantle plume centre, proposed for the Melville Bugt swarm by  
1095 Klausen and Nilsson (2019).

1096

1097 **Fig. 16.** Age compilation for detrital zircon grains from four samples collected from the lower Anti-  
1098 Atlas Supergroup in the Zenaga and Bou Azzer El Graara inliers. Ages older than 1.7 Ga indicate  
1099 the West African provenance (modified from Linnemann et al., 2010 and Abubaker et al., 2017).  
1100 See the text for more explanation. Abbreviations: Conc.-Concordance

1101

1102 **Fig. 17.** Distribution of U-Pb ages for detrital zircon grains analyzed in this study from the lower  
1103 Anti-Atlas Supergroup in the Zenaga and Bou Azzer El Graara inliers (all grouped together) and  
1104 potential sources compiled from the literature. Gray stripes: detrital zircons; black stripes: zircons  
1105 from magmatic rocks. Detrital zircon data sources: Pyrenees (Margalef et al., 2016); Lybia (Altumi  
1106 et al., 2013); Algeria, Ahaggar (Linnemann et al., 2011); Meguma, Ganderia, and Avalonia detrital  
1107 zircons (Fyffe et al., 2009; Gärtner et al., 2016; Henderson et al., 2016; Murphy et al., 2004a,  
1108 2004b; Waldron et al., 2009 and references therein); Anti-Atlas and Sardinia Cambrian (Avigad et  
1109 al., 2012); Anti-Atlas Neoproterozoic (Abati et al., 2010; Karaoui et al., 2015). Anti-Atlas mafic  
1110 rocks (Youbi et al., 2013 and references therein). Modified after Linnemann et al., (2011), Pratt et  
1111 al., (2015), and Marzoli et al., (2017), and references therein. Abbreviations: ANS = Arabian-  
1112 Nubian Shield; CAMP-Central Atlantic Magmatic Province; Camb.-Cambrian; Dev.- Devonian;  
1113 Jur., Jurassic; Neoprot.- Neoproterozoic; Ord.- Ordovician; Sil.-Silurian

1114

1115 **Fig. 18.** A tentative correlation scheme of the lower Anti-Atlas Supergroup and its stratigraphic  
1116 equivalents in the Anti-Atlas inliers. U1 to U8 are unconformities. U-Pb geochronology data are

1117 from (1) this study, (2) Letsch (2018), (3) Youbi et al., (2018) and Bouougri et al., (2018; 2020), (4)  
1118 Abati et al., (2010), (5) Walsh et al., (2012), (6) Thomas et al., (2002), (7) Soulaïmani et al.,  
1119 (2013a), Hafid et al., (2013), Blein et al., (2014b), (8) Soulaïmani et al., (2019), (9) Walsh et al.,  
1120 (2002), (10) Ikenne et al., (2017), (11) Aït Malek et al., (1998), (12) Youbi et al., (2019), (13)  
1121 Benziane et al., (2016), and (14) Schulte et al., (2016).

1122

### 1123 **Table captions**

1124

1125 **Table 1.** Site location and summary of the data for the studied samples. The names of units and  
1126 formations are sensu Bouougri and Saquaque (2004).

1127

1128 **Table 2.** Representative whole-rock major and trace element geochemical data for dykes and sills  
1129 from the Zenaga, Agadir Melloul, Iguerda-Taïfast, and Tagragra of Akka inliers including the dated  
1130 dykes, samples TA1 from the Zenaga Inlier ( $1676 \pm 37$  Ma; this work), DZ 36 ( $1640 \pm 2$  Ma and  
1131  $1641 \pm 3$  Ma; Kouyaté et al., 2013) and DZ 40 ( $1656 \pm 9$  Ma; Kouyaté et al., 2013) from the Zenaga  
1132 Inlier (Kouyaté et al, 2013), and AMHA069 ( $1654 \pm 16$  Ma; Kouyaté et al, 2013) from the Agadir  
1133 Melloul Inlier. The geochemical data are from unpublished theses (Hafid, 1999; Mortaji, 1989) and  
1134 published papers (Hafid et al., 2001; El Aouli et al., 2010; Youbi et al., 2013).  $\text{Fe}_2\text{O}_3\text{t}$  = total Fe  
1135 expressed as  $\text{Fe}_2\text{O}_3$ . Loss-on-ignition (L.O.I.) determined by sample weight loss at  $1000^\circ\text{C}$ .

1136

### 1137 **Appendix: Supplementary data**

1138

1139 **Figure S1** of supplementary data. Cathodoluminescence (CL) images of zircon grains obtained  
1140 from the sample TA 1 showing their internal texture, spots (yellow circles), and apparent ages in  
1141 Ma ( $1\sigma$  error).

1142

1143 **Figure S2** of supplementary data. Cathodoluminescence (CL) images of zircon grains obtained  
1144 from the sample TG-15-1 (metasiltstone of the Tasserda Formation) of the lower Anti-Atlas  
1145 Supergroup showing their internal texture, laser spots (yellow circles=concordant ages; red  
1146 circles=discarded ages due to high discordance and/or U/Th/Pb contents), and apparent ages in Ma  
1147 ( $1\sigma$  error).

1148

1149 **Figure S3** of supplementary data. Cathodoluminescence (CL) images of zircon grains obtained  
1150 from the sample BA21 (quartzites of the Agoumy Formation) of the lower Anti-Atlas Supergroup  
1151 showing their internal texture, laser spots (yellow circles=concordant ages; red circles= discarded  
1152 ages due to high discordance and/or U/Th/Pb contents), and apparent ages in Ma ( $1\sigma$  error).

1153

1154 **Figure S4** of supplementary data. Cathodoluminescence (CL) images of zircon grains obtained  
1155 from the sample ALBL29 (quartzites of the Tachdamt Formation) of the lower Anti-Atlas  
1156 Supergroup showing their internal texture, laser spots (yellow circles=concordant ages; red circles=  
1157 discarded ages due to high discordance and/or U/Th/Pb contents), and apparent ages in Ma ( $1\sigma$   
1158 error).

1159

1160 **Figure S5** of supplementary data. Cathodoluminescence (CL) images of zircon grains obtained  
1161 from the sample BA24 (metasandstone of the Bleida Formation) of the lower Anti-Atlas  
1162 Supergroup showing their internal texture, laser spots (yellow circles=concordant ages; red circles=  
1163 discarded ages due to high discordance and/or U/Th/Pb contents), and apparent ages in Ma ( $1\sigma$   
1164 error).

1165

1166 **Table S1** of supplementary data. SHRIMP zircon U-Th-Pb data for the Taghdout Sill (Sample  
1167 TA1).

1168



1169 **Table S2** of supplementary data. LA-ICP-MS zircon U-Th-Pb data for the sample TG-15-1  
1170 (metasiltstone of the Tasserda Formation) of the lower Anti-Atlas Supergroup.

1171

1172 **Table S3** of supplementary data. SHRIMP zircon U-Th-Pb data for the sample BA21 (quartzites of  
1173 the Agoumy Formation) of the lower Anti-Atlas Supergroup.

1174

1175 **Table S4** of supplementary data. LA-ICP-MS zircon U-Th-Pb data for the sample ALBL29  
1176 (quartzites of the Tachdamt Formation) of the lower Anti-Atlas Supergroup.

1177

1178 **Table S5** of Supplementary data. SHRIMP zircon U-Th-Pb data for the sample BA24  
1179 (metasandstone of the Bleida Formation) of the lower Anti-Atlas Supergroup.

1180

1181

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1182

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Table 1. Site location and summary of the data for the studied samples. The names of units and formations are sensu Bouougri and Saquaque (2004).

Sample ID	Coordinates (GPS WGS84)	Inlier/ Subinlier	Lithology	Units/Formations/Suites	Methodes/ Mineral	Number of spots	U-Pb zircon Age (Ma)
BA24	N 30°23'7.90", W 6°30'33.40"	Bou Azzer El Graara/Bleida	Fine-grained sandstone	Upper Unit/Bleida Formation	SHRIMP/II/Zircon	44 spots	1855 ± 6.5 - 977 ± 10*
ALBL29	N 30°20'19.28", W 6°24'48.70"	Bou Azzer El Graara/Bleida	Medium to coarse-grained quartzite	Middle Unit/Tachdamt Formation	LA-ICP-MS/Zircon	96 spots	1060 ± 5.5 - 930 ± 12*
BA21	N 30°29'11.40", W 6°52'44.00"	Bou Azzer El Graara/Tachdamt	Medium-grained quartzite	Lower Unit/Agoumy Formation	SHRIMP/II/Zircon	55 spots	1818 ± 6.4*
TA1	N 30°36'43.6", W 7°17'13.4"	Zenaga	Coarse-grained to a fine-grained microlitic porphyric dolerite	Ifzwane Suite	SHRIMP/II/Zircon	18 spots	1676 ± 37**
TG-15-1	N 30°35'11.76", W 7°20'06.92"	Zenaga	Purple coarse-grained and quartz rich metasilstone	Lower Unit/Tasserda Formation	LA-ICP-MS/Zircon	99 spots	2049.3 ± 4.6*

\* U-Pb detrital zircon, maximum depositional age

\*\* U-Pb age emplacement

SHRIMP/II: Sensitive High Resolution Ion Microprobe

LA-ICP-MS: Laser Ablation Inductively Coupled Plasma Mass Spectrometry

ID: Identification

GPS WGS84: The Global Positioning System (GPS) uses the World Geodetic System (WGS84) as its reference coordinate system.



Table 1. Site location and data summary of the investigated samples. The names of Units and Formations are sensu Bouougri and Saquaque (2004).

Sample ID	Coordinates (GPS WGS84)	Inlier/ Subinlier	Lithology	Units/Formations/Suites	Methodes/ Mineral	Number of spots	U-Pb zircon Age (Ma)
BA 24	N 30°23'7.90", W 6°30'33.40"	Bou Azzer El Graara/Bleida	Fine-grained sandstone	Upper Unit/Bleida Formation	SHRIMP/II/ Zircon	44 spots	1957 ± 63*
ALBL29	N 30°20'19.28", W 6°24'48.70"	Bou Azzer El Graara/Bleida	Medium to coarse-grained quartzarenite	Middle Unit/Tachdamt Formation	LA-ICP-MS/Zircon	96 spots	1060 ± 5.5*
BA 21	N 30°29'11.40", W 6°52'44.00"	Bou Azzer El Graara/Tachdamt	Medium-grained quartzarenite	Lower Unit/Agoumy Formation	SHRIMP/II/ Zircon	54 spots	1818 ± 6.4*
TA 1	N 30°36'43.6", W 7°17'13.4"	Zenaga	Coarse-grained to a fine-grained microlitic porphyric dolerite	Ifzwane Suite	SHRIMP/II/ Zircon	18 spots	1639 ± 34 **
TG-15-1	N 30°35'11.76", W 7°20'06.92"	Zenaga	Purple coarse-grained and quartz rich siltstone	Lower Unit/Tassarda Formation	LA-ICP-MS/Zircon	99 spots	2047 ± 9.7*

\* U-Pb detrital zircon, maximum depositional age

\*\* U-Pb age emplacement

SHRIMP/II: Sensitive High Resolution Ion Microprobe

LA-ICP-MS: Laser Ablation Inductively Coupled Plasma Mass Spectrometry

ID: Identification

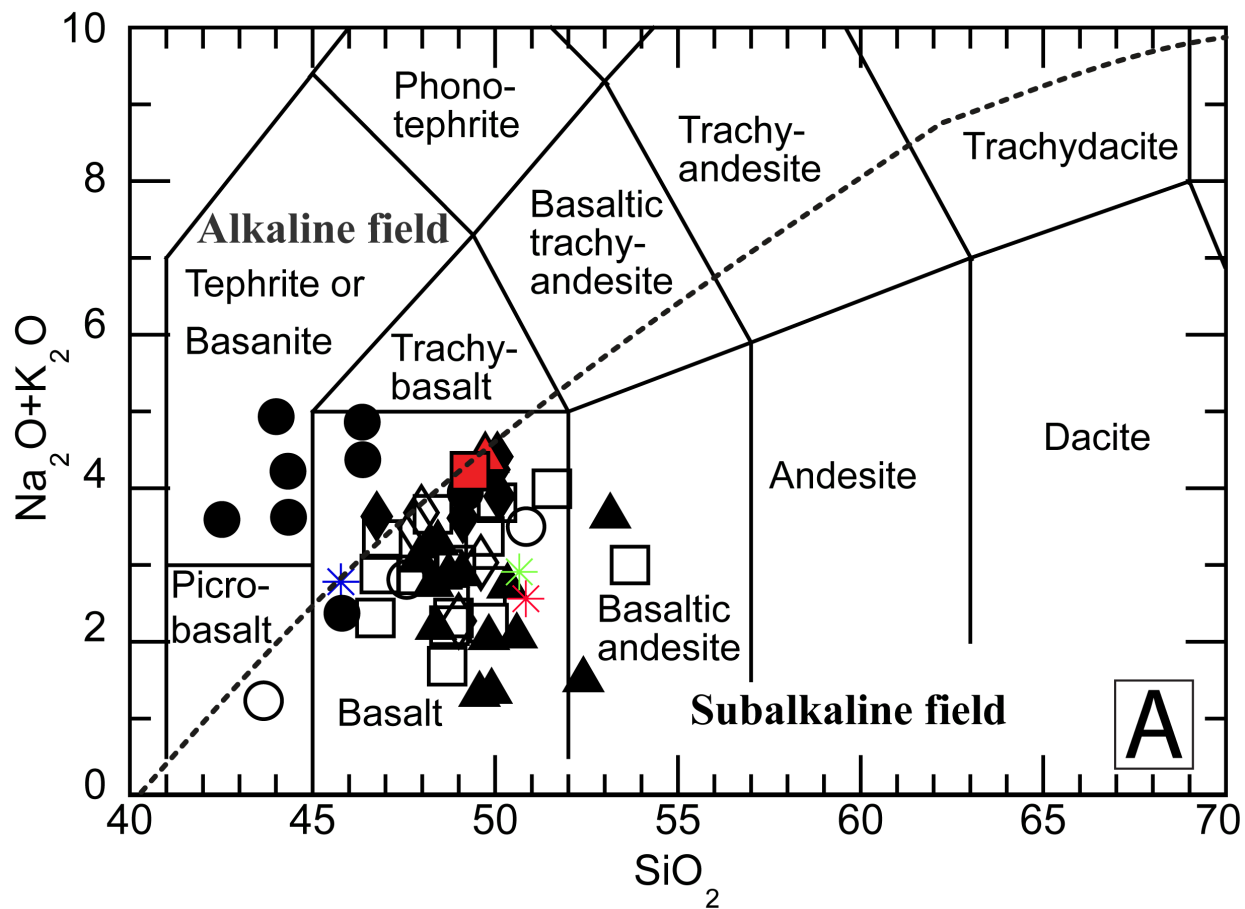
GPS WGS84: The Global Positioning System (GPS) uses the World Geodetic System (WGS84) as its reference coordinate system.

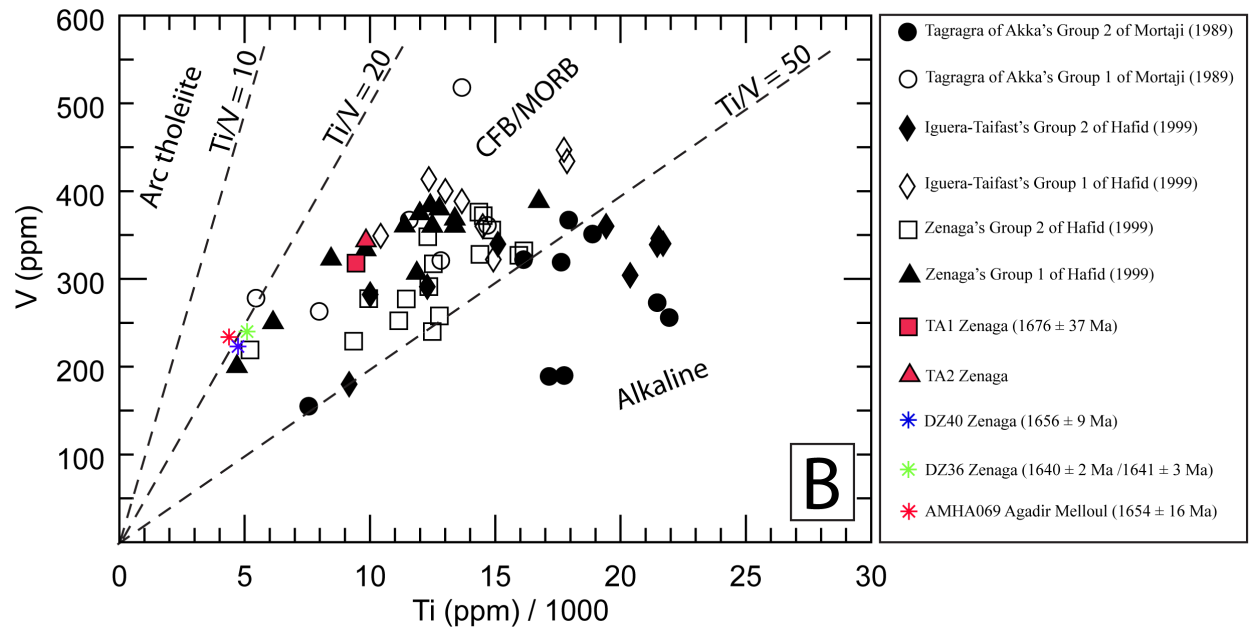
Table 2. Representative whole-rock major and trace element geochemical data for dykes and sills from the Zenaga, Agadir Melloul, Iguerda-Taïfast, and Tagragra of Akka inliers including the dated dykes, samples TA1 from the Zenaga Inlier ( $1676 \pm 37$  Ma; this work), DZ 36 ( $1640 \pm 2$  Ma and  $1641 \pm 3$  Ma; Kouyaté et al., 2013) and DZ 40 ( $1656 \pm 9$  Ma; Kouyaté et al., 2013) from the Zenaga Inlier (Kouyaté et al., 2013), and AMHA069 ( $1654 \pm 16$  Ma; Kouyaté et al., 2013) from the Agadir Melloul Inlier. The geochemical data are from unpublished theses (Hafid, 1999; Mortaji, 1989) and published papers (Hafid et al., 2001; El Aouli et al., 2010; Youbi et al., 2013). Fe<sub>2</sub>O<sub>3</sub>t= total Fe expressed as Fe<sub>2</sub>O<sub>3</sub>. Loss-on-ignition (L.O.I.) determined by sample weight loss at 1000°C.

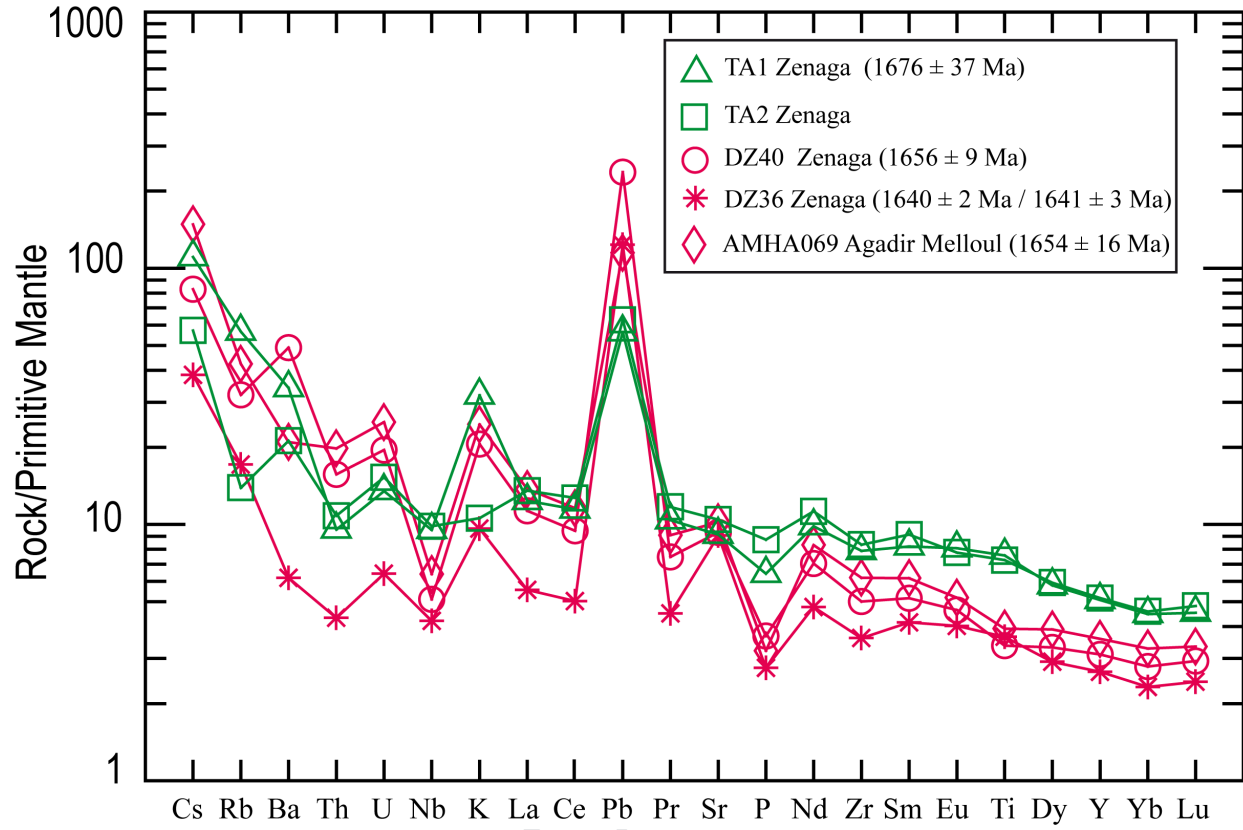
Sample	TA1	TA2	DZ36	DZ40	AMHA69	ZEN-4	ZEN-13	ZEN-1	ZEN-11	IT-5	IT-7	IT-12	IT-13	NC40	NC379	NC314aMor	NC 361
Inlier	Zenaga	Zenaga	Zenaga	Zenaga	Agadir Melloul	Zenaga	Zenaga	Zenaga	Zenaga	Iguerda-Taïfast	Iguerda-Taïfast	Iguerda-Taïfast	Iguerda-Taïfast	Tagragra of Akka	Tagragra of Akka	Tagragra of Akka	Tagragra of Akka
Latitude N	30°36'43.6"	30°36'43.7"															
Longitude W	7°17'13.4"	7°17'13.3"															
Orientation	N45°	N45°	N125°	N140°		N170°	N125°	N85°	N170°	N00°	N30°	N45°	N50°-N110°	N50°-N110°	N30°-N160°	N30°-N160°	
Ages (Ma)	1676 ± 37		1640 ± 2/ 1641 ± 3	1656 ± 9	1654 ± 16												
Reference	(this work)	(this work)	Kouyaté (2013)	Kouyaté (2013)	Kouyaté (2013)	Hafid (1999)	Hafid (1999)	Hafid (1999)	Hafid (1999)	Hafid (1999)	Hafid (1999)	Hafid (1999)	Hafid (1999)	Mortaji (1989)	Mortaji (1989)	Mortaji (1989)	Mortaji (1989)
SiO <sub>2</sub>	49.31	49.73	50.66	50.84	45.78	48.56	49.71	48.38	48.30	49.12	49.98	49.00	47.98	50.85	47.58	44.35	49.18
TiO <sub>2</sub>	1.64	1.58	0.85	0.73	0.79	2.66	2.09	1.98	1.64	2.05	2.52	2.17	2.06	0.91	1.93	2.86	2.99
Al <sub>2</sub> O <sub>3</sub>	13.79	13.48	14.55	14.61	13.73	13.58	14.08	13.97	13.97	13.50	13.67	13.06	13.20	14.72	13.62	15.25	12.75
Fe <sub>2</sub> O <sub>3</sub> t	14.51	14.35	10.27	9.78	9.50	14.43	14.83	13.71	13.16	15.17	15.02	14.87	15.54	12.60	15.63	16.53	16.06
MnO	0.23	0.23	0.16	0.18	0.25	0.17	0.19	0.16	0.20	0.26	0.17	0.23	0.26	0.19	0.20	0.18	0.12
MgO	6.34	6.68	7.31	7.55	9.87	6.61	5.76	6.55	7.73	5.28	5.24	5.91	6.24	6.69	6.29	5.95	5.21
CaO	6.70	6.20	10.01	11.27	12.18	8.71	8.51	10.18	10.34	8.52	6.98	10.04	8.40	7.22	8.82	7.16	4.36
Na <sub>2</sub> O	3.26	4.06	2.17	1.94	2.49	2.40	2.59	1.86	2.10	2.46	2.45	1.91	2.73	3.36	2.04	2.58	2.63
K <sub>2</sub> O	0.96	0.32	0.74	0.62	0.29	0.54	0.78	0.32	0.64	1.15	1.79	0.36	0.95	0.14	0.77	1.04	1.31
P <sub>2</sub> O <sub>5</sub>	0.14	0.19	0.07	0.08	0.06	0.24	0.33	0.18	0.15	0.43	0.36	0.19	0.17	0.14	0.24	0.69	0.50
L.O.I.	2.47	3.04	2.48	2.38	4.37	2.35	2.18	2.67	2.02	1.46	1.73	2.07	1.99	3.67	2.69	3.80	5.00
Total	99.34	99.84	99.49	100.25	99.68	100.25	101.05	99.96	100.25	99.40	99.91	99.81	99.52	100.49	99.81	100.39	100.11
V	343.24	317.91	240.00	233.70	223.00	326.90	317.14	306.30	332.92	290.97	339.52	400.00	413.57	278.00	367.00	189.00	367.00
Cr	18.95	18.83	307.70	410.10	802.70	189.80	96.76	197.10	181.59	114.14	126.86	113.12	180.31	139.00	114.00	180.00	118.00
Co	51.53	54.75	46.00	44.67	57.36	43.13	50.47	40.43	55.85	44.43	46.07	44.10	43.62	57.00	54.00	46.00	46.00
Ni	39.42	37.22	143.00	185.20	178.30	117.00	69.71	94.16	121.74	44.20	64.28	79.64	92.72	55.00	95.00	109.00	65.00
Zn	126.00	158.84	90.36	92.06	228.20	150.20	152.60	119.80	161.27	131.25	129.48	134.44	155.27		116.00	153.00	62.00
Rb	35.95	8.84	26.89	20.34	10.88	16.85	20.90	8.44	19.70	63.40	68.05	24.02	45.85	10.00	43.00	29.00	81.00
Sr	192.75	220.51	215.20	196.60	191.10	391.10	410.50	269.90	256.15	372.90	370.30	253.56	200.55	593.00	185.00	247.00	120.00
Y	23.11	23.50	16.27	14.16	12.11	28.80	30.91	27.12	22.87	37.00	28.97	29.81	28.79	19.46	30.85	41.74	26.22
Zr	88.44	93.50	69.43	56.11	40.35	151.90	123.40	87.81	75.23	142.31	178.72	111.29	108.45		122.00	293.00	167.00
Nb	6.80	7.02	4.56	3.64	2.99	13.10	7.60	4.62	3.77	8.57	15.77	7.74	8.29		10.00	36.00	20.00
Ba	238.53	148.42	146.60	342.30	43.29	169.70	482.90	70.51	113.54	762.92	700.14	224.48	425.88	103.00	274.00	528.00	418.00

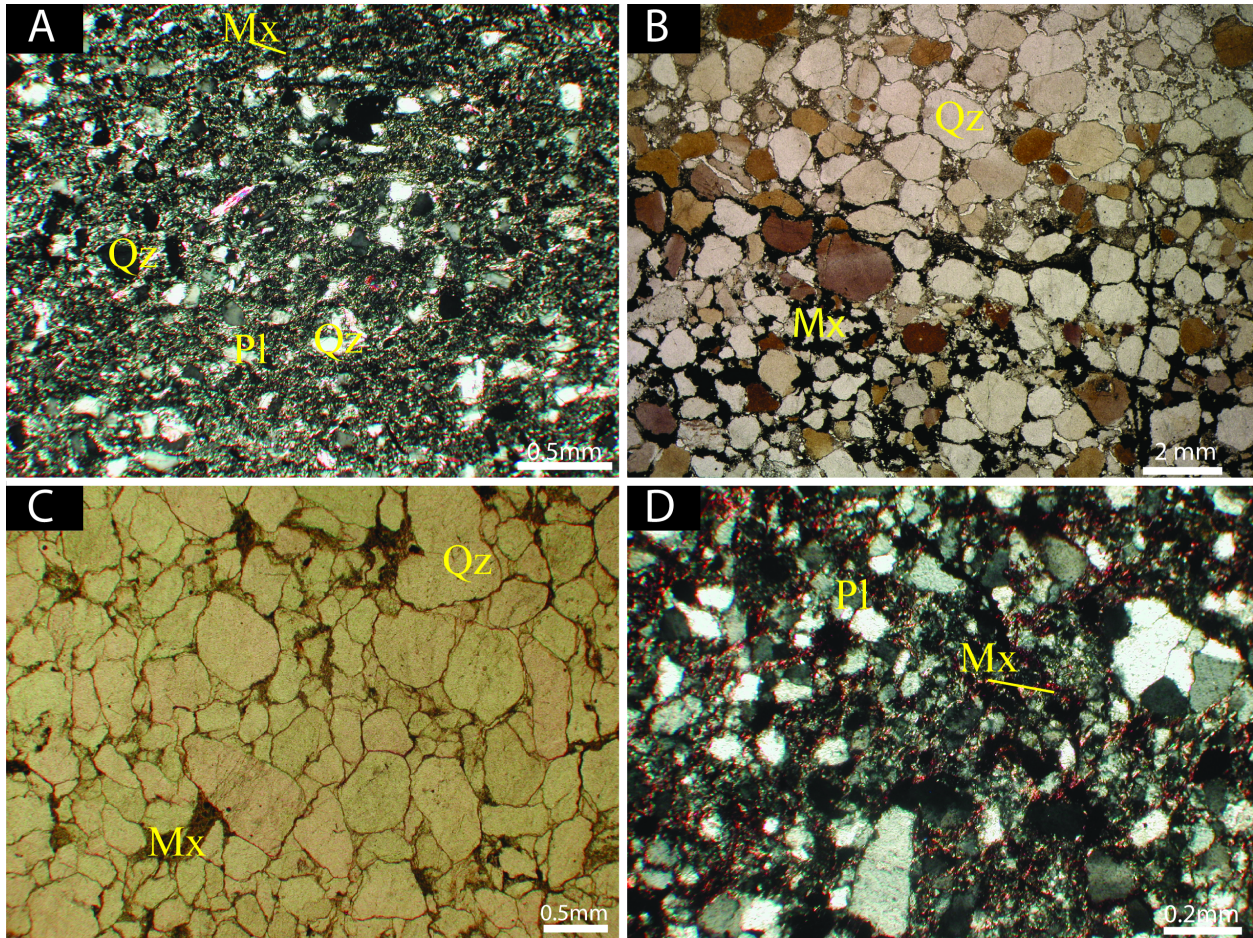
La	8.50	9.31	9.45	7.77	3.82	18.75	25.71	8.90	8.06	26.09	30.90	7.84	8.38	10.47	9.79	31.79	18.87
Ce	20.35	22.54	20.54	16.79	8.92	50.90	63.10	23.08	19.89	59.41	70.09	21.36	21.78	27.83	28.15	70.15	45.07
Pr	2.87	3.24	2.51	2.06	1.24	5.75	6.45	3.18	2.75	7.76	8.92	3.23	3.28				
Nd	13.34	15.24	11.26	9.54	6.45	25.95	26.89	14.22	12.43	33.49	37.03	16.27	15.26	13.43	17.19	40.17	27.38
Sm	3.65	4.05	2.74	2.29	1.84	6.44	6.11	4.55	3.81	7.32	7.46	4.56	4.44	3.60	5.26	8.75	7.06
Eu	1.36	1.31	0.87	0.78	0.68	2.21	2.07	1.56	1.31	1.96	2.42	1.68	1.53	1.01	1.52	1.87	2.32
Gd	3.98	4.24	2.81	2.42	2.11	6.61	6.10	4.47	3.55	6.89	7.19	5.96	5.53	3.40	5.49	7.93	6.94
Tb	0.66	0.69	0.47	0.39	0.35	1.07	0.99	0.84	0.67	1.06	1.05	0.96	0.93				
Dy	4.27	4.38	2.87	2.44	2.14	5.68	5.75	5.02	4.05	6.69	5.91	5.58	5.70	3.35	5.65	7.94	5.55
Ho	0.89	0.93	0.57	0.49	0.43	1.11	1.16	1.03	0.83	1.41	1.18	1.13	1.19				
Er	2.38	2.41	1.61	1.39	1.19	2.68	3.00	2.66	2.24	3.66	2.99	3.04	3.08	1.99	2.97	4.35	2.55
Tm	0.34	0.34	0.24	0.21	0.18	0.37	0.44	0.39	0.32	0.56	0.42	0.45	0.46				
Yb	2.21	2.26	1.62	1.38	1.14	2.39	2.70	2.45	1.88	3.34	2.53	2.68	2.77	2.01	2.82	4.12	2.04
Lu	0.34	0.36	0.25	0.22	0.18	0.35	0.39	0.36	0.29	0.51	0.37	0.40	0.42	0.37	0.46	0.67	0.36
Cu	279.90	20.75	67.27	60.43	193.30	22.64	61.31	68.13	10.68	73.30	20.24	200.74	18.26	15.00	148.00	33.00	18.00
U	0.28	0.32	0.53	0.41	0.14	0.54	0.47	0.24	0.25	0.40	1.65	0.18	0.21				
Th	0.81	0.92	1.69	1.33	0.37	1.77	2.42	0.89	0.79	1.70	3.65	0.64	0.80				
Ta	0.51	0.54	0.36	0.28	0.23	0.82	0.24	0.19	0.20	0.48	1.00	0.59	0.59				
Hf	2.45	2.65	2.00	1.61	1.16	3.86	3.03	2.29	1.95	3.99	4.92	3.36	3.29				
Sc	48.93	48.88				32.09	41.73	39.63	48.49	38.97	28.38	38.58	41.10		45.00	35.00	30.00
Ga	20.82	19.25	16.88	15.52	13.62	21.02	21.94	18.46	19.54	21.78	22.31	19.06	18.76				
Pb	4.00	4.47	8.16	16.85	8.78	7.07	9.65	5.42	5.55	8.50	6.98	5.52	5.90				
Cs	0.87	0.45	1.17	0.65	0.30	1.05	1.15	0.73	1.21	7.09	1.39	1.41	2.00				



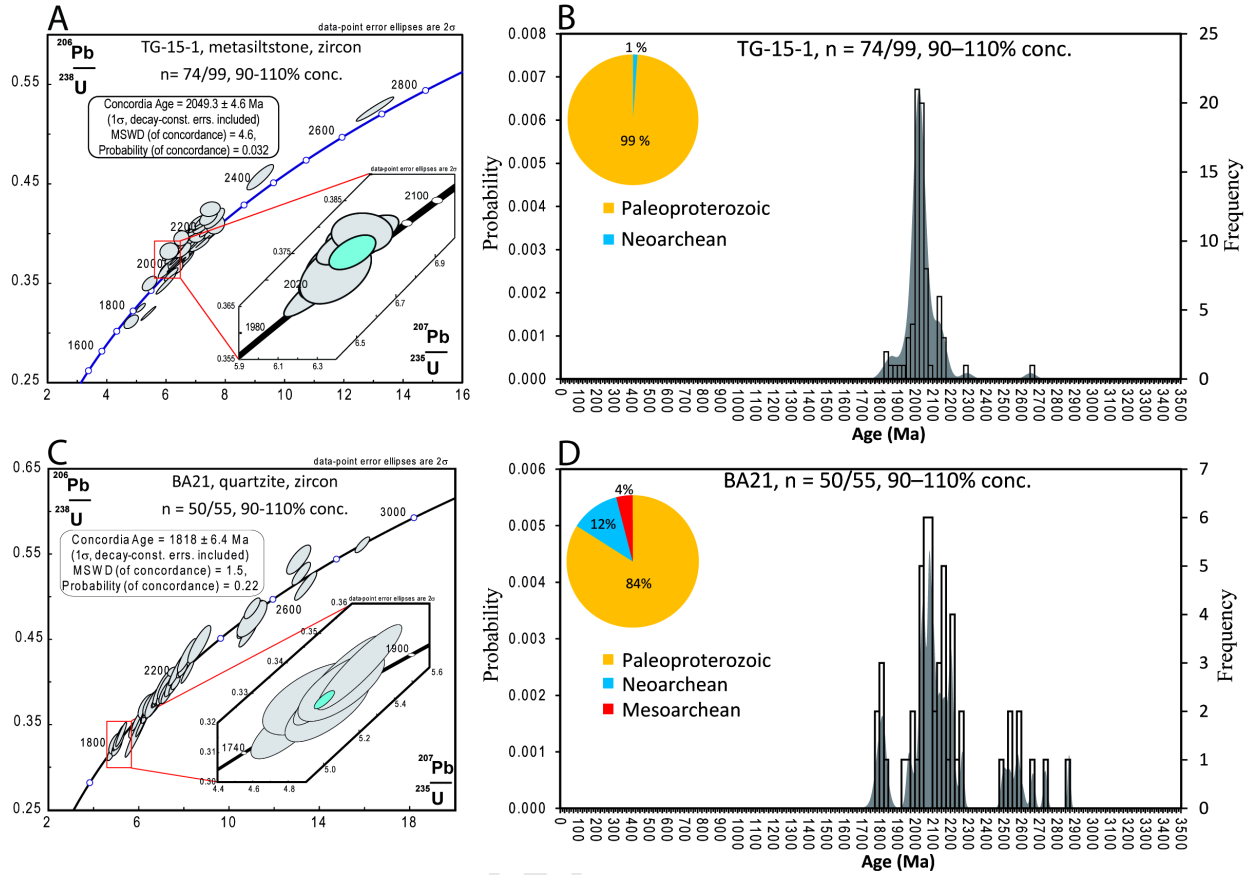


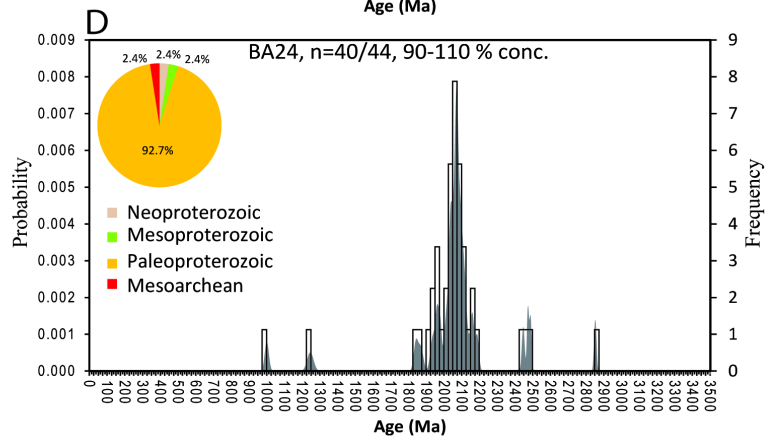
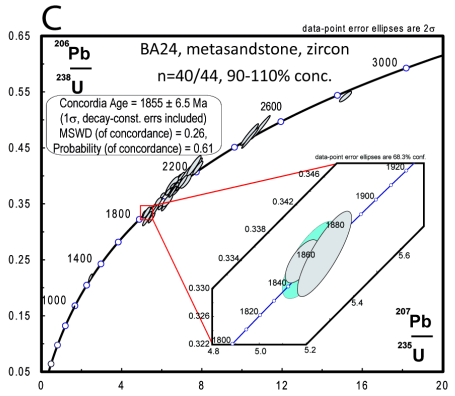
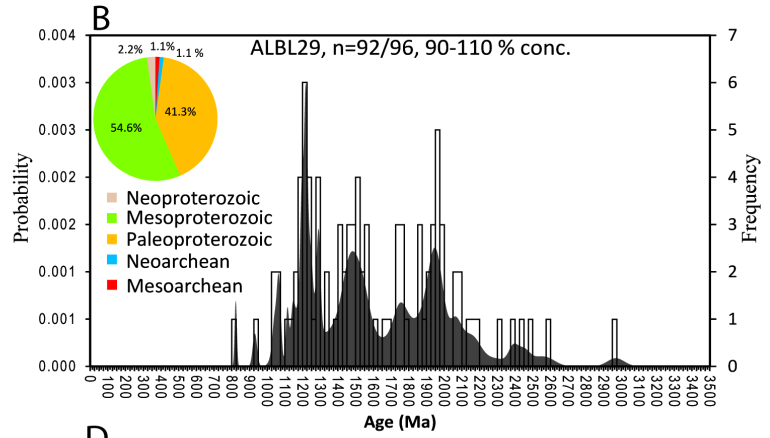
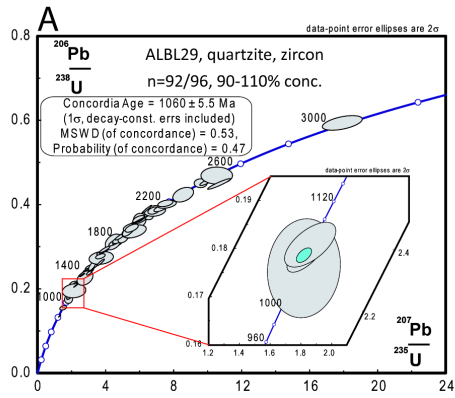


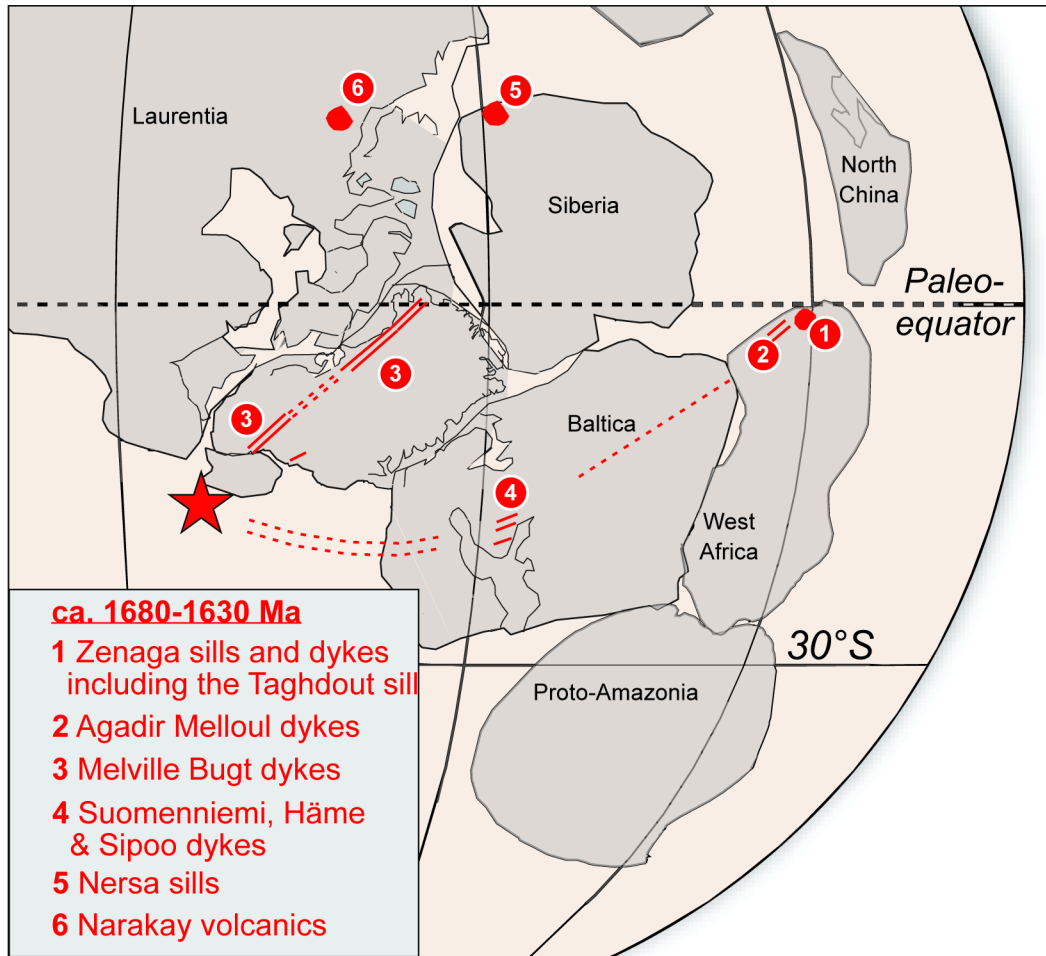


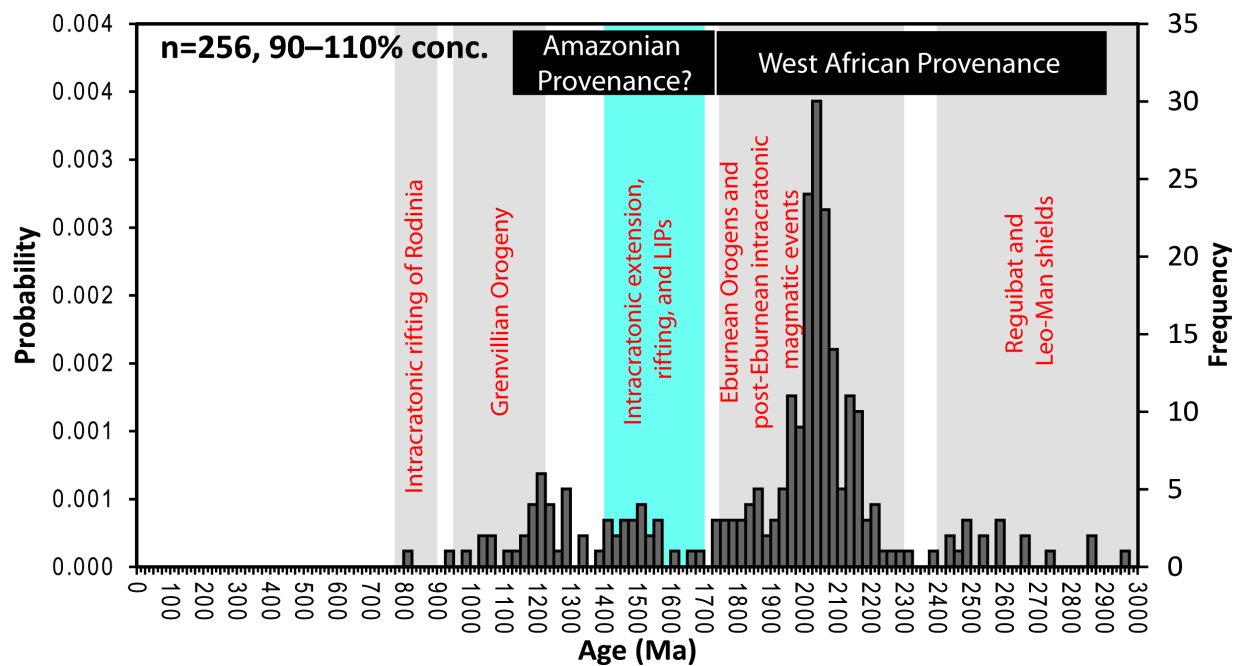


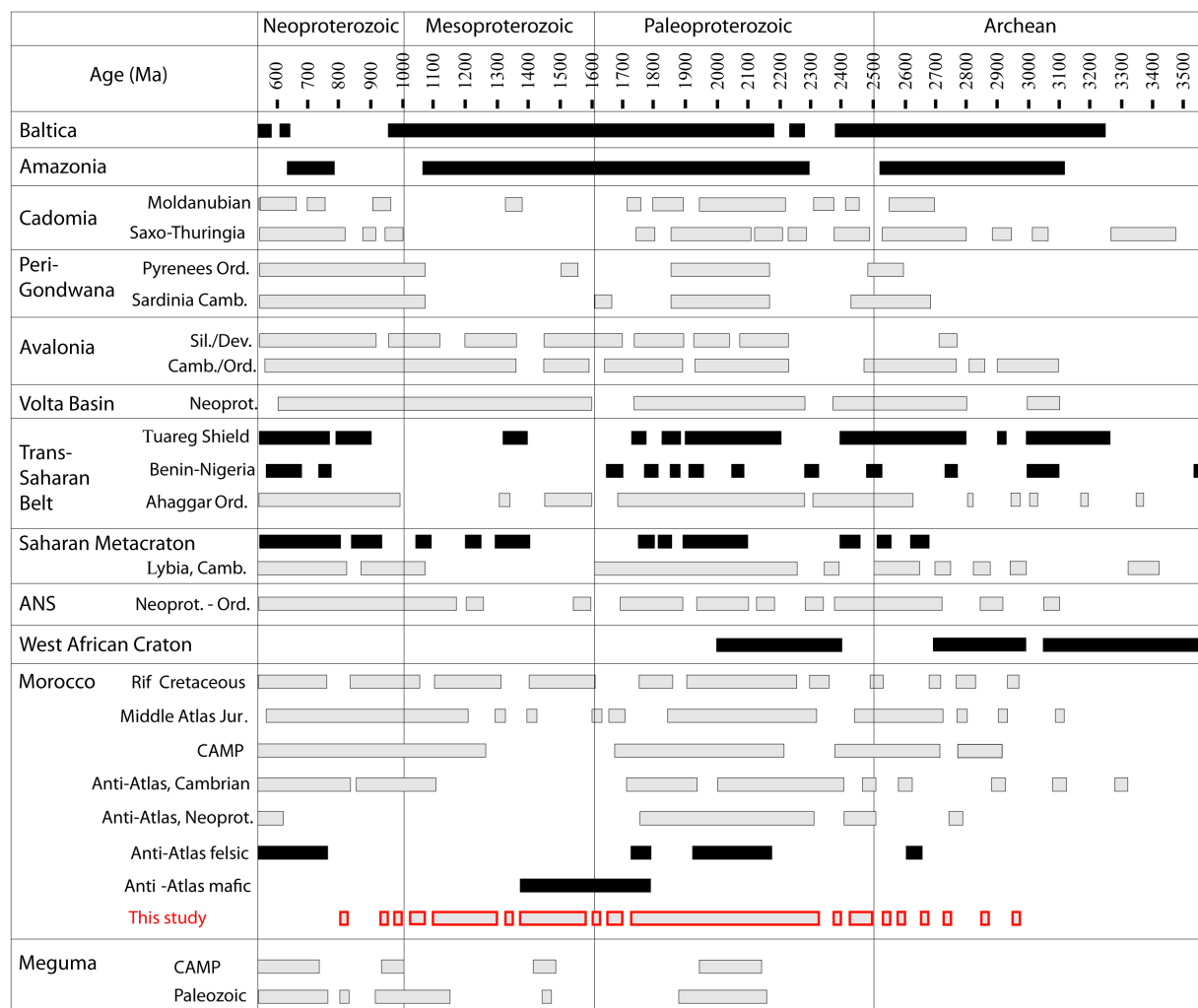


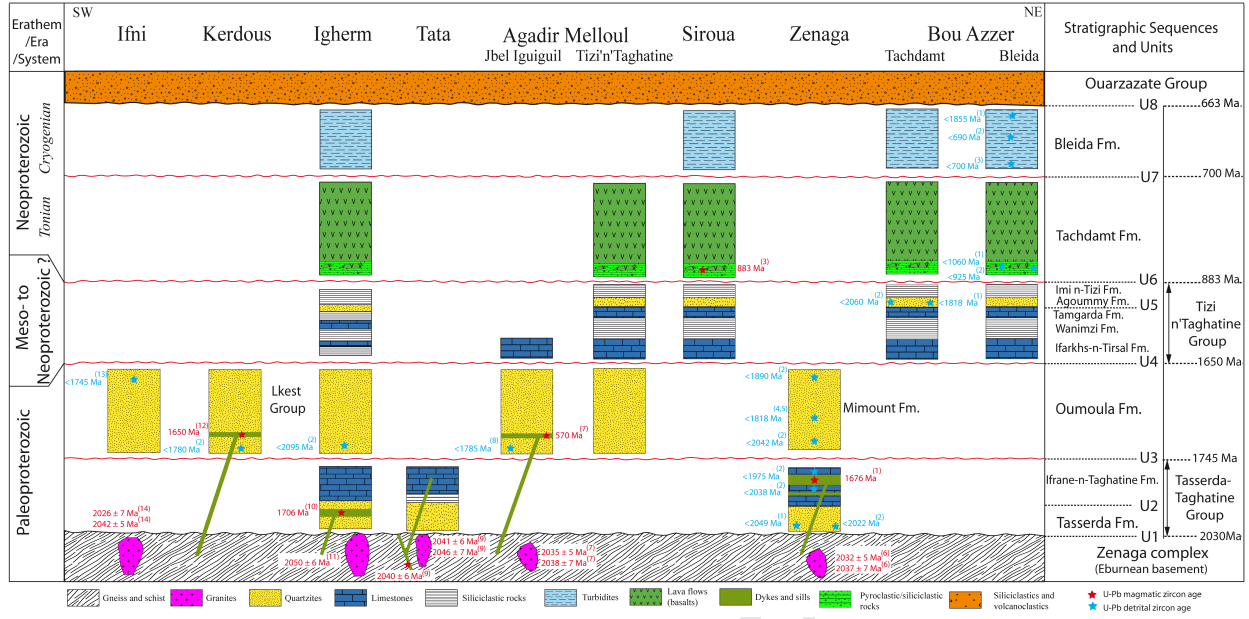


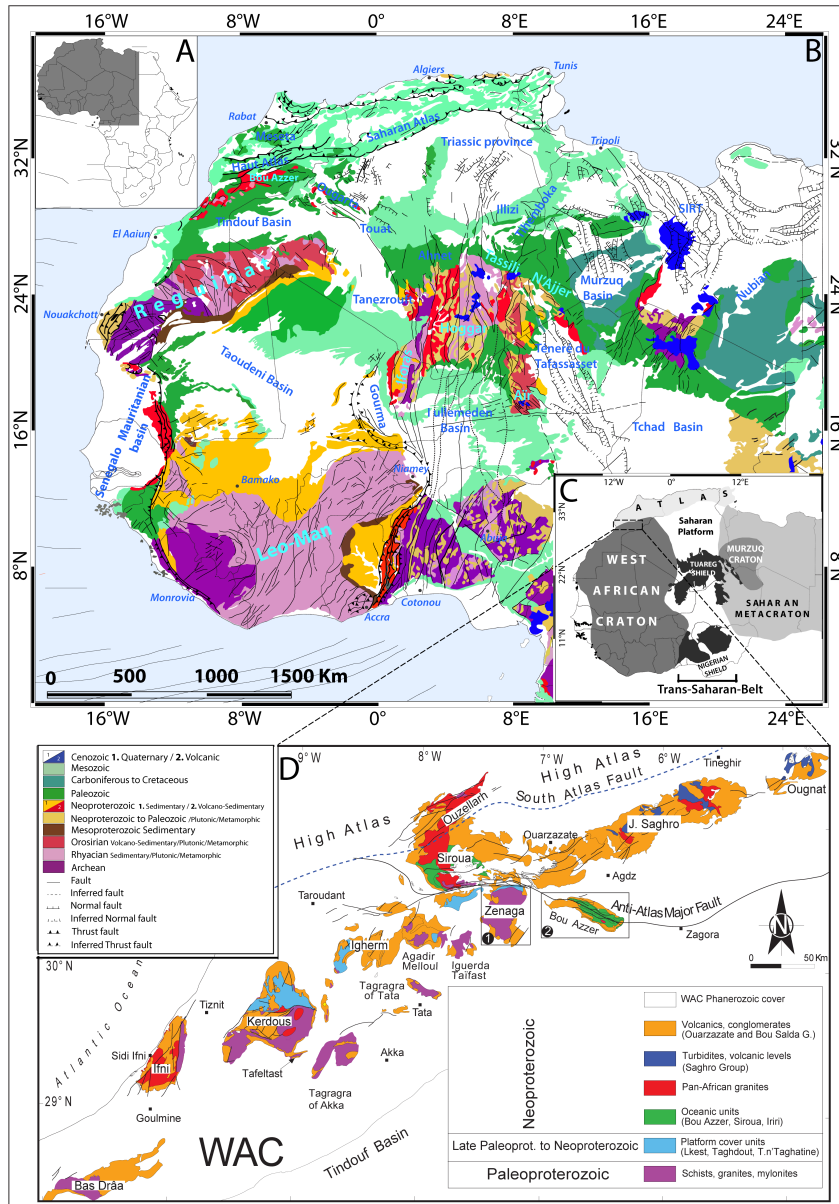


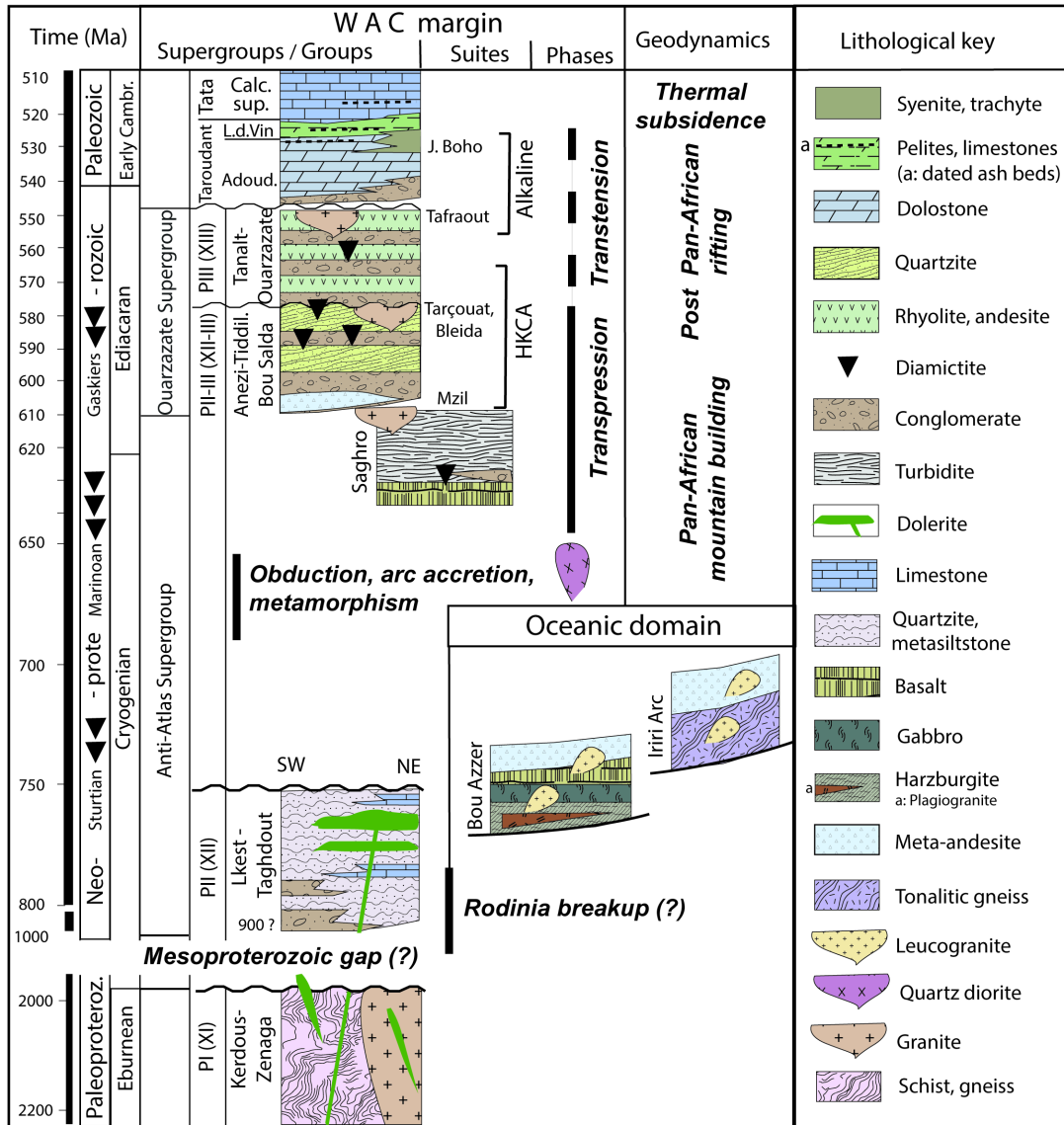




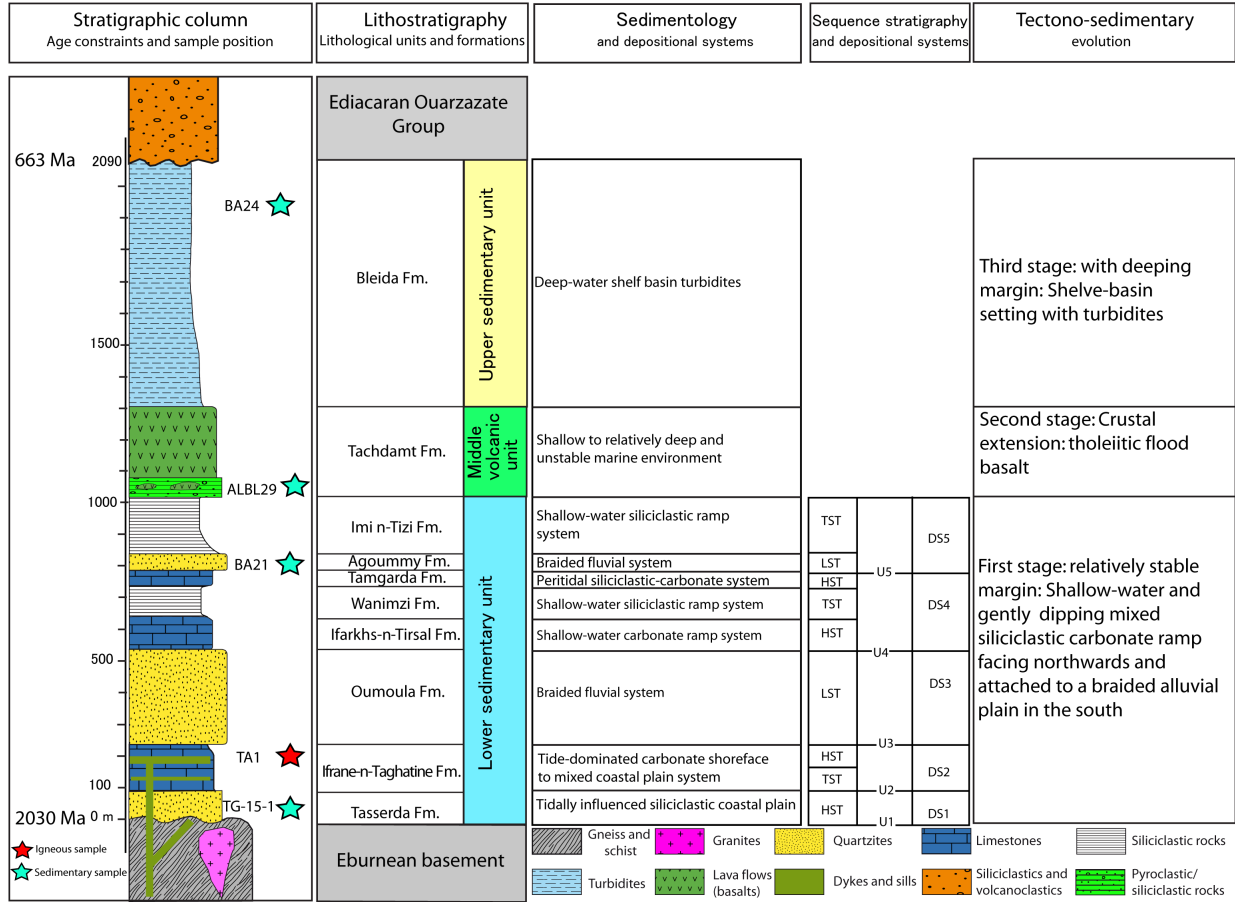


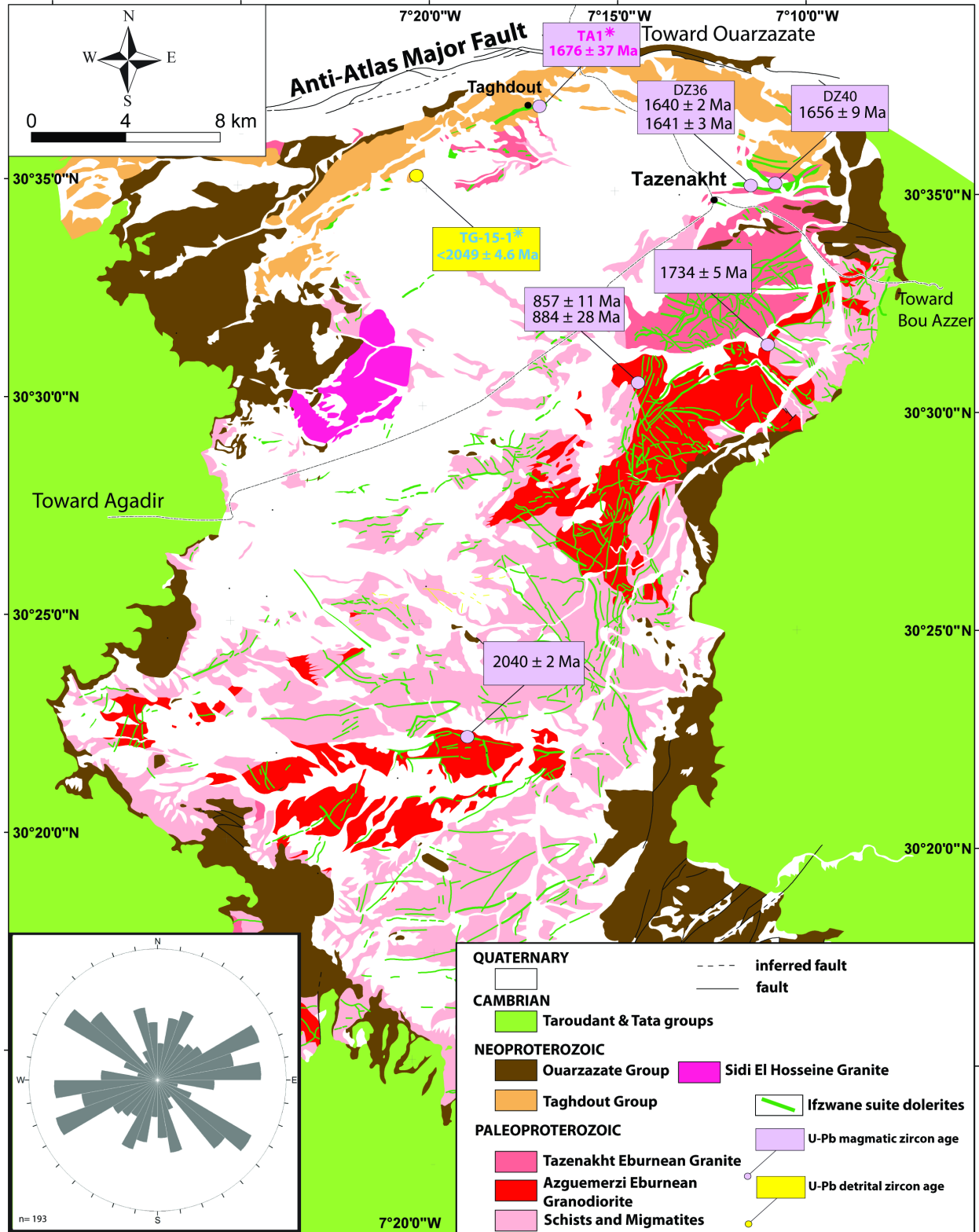


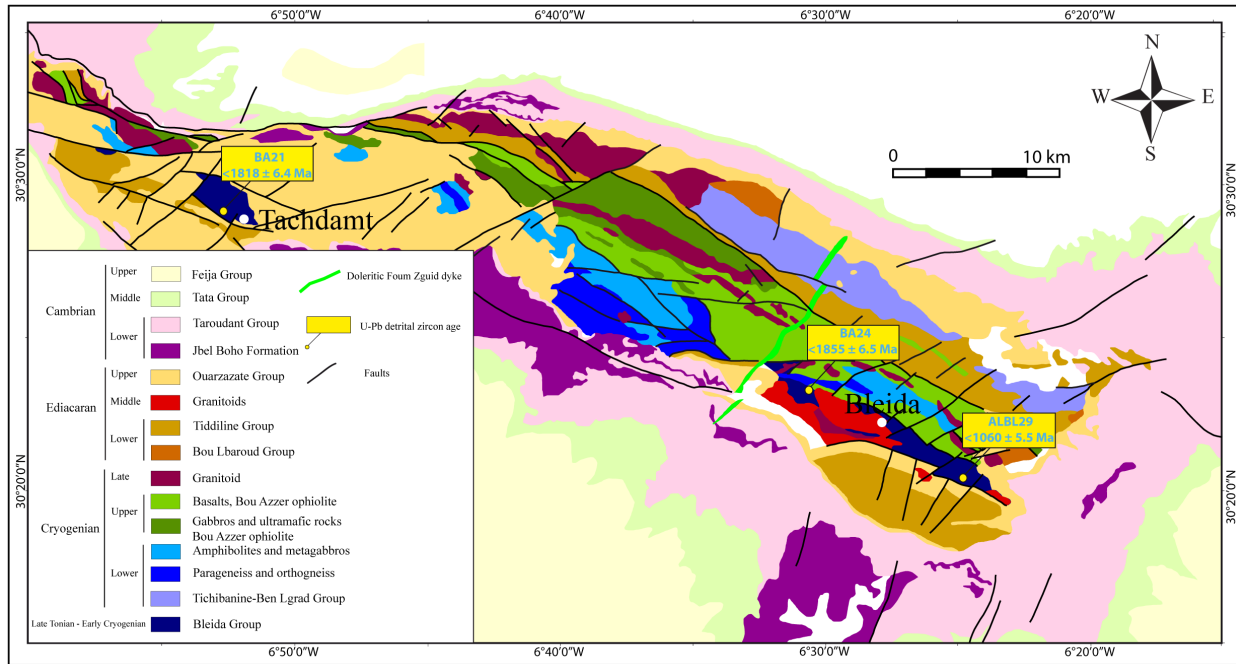


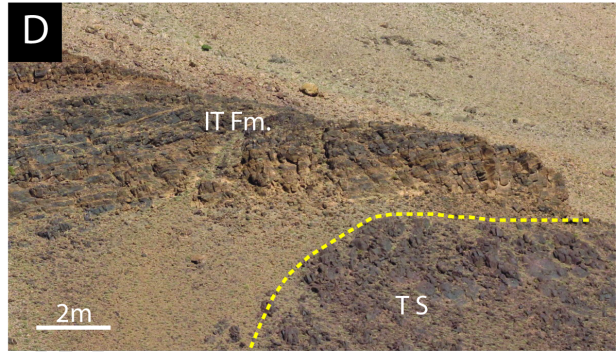
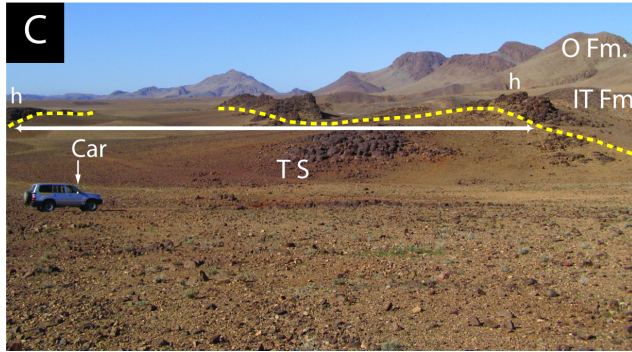
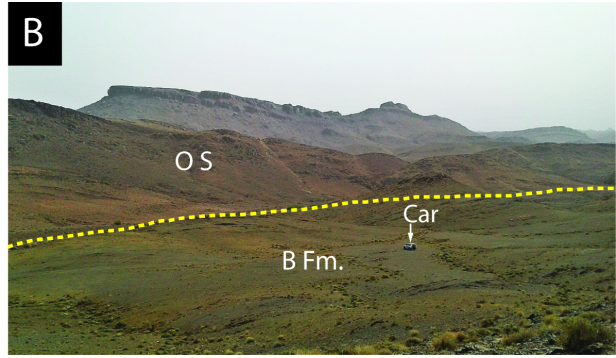
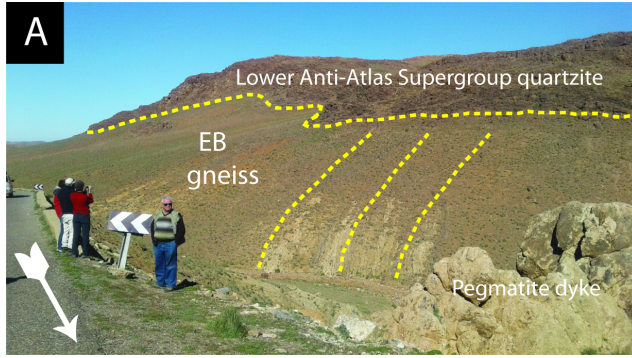




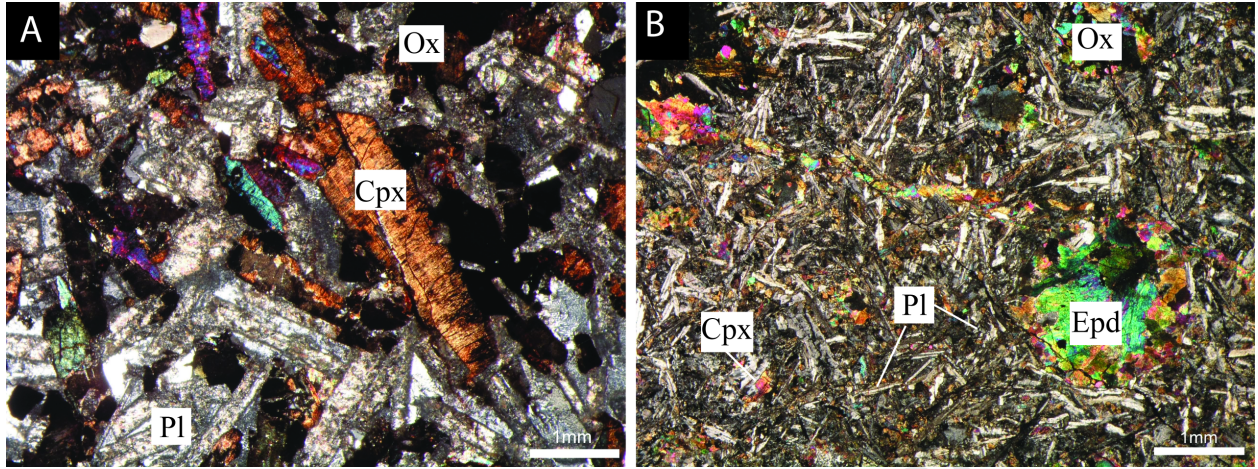


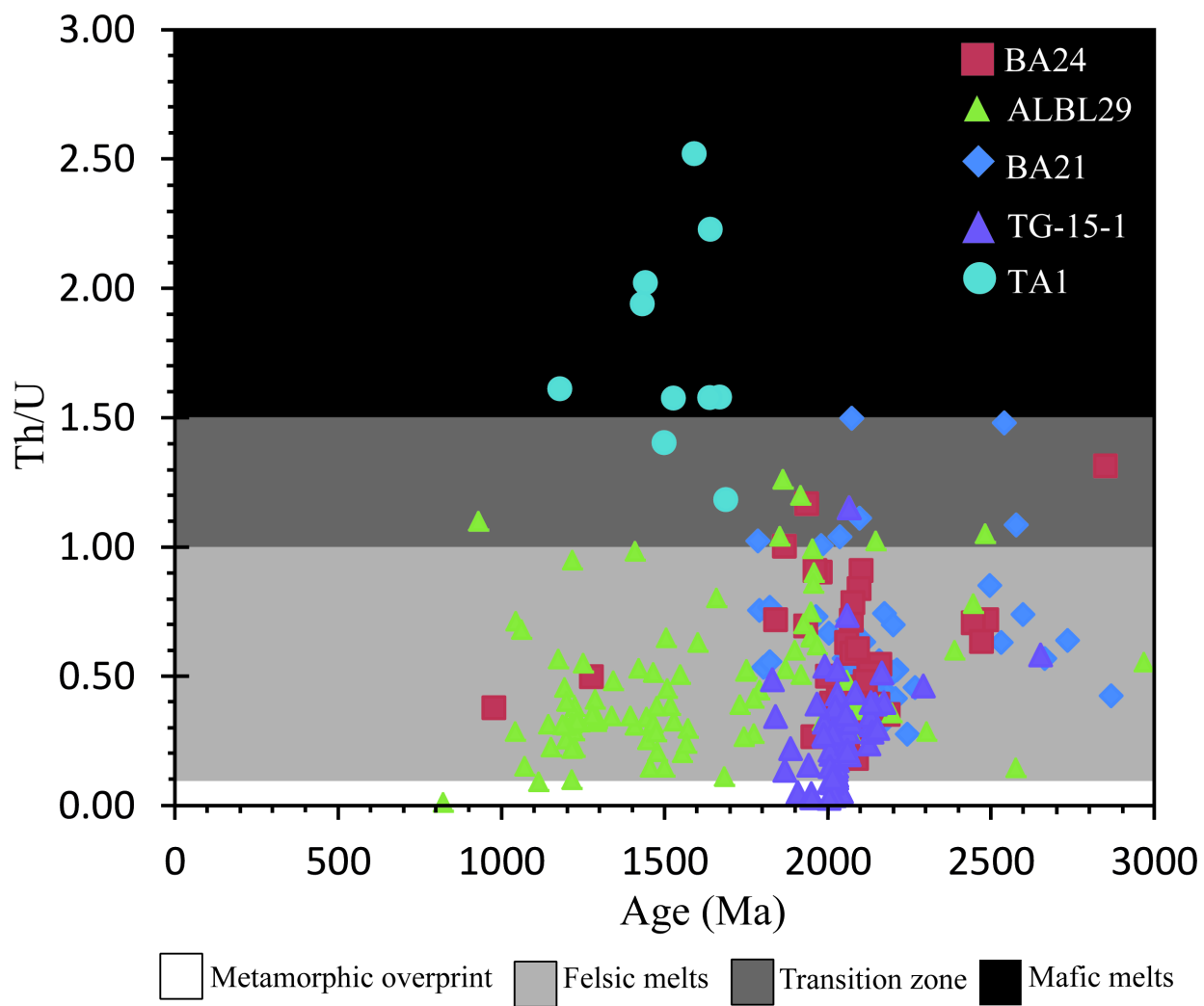


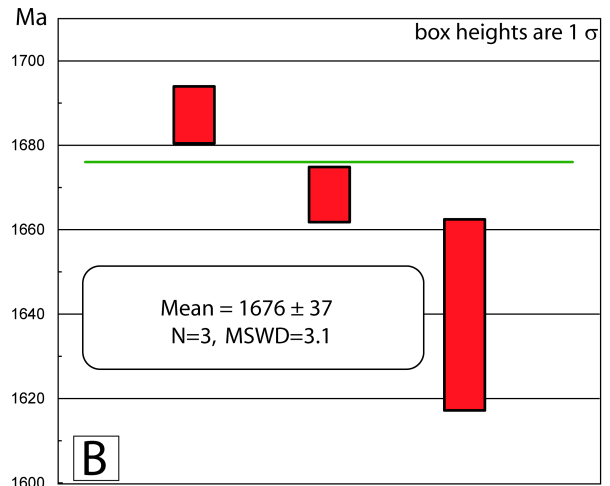
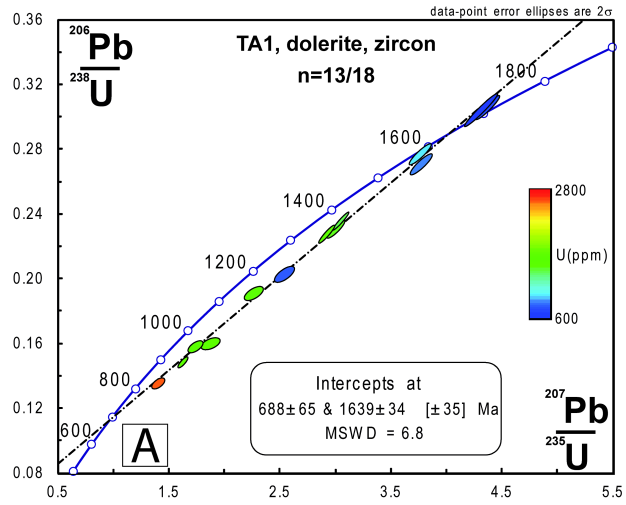




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## Highlights

- ▶ The Taghdout mafic sill of the Ifzwane Suite yielded an age of  $1676 \pm 37$  Ma
- ▶ The maximum depositional age of the lower Anti-Atlas Supergroup ranges from 2049 Ma (bottom) to 700 Ma (top) and the provenance is mainly from the WAC
- ▶ A new 5-part lithostratigraphic framework for the Proterozoic lower Anti-Atlas Supergroup of the Anti-Atlas Belt is proposed
- ▶ The 1650-1640 Ma Zenaga and 885-883 Ma Iguerda-Taïfast LIPs provide important stratigraphic constraints
- ▶ Correlation between the lower Anti-Atlas Supergroup and the Taoudeni Basin succession in Mauritania



**Declaration of interests**

The authors declare that they have no known competing financial interests or personal relationships that could have appeared to influence the work reported in this paper.

The authors declare the following financial interests/personal relationships which may be considered as potential competing interests:

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