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1 **Temporal and spatial changes of the submarine Cretaceous paleoslope in Northern**
 2 **Tunisia, inferred from Slump folds analysis**

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 16 **ABSTRACT**

17 This paper presents the first comprehensive, non-exhaustive, study of the genetic relationship
 18 between slump folds and the synsedimentary paleoslope during Cretaceous time in northern Tunisia.
 19 Slump folds occur mainly in the Cretaceous marl-dominated lithofacies, which exposes numerous
 20 slump folds structures. In addition, fault kinematic analysis is conducted to define the paleostress fields
 21 and the stress states characterizing the Cretaceous extension that triggers soft-sediment deformation
 22 and slumping. The MAM and the APM methods are used to deduce the paleoslope in several localities.
 23 The calculated values of paleoslope trend derived from MAM and APM methods precise the variation
 24 of the paleoslope trend during Cretaceous times in northern Tunisia. This paleoslope is ~NW-dipping
 25 during Berriasian, ~SSW-dipping during Valanginian, ~NW-dipping during the Barremian and ~N- to
 26 ~NNE or ~S- to ~SSW-dipping during Aptian-Albian period. The results of the back-tilted fault diagram
 27 show a ~North to ~Northeast-trending tectonics extension. The back-tilting of Cenomanian slump axis
 28 and poles of axial planes (MAM and APM methods) give close results with ~Southward or ~Northward-
 29 dipping paleoslope. The restored fault diagrams show ~North to ~Northeast-trending extension during
 30 Cenomanian times. Coniacian-Santonian marls deposits seal all the gravity-driven deformation
 31 structures. North Tunisian area exposes evidences for abundant soft-sediment deformation and
 32 slumping atop a northward facing submarine slope, which was probably dominant from the Early
 33 Cretaceous to Santonian with ~North-South tectonic extension related to the Southern Tethyan rifted
 34 continental margin evolution.

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 37 **Keywords:** Cretaceous; submarine paleoslope; slump folds; Tunisia

38 1. Introduction

39 Slumps folds, generated within unlithified sediments atop existent slope, form one of the most
40 remarkable features indicative of slope and basin-floor settings. Submarine slumping characterizes the
41 sediment-shape in both active and passive margins context (Woodcock, 1979, Strachan and Alsop,
42 2006; Alsop and Marco, 2014; Sharman et al., 2015). In passive margin framework, the driving force of
43 such structures is acknowledged to be the gravitational instability over a submarine slope. During
44 and/or after sedimentation, and before complete lithification, sediments moving downward (naturally
45 from continental shelf to basin) in successive depositional events. Subsequently, Slump horizons are
46 classified as primary structures that reflect paleoslope direction. Geoscientists have developed several
47 techniques, from slump folds, to deduce the direction of the paleoslope of both recent and ancient
48 margins (Heifetz et al., 2005; Strachan, 2008; Peel, 2014). North Tunisian area (Fig. 1) exposes
49 evidences for abundant soft-sediment deformation and slumping that characterized the evolution of
50 the Southern Tethyan rifted continental margin. However, to our knowledge, studies of the gravity-
51 driven deformational structures are still insufficiently characterized or even absent in this region.

52 By the evaluation and analysis of slump folds and associated soft-sediment structures, many
53 paleoslope and paleogeographical reconstructions were tested worldwide (Hansen, 1971; Woodcock,
54 1979; Debacker et al., 2009). Several techniques have been developed for the deduction of the
55 direction of the paleoslope from slump folds. Jones (1940) was the first who quantified the application
56 of this idea using the Mean Axis Method (MAM) based on the correlation of mean slump fold axis with
57 paleoslope strike in estimates of paleoslope direction (Fig. 2). Alternative method thereafter
58 designated by Hansen (1965, 1967) used the Separation Arc Method (SAM) is based on determining
59 the paleoslope direction from the bisector of a separation Arc named Separation Arc Method.
60 Woodcock (1979) reviewed these methods and suggested a procedure for determining paleoslope
61 directions which reflects the respective strengths of the two methods. More recently, other studies
62 have focused on the study of slump folds and described other methods to estimate the paleoslope
63 direction from these structures (Bradley and Hanson, 1998; Strachan and Alsop, 2006; Strachan, 2008;
64 Debacker and Meester, 2009; Alsop and Marco, 2011, 2012a, b, 2013, 2014; Debacker, 2012; Yong and
65 al., 2013; Sharman, 2014). These methods use well-determined parameters of slumps such as slump
66 fold axis, axial plane, interlimb angle...etc. Alsop and Marco (2012a) reviewed the methods (i.e. fold
67 facing directions, mean axis method -MAM-, mean axial plane strike method -MAPS-, mean axial-
68 planar dip method -MAD-, separation arc method -SAM-, axial planar intersection method -AIM-,
69 estimating the paleoslope direction from slumps folds structures (Fig. 2) with the contemporary
70 understanding. Authors gave a detailed description highlighting similarities and differences between

71 methods to deduce the direction of paleoslope from slump folds (for more detail see [Alsop and](#)
72 [Marco, 2012a](#)).

73 The analysis of slump folds and faults data together with sequences correlation are used in this study
74 to evaluate Cretaceous soft-sediment deformation in north Tunisia. This work is the first
75 comprehensive, non-exhaustive, study in north Tunisia applying known techniques to deduce
76 paleoslope of the Cretaceous basin based on the slump folds analysis. The work intention is to present
77 a structural contribution, via syndepositional deformation, to the debate about the basin
78 paleogeography which is usually studied in this basin by stratigraphy and/or sedimentology. The
79 northeast Tunisia exposes evidences for abundant soft sediment deformation features atop a
80 synsedimentary Cretaceous submarine slope. This study presents the compilation of 100 slump folds
81 collected along 54 km section from the Sidi Salem in the Southeast to the Jebel Boulahouadjeb in the
82 Northwest.

83 **2. Geological setting**

84 Northern Tunisia is a part of the northern edge of the African plate and represents the eastern part of
85 the Atlas System. The Atlas fold-and-thrust belt, extending from Morocco to Tunisia, is itself a part of
86 the alpine chain of Africa named “the Maghrebides belt” which is related to the Alpine orogeny. Tunisia
87 makes the eastern frontier of the Maghrebides belt. Except its northerly edge which corresponds to
88 the Numidian range, the Northern Tunisian Atlas ([Fig. 1](#)) is divided into two distinguished structural
89 units: (1) First, a major southeastward major thrust unit called Teboursouk thrust unit, which
90 corresponds to the front of the Alpine Range and shows a thick Aptian–Albian sequences. This area
91 exhibits also numerous outcropping salt structures belonging to the northeastern Maghreb salt
92 province ([Masrouhi et al, 2013](#); [Jaillard et al., 2013](#)), (2) second, the Zaghouan-Ressas unit,
93 acknowledged by last reviews to be the front of the Northern Tunisian Alpine Range. The present
94 Zaghouan Thrust Fault corresponds to an inverted inherited fault, which during Mesozoic time, makes
95 a paleogeographic line dividing a relatively shallow platform to the south, with a condensed Aptian
96 section from a deep basin in the north ([Morgan et al., 1998](#); [Soua, 2016](#)) with a thick Aptian–Albian
97 section ([Chihaoui et al., 2010](#)).

98 The present-day fold-and-thrust belt in northern Tunisia is the result of the Meso–Cenozoic geological
99 evolution of the northern African margin, which can be summarized in two main periods. The first is
100 the Mesozoic rifting, highlighted from Late Permian(?) to the Early Cretaceous ([Guiraud, 1998](#)). This
101 rifting is related to the Tethyan and Atlantic Oceans opening. Early Triassic salt basins were shaped,
102 overall Tunisia, during the earlier stage of this evolution ([Soussi et al., 2017](#)). The extensional setting
103 prevailed during Jurassic and Early Cretaceous ([Boughdiri et al., 2007](#); [Souquet et al., 1997](#); [Masrouhi](#)
104 [and Koyi, 2012](#); [Dahri and Boukadi, 2017](#)), led to the opening of Central Atlantic. Late reviews (e.g.

105 Guiraud et al., 2005; Gharbi et al., 2015) demonstrate a paleogeographic differentiation of the
 106 continental platform, which is accompanied in some localities by an associated volcanic and halokinetic
 107 activity (Mattoussi Kort et al., 2009; Dhahri and Boukadi, 2017, Jaillard et al., 2017). A major
 108 synsedimentary normal faulting delivering a general tilted blocks geometry, for which the Aptian–
 109 Albian ages are perceived to have the most extreme extensional related-structures time of the south
 110 Tethyan edge in North Tunisia (Souquet et al., 1997; Gharbi et al., 2013; Masrouhi et al., 2014b).
 111 Likewise, related Cretaceous salt movement confirms an hyperactive extension, thick and/or thin-
 112 skinned tectonic extension (Masrouhi et al., 2014a). The frequent NW- to- WNW-trending faults,
 113 related to N-to-NE-trending extension, are likely in control of an irregular sea-floor highlighted by
 114 significant thickness and/or facies changes during Cretaceous (Dhahri and Boukadi, 2017, Jaillard et al.,
 115 2017). The associated series show abundant slumps and syntectonic sequences (Gharbi et al., 2013).
 116 This margin is generally acknowledged, by all accounts of past reviews, to exposes a general NW-
 117 dipping submarine slope (Souquet et al., 1997; Masrouhi et al., 2013).

118 The second main period, started since Late Cretaceous (approximately Campanian–Maastrichtian),
 119 causes a basin positive inversion related to the Africa and Eurasia plate convergence (Guiraud and
 120 Bosworth, 1997; Gharbi et al., 2015). Two markedly compressional events affected the Atlas belt: the
 121 Atlassic mid-to-late Eocene contractional event (Frizon De Lamotte et al., 2002; Masrouhi et al., 2008;
 122 Dhahri and Boukadi, 2010) causes significant inversion of the basins, followed by the Miocene Alpine
 123 event which is the major orogenic contractional event that propagates deformation southward and
 124 responsible for the present-day NE-trending complex northern Tunisia Fold-and-Thrust belt (Chihi,
 125 1992; Dhahri and Boukadi, 2010; Gharbi et al., 2014). The Mesozoic basins of northern Tunisia are
 126 geometrically disturbed after having undergone the Neogene compressional tectonics that affected
 127 the northern African margin. This makes difficult the analysis of soft-sediment deformation and
 128 slumping that occurred previously to the tectonic inversion of this area.

129 **3. Tectono-sedimentary data**

130 To date and to precise/approximate the successive depositional events, detailed new key-facies
 131 description is used in this study, based on micropaleontological existent data (for more details on
 132 micropaleontological data, see Turki, 1988; El Ouardi, 1996; Rami, 1998; Masrouhi, 2006; Ben Fadhel
 133 et al., 2011; El khazri et al., 2013). Correlation of Cretaceous sequences is also significant because the
 134 facies and thickness of sediments are greatly affected by the synsedimentary faults activity of the faults
 135 bordering the adjacent tectonic blocks.

136 **3.1. Lithostratigraphy and correlation of measured sections**

137 The section of Jebel Oust (Fig. 3) is the reference type-section in northern Tunisia lower Cretaceous
 138 basin sequences (Souquet et al., 1997). It exposes a 2600 m thick Barremian–Aptian sequences.

139 According to [Souquet et al. \(1997\)](#), the Tithonian-Berriasian is marked by debris-flow and mud-flow
140 deposits. The Berriasian *p.p.* is composed of greenish muds and thin-bedded sandstones and siltstones.
141 The Valanginian-Hauterivian unit is made by a siliciclastic carbonate unit Valanginian in age and
142 acknowledged as Seroula Formation ([Khessibi, 1967](#)) and the Hauterivian deposits made by mixed shelf
143 deposits ended with carbonate platform.

144 The Barremian sequences began with mixed tempestites, followed by a thick carbonate shoals with
145 reworked corals, topped by storm deposits within offshore muds, themselves followed by anoxic marl-
146 mudstone parasequences ([Souquet et al., 1997](#)). In northern Tunisia, the Valanginian–Aptian open
147 marine shaly series are attributed to the M'Cherga Formation ([Buroillet, 1956; Ben Ferjani et al., 1990](#)).

148 The Aptian–Early Albian sequences are relatively made of deep-water type showing gray to black
149 marls. The Aptian series in this region present frequent traces of lamination which is the witnesses of
150 the local thickness reduction of these series ([Turki, 1980](#)). In the northern portion of Jebel
151 Boulahouadjeb, the Aptian marls are covered by Triassic salt suggesting an active salt tectonic period.
152 In addition, the Aptian sequences show abundant slumps and calcareous nodules clearly linked to a
153 contemporaneous sub-marine slope. Furthermore, salt tectonics which is active during this period, and
154 responsible for the emplacement of large submarine salt sheet and salt rise can also enhance this slope
155 ([Masrouhi et al., 2013](#)).

156 The Correlation of the Lower Cretaceous series ([Fig. 3](#)), drawn from Jebel Sidi-Salem in the Southeast
157 to Jebel Boulahouadjeb in the Northwest, shows a significant thicknesses changes ranging from tens
158 to hundreds of meters but with preservation of the lithology which remains in all sections a pelagic to
159 hemi-pelagic facies.

160 A significant thickening is perceived at Jebel Oust (2600 m-thick Barremian–Aptian series) and Jebel
161 Boulahouadjeb (2000 m-thick Hauterivian–Barremian series), however these series are relatively
162 reduced with 310 m at Jebel Sidi-Salem and 450 at Oued Tazega ([Fig. 3](#)). Jebel Boulahouadjeb was,
163 likely, a significant depocenter with a very high rate of sedimentation, created in response to active
164 and extensional Cretaceous tectonics ([Souquet et al., 1997; Masrouhi et al., 2013](#)). Furthermore, the
165 series outcropping from either side of the Zaghuan fault reflects a reduction in thickness toward the
166 southeast compared to the Jebel Oust's type-section. Apparently, no significant flexure has worked at
167 the base of the current Zaghuan fault during the Barremian, but we are led to imagine a
168 synsedimentary flexure or rather a set of flexures further west separating subsiding areas of the
169 Tunisian trough (Jebel Oust) from other edge series with reduced thickness ([Turki, 1988, El khazri et
170 al., 2013](#)).

171 The correlation, of the Lower Cretaceous sequences, likely reflects an irregular sea floor probably
172 created by tectonic extension ([Masrouhi et al., 2013](#)). The Lower Cretaceous extension is also testified

173 by numerous synsedimentary features (nodule reworking, slumps, olistholites, active salt tectonics...).

174 The Lower Cretaceous deposits in central and northern Tunisia can be divided into four second order

175 cycles. The first two cycles extending from Late Tithonian to Hauterivian are included within the rifting

176 phase where basement block-faulting predominates (tilted blocks). Third and fourth cycles extend

177 from Barremian to Albian are likely produced by the intra-basin growth faulting (Souquet et al., 1997).

178 Upper Cretaceous deposits show, in all northern Tunisia, homogeneous facies mainly composed of

179 marls and fine-grained limestones (Fig. 4). The Albian-Cenomanian series are defined as Fahdene

180 Formation wholly formed by limestone-marl alternations and topped by the thinly laminated organic-

181 rich limestones of Bahloul Formation, dated as Uppermost Cenomanian-Turonian (Soua and

182 Tribovillard, 2007). Over the Fahdene Formation, rest a thick series of gray marl and shale interbedded

183 with fine-grained limestones. These series are attributed to the Turonian–Early Campanian Aleg

184 Formation (Buroillet, 1956). The Campanian–Maastrichtian series (Abiod Formation) are essentially

185 made by chalky limestones.

186 The correlation of the upper Cretaceous series (Fig. 4) shows no significant variation in facies (all

187 sequences are of pelagic to hemi-pelagic facies; Florida, 1963; Ben Yagoub, 1978, Turki, 1988; Soua

188 and Tribovillard, 2007). However, clearly considerable thickness variation is perceived along the

189 northern Tunisia. From Upper Albian to Coniacian-Santonian, thicknesses significantly vary from tens

190 to hundreds of meters for the equivalent formations from section to other. This correlation clearly

191 reflects a persisted tilted blocks geometry, revealed for the Early Cretaceous, and likely shaped by

192 major synsedimentary normal fault systems.

193 **3.2. Slumping and faulting analysis**

194 **3.2.1. Dataset and methodology:**

195 To study the tectono-sedimentary evolution and the related submarine paleoslope of the southern

196 Tethyan margin in the northern Tunisia, field data collection and detailed geologic mapping have been

197 carried out. Stereonet 9 (Allmendinger et al., 2013; Cardozo and Allmendinger, 2013), failles software

198 (Carey, 1979; Carey and Brunier, 1974) and canvas drawing software have been used in data plotting

199 and analysis. Particularly, two elements are surveyed: the faults kinematics related to the the tectonic

200 regime and slump folds analysis which supports the reconstruction of the paleoslope orientation

201 (Alsop and Holdsworth, 2007; Van Loon, 2009; Lacelle, et al., 2015).

202 The brittle deformation is frequently quantified using fault kinematic analysis methods. These methods

203 are based on measurements of centimetric to metric scales faults planes and striae. Fault kinematic

204 analysis commonly determines the reduced stress tensor and the directions and plunges of principal

205 stresses (σ_1 , σ_2 , σ_3). This inferred stress tensors reflect a local state of stress. To deduce the original
206 stress tensors, we rotated all fault data to restore the bedding plane to its horizontal orientation.

207 Several techniques have been developed for the deduction of the paleoslope direction from the study
208 of slump fold axis. Jones (1940) was the first who applied this idea quantitatively. Recently, other
209 studies have focused on the study of slump folds and have described other methods estimating the
210 paleoslope direction from these structures (e.g. Hansen, 1965, 1967; Woodcock, 1979; Bradley and
211 Hanson, 1998; Alsop et al., 2001; Strachan and Alsop, 2006; Strachan, 2008; Debacker and Meester,
212 2009; Alsop and Marco, 2011, 2012a, b, 2013, 2014; Debacker, 2012; Yong et al., 2013; Sharman, 2014).
213 Two methods are used in this study (Fig. 2): the MAM (Mean Axis Method) and the APM (the Axial-
214 Planar Method).

215 The MAM, proposed initially by Jones (1939), assumes that slumps fold axes are statistically parallel to
216 the paleoslope and perpendicular to the translation direction of the sheet. To apply this method, we
217 need to rotate either the original data, or at least the derived paleoslope directions, in order to remove
218 later tectonic effects. This method brings about two diametrically opposed possibilities for the
219 direction of the paleoslope (Fig. 2). The latter is usually chosen basically on regional paleogeographic
220 constraints (Jones, 1939), and/or the vergence of slump folds (Woodcock, 1976b). The MAM has the
221 advantage of being statistically robust because it is based on average properties of slump
222 measurements. The major disadvantage of the MAM is that it does not allow for fold axes that are
223 oriented in parallel or oblique way to the downslope direction (Woodcock, 1979). The APM, treating
224 the Fold axial surfaces (approximated as axial planes), yields important kinematic information in a
225 similar way as fold axes (Woodcock, 1976a, 1979b; Farrell and Eaton, 1987, Sharman, 2014). Through
226 APM (Fig. 2), poles to axial planes tend to align in a great circle around the mean fold axis, and thus
227 the plane through the best-fit girdle to these poles gives a bi-directional estimate of the slumps
228 transport direction (Woodcock, 1976b, Sharman, 2014). Furthermore, the imbrication of the mean
229 axial plane with respect to the global slumps orientation can be used to evaluate the paleoslope dip
230 direction (Woodcock, 1976b).

231 3.2.2. Results of synsedimentary faults analysis

232 The current study, of faults and slump folds, is based on new structural data collected from seven
233 localities which are respectively from the Southeast to the Northwest the Jebel Sidi Salem – Messella,
234 Jebel Oust, Jebel Rihane, Jebel Kechtilou, Oued Tazegga, Jebel Bechtab and Jebel Boulahouadjeb.

235 The Sidi Salem–Messella is a NE-trending structure located along the Zaghouan–Ressas thrust system.
236 The Present survey together with previous works (Turki, 1988; Morgan et al., 1998) affirms that this
237 present-day thrust system (Fig. 5a) is the result of the inversion of major inherited normal faults. Field
238 data shows a layer's geometry of growth strata likely related to an early normal faulting acting during

239 Aptian-Albian times and as well probably active before. Likewise, Aptian-Albian sequences display
240 abundant conglomeratic horizons and slumping, in addition to significant thickness changes. Besides,
241 the Sidi Salem–Messella structure shows abundant centimetric, metric to decametric-scale sealed
242 normal faults (Fig. 5). The collected faults data are rotated, using fault diagram, to their original
243 orientation to restore initial tectonic extension. The back-tilting, of faulting, highlights a normal fault
244 type predating the Cenozoic folding (Fig. 5). This kinematics analysis of normal faulting (Fig. 5b) typifies
245 a minimum stress axis σ_3 -trending $\sim N25^\circ E$ during the Aptian time. During Albian times, the tectonic
246 regime is characterized by $\sim N201^\circ E$ -trending minimum stress (σ_3) axis (Fig. 5c). This setting was likely
247 responsible for half-graben geometry structure of the Jebel Sidi Salem–Messella area.

248 15 km northward, the Jebel Oust is a 15 km-large $N40^\circ E$ -trending anticline structure with Jurassic to
249 Maastrichtian outcrops (Fig. 6a). In this locality, considered as the type-section of Lower Cretaceous
250 (Ben Ferjani et al., 1990; Souquet et al., 1997), 2600 m-thick siliciclastic and marly sequences have
251 been deposited. The Barremian-Aptian series show numerous small-scale synsedimentary normal
252 faults preserved within siliciclastic sediments. The Barremian sediments are affected by normal
253 faulting (Fig. 6b) locally characterized by $\sim N272^\circ E$ -trending minimum stress axis (σ_3). During Aptian
254 times, the Jebel Oust area seems also to be controlled by the same fault population acting as normal
255 faults with an extensional trending $\sim N251^\circ E$ (Fig. 6c, d).

256 The Jebel Rihane corresponds to a WNW-to NW-trending large syncline structure (Fig. 7a) situated
257 eastern of Jebel Oust structure and it is mainly made of Upper Cretaceous deposits (Arfaoui, 2007).
258 This structure seems to be controlled essentially by E-to NE-trending faults. Locally, field results
259 highlight that most faults have steep planes (Fig. 7b, c) suggesting that they occurred in a local
260 transtensive stress regime which controlled the sedimentation in this small area since the Albian times.
261 The back-tilted Albian fault diagram shows a transtensive tectonic regime with $\sim N354^\circ E$ trending
262 tectonic extension.

263 The Jebel Kechtilou is a NE-trending faulted anticline (Fig. 8a). This is one of numerous structures
264 disturbed during Cretaceous rifting by the ascendant movement of Triassic evaporites. Active
265 extensional tectonics recorded all along the Jebel Kechtilou within Aptian-Albian deposits (Haggui,
266 2012). The Albian series show number of a mesoscale synsedimentary normal faults. The back-tilted
267 fault diagram shows $\sim N359^\circ E$ trending tectonic extension during Albian time (Fig. 8b). In addition,
268 numerous metric Aptian 'olistostromes' are interpreted to be derived from submarine gravity sliding
269 in Albian series (Fig. 8c). Frequent reworked blocks, slumps, and nodules joined the Aptian-Albian
270 extensional tectonic regime. Furthermore, these tectonic events induced variation in thickness (Fig.
271 8d) and facies of Cretaceous series prior to the tectonic inversion.

272 The Jebel Bechtab located in the Lansarine belt (Fig. 9a) shows a NNE-trending faulted structure,
273 intersected by Triassic evaporitic bodies. This structure is made of Hauterivian-Barremian to Oligo-
274 Miocene sequences (Masrouhi et al., 2013). Mains faults observed in this zone are NW-dipping normal
275 faults. Detailed observation of fault planes argues for an early normal displacement before Cenozoic
276 compression which reverse them as strike-slips. Albian fault population of Glib El Abiod locality (Fig.
277 9b) indicates a normal faulting resulting from extensional tectonic regime locally characterized by
278 ~N344°E-trending minimum stress axis (σ_3). During Cenomanian times, the Bechtab area is also
279 controlled by comparable fault population acting as normal faulting with ~N002°E-trending
280 extensional tectonic (Fig. 9c).

281 3.2.3. Slumps analysis results

282 The Lower Cretaceous sequences are marl-dominated and deposited in an unstable basin floor
283 affected by extensional tectonics. In such conditions, the instability of locally oversteepened slopes
284 together with seismotectonic activity can trigger slumps within shallow marls and mudrocks before
285 they are lithified. Numerous slump folds have been identified. After rotation of bedding data to restore
286 the bedding planes in its horizontal orientation, these deformations are observed only within a
287 restricted stack of marly strata argue for synsedimentary slump folding and not micro and/or
288 mesofolding resulting from permanent compressions.

289 21 slump folds were studied in the Jebel Oust locality, which are recorded throughout Berriasian,
290 Valanginian and Barremian sequences. Slump folds of Jebel Oust are upright to recumbent folds
291 (Fig.10). Both MAM and APM methods applied for 10 samples of slumps observed within the Berriasian
292 series gives ~N331 trending paleoslope (Fig.10a). For the Valanginian slump folds, the use of the APM
293 and MAM reveals ~SSW-mean downslope transport direction. The back-tilting of 6 slump folds from
294 the Valanginian siliciclastic carbonates deposits using MAM and APM methods give very close values
295 of paleoslope trend, respectively ~N202 and ~N201 (Fig. 10b). The Barremian sequences are topped
296 by a carbonate level bar which is formed by Black nodular limestones followed by thick fine clayey
297 limestones and green marls alternations. Within this carbonate level bar 5 slump folds are measured
298 and unfolded-back to their horizontal position. Results of the determination techniques display a
299 general ~NW-dipping paleoslope. The back-tilted axis gives ~N310 value and the back-tilted poles of
300 axial planes gives ~N318 value (Fig. 10c).

301 47 slump folds within the Aptian-Albian series are analyzed in this work. Slump folds are observed
302 within green-black marls series interbedded by fine marly limestone in the base and within the middle
303 marls interbedded with black clayey limestone. All of them are topped by clayey platy limestone with
304 dark gray marls of Upper Albian. Abundant slump folds are accompanied by frequent resedimented
305 blocks and nodules. Albian epoch is considered by authors (e.g. Saadi et al., 1994; Masrouhi et al.,

2014a, Jaillard et al., 2017) as a maximum regional extension period of the Cretaceous rifting in northern Tunisia. Slump folds' geometries, observed in the studied sections, suggest a wide fold types ranging from open hinge folding to steeply inclined (Figs. 11, 12). In Jebel Sidi Salem–Messella, 25 slump folds are analyzed. The Analysis of unfolded slump axis gives ~N030-trending paleoslope, whereas the unfold poles of axial planes (APM method) give comparable value of ~N025-trending paleoslope (Fig. 11a). These results are in accordance with a general Aptian–Albian NNE-trending paleoslope. Westward, slump folds belonging to the Jebel Rihane structure show a general ~SSE-trending mean slump transport direction. The computed values of paleoslope trend derived from MAM and APM methods are: ~N150 and ~N160 respectively (Fig. 11b). Northward, in Jebel Kechtilou numerous slump folds are measured within the Aptian–Albian marls and limestones alternations. Using the two methods above, a general ~NNE-trending paleoslope of the basin floor seems to be dominant during Aptian–Albian times. The back-tilted axis method gives a paleoslope trend of ~N017, and, the back-tilted poles of axial planes gives ~N030 (Fig. 11c). The structures of Jebel Bechtab, Oued Tazegga and Jebel Boulahouadjeb lie on the northern tip of the study area. The Cretaceous series of these structures are characterized by larger, sometimes decametric, slump folds. The pre-tilted slump axis and poles to axial surfaces give a general ~S-trending paleoslope (~S, ~SSE, and ~SSW respectively in the Jebel Bechtab, Oued Tazegga and Jebel Boulahouadjeb). The following results are obtained: in the Jebel Bechtab, the back-tilted axis gives ~N175-trending paleoslope, and, the back-tilted poles of axial planes give ~N173-trending slope (Fig. 12a). In the Oued Tazegga, the results of two MAM and APM methods are respectively ~N151 and ~N156-trending paleoslope (Fig. 12b). In the Jebel Boulahouadjeb structure, the analysis of unfolded slump axis gives ~N202-trending paleoslope, whereas the unfold poles of axial planes (APM method) give a ~N205-trending paleoslope (Fig. 12c). The Cenomanian series show an abundant instability features classified also to be soft-sediment deformation related to the Cretaceous extension in north-eastern Tunisia. In this area, 32 slump folds were analyzed in the Jebel Sidi Salem–Jebel Messella, Jebel Kechtilou and Jebel Bechtab structures. The analysis of slump folds' data shows a general N- downslope transport direction. The unfolded slump axis (MAM Method) gives ~N004-trending paleoslope, and, the unfold poles of axial planes (APM method) show a similar paleoslope direction with ~N008-trending paleoslope (Fig. 13a). Westward, In Jebel Kechtilou, a general ~S-dipping paleoslope is highlighted. The Analysis of unfolded slump axis (MAM method) gives ~N196-trending paleoslope, and the unfold poles of axial planes (APM method) gives ~N186-trending paleoslope (Fig. 13b). Northward, in the northern flank of Jebel Bechtab (Glib El Abiod locality), Cenomanian series exhibit well identified slump features. The back-tilted slump axis and poles of axial planes give very close results highlighting a general ~S-dipping paleoslope. The MAM and APM methods give respectively ~N172 and ~N167-trending paleoslope (Fig. 13c).

340 **4. Discussion:**

341 North Africa is part of the Southern Tethyan rifted continental margin during Mesozoic times. The rift
342 stage, started during Permian (?)–Triassic, is followed during Jurassic and Cretaceous times by passive
343 margin basin's style (Tlig, 2015). During the syn-rift and post-rift stages, which is associated with the
344 opening of the Tethyan and Central Atlantic oceans, north Tunisia area corresponds to basin that was
345 filled with ~5-to 6-km-thick sedimentary succession. Although the geodynamic evolution of the
346 southern Tethyan margin in Tunisia have been considered by several authors, the gravity-driven
347 deformational structures are still insufficiently studied.

348 The Lower Cretaceous series, showing a significant thicknesses variation, were likely deposited above
349 an irregular sea floor. Synsedimentary features (slumps, sealed normal faults, conglomerates,
350 thickness variation...) testify for Mesozoic extensional tectonic regime and display character of related
351 growth strata. The basin is itself part of the Southern Tethyan rifted continental margin (Basilone et
352 al., 2014; Basilone and Sulli, 2016). Using slump folds transport determination techniques (MAM and
353 APM), the basin paleoslope exhibit a variable trend through Early Cretaceous times (Table 1); this
354 paleoslope is ~NW-dipping during Berriasian, ~SSW-trending during Valanginian and ~NW-dipping
355 during the Barremian. During Aptian-Albian period, perceived to have the most extreme extensional
356 related-structures time of the south Tethyan edge in North Tunisia (Masrouhi et al., 2013; Dhahri and
357 Boukadi, 2017, Jaillard et al., 2017), a significant paleogeographic differentiation is identified in Tunisia.
358 The direction of paleoslope inferred from slump axis and from poles of Axial planes confirms this
359 paleogeography by giving ~N- to ~NNE or ~S- to ~SSW-dipping paleoslope (Table 1). Fault kinematics
360 deduced from the study of back-tilting striated fault planes display evidence for ~N-S to ~NE-SW
361 direction of the tectonic extension. Based on this study and numerous previous paleogeographic data
362 (e.g. Souquet et al., 1997; Ben Fadhel et al., 2011; Soua, 2016), the Lower Cretaceous deposits of the
363 Tunisian Atlas can be divided into four second order cycles. The first two cycles extending from Late
364 Tithonian to Hauterivian are included within the rifting phase where basement block-faulting
365 predominates (tilted blocks). Third and fourth cycles extend from the Barremian to Albian are likely
366 produced by the intra-basin growth faulting (Fig. 14). Since the Cenomanian, the facies heterogeneity
367 decreases. The homogeneity of the Upper Cretaceous facies likely reflects a basin post-rifting
368 subsidence history. The back-tilted fault diagrams show ~N-S to ~NE-SW extension during Cenomanian
369 time which indicates that the Albian E- and NW-trending faults remain active during Cenomanian. This
370 faults system is likely an inherited Albian faulting that controls a general northward basin's submarine
371 slope (Fig. 14). The back-tilted slump axis and poles of axial planes (MAM and APM respectively) give
372 very close results with a ~S or ~N-dipping paleoslope which confirms the ~N to NE-trending extension
373 during Cenomanian time. During Upper Cretaceous, an extensional period of post-rift stage is

374 highlighted, showing pelagic deposits controlled probably by the isostatic response of different major
375 fault systems and/or depocenters (Gharbi et al., 2013). Coniacian–Santonian deposits seem to seal all
376 the aforementioned differentiation and described as post-rift marl-rich sequences followed by
377 limestone and marl sequences approximately homogeneous in the entire basin. During the Upper
378 Cretaceous, this region was a transtensive paleomargin, at least until earliest the Campanian
379 (Masrouhi et al., 2008, Gharbi et al., 2013).

380 5. conclusion

381 The analysis of slump folds and the synsedimentary normal faults data together with sequences
382 correlation permits us to precise the variation of the paleoslope trend during Cretaceous times in
383 northern Tunisia. In detail, this paleoslope is ~NW-dipping during Berriasian, ~SSW-trending during
384 Valanginian and ~NW-dipping during the Barremian and ~N- to ~NNE or ~S- to ~SSW-dipping during
385 Aptian-Albian period. In this area, passive margin gravity deformations occurred during Lower
386 Cretaceous and continued at least to the Cenomanian. During Cenomanian, the inherited fault systems
387 likely active since Albian controls a general northward basin's submarine slope. with a ~S or ~N-dipping
388 paleoslope. Abundant soft-sediment deformations and slumping are shaped onto a regional
389 northward to southward facing submarine slope that reflect the intra-basin growth faulting associated
390 to Triassic salt rise up during Cretaceous times. The extensional regime that qualified this is responsible
391 for the development of major synsedimentary normal fault systems producing tilted blocks basin
392 geometry characterizing a general ~North-South tectonic extension, related to the Southern Tethyan
393 rifted continental margin.

394

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401

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- 571

572 **Figures captions**

573 **Figure 1.** Location of the studied sections in northeastern Tunisia. The inset on the top left shows the
574 location of Tunisia in North Africa.

575 **Figure 2.** fold's patterns in slump sheet. A) Equal area plot of hypothetical fold axis and poles of axial
576 planes distribution to illustrate the mean axis, the separation, and the axial plane methods. The best-
577 fit girdle to axes and poles to axial planes of the slump sheet are horizontal. The paleoslope dips to
578 south (modified from [Woodcock, 1979](#)). B) folding generating during layer-parallel shearing (a) and
579 layer-normal shearing (b), see Alsop and Marco, 2011 for more detail, and, associated facing direction
580 (for more detail see Alsop and Marco, 2011, 2012a, b). C) fold facing direction in schematic illustrations
581 summarizing the relationships between folds vergence, axial plane and up-ward or downward facing
582 states (see Alsop and Marco, 2012a).

583 **Figure 3.** NE-SW lithostratigraphic correlation of Lower Cretaceous series with the Jebel Oust type-
584 section in northeastern Tunisia.

585 **Figure 4.** NE-SW lithostratigraphic correlation of Upper Cretaceous series in northeastern Tunisia.

586 **Figure 5.** A) detailed geologic map of Sidi-Salem–Messella structure ([Gharbi et al.](#), forthcoming). B)
587 Synsedimentary Aptian normal fault with lower hemisphere stereograms projection of back-tilted
588 striae. C) Synsedimentary Albian normal faults with lower hemisphere stereograms projection of back-
589 tilted striae.

590 **Figure 6.** A) detail geologic map of Jebel Oust ([Castany et al., 1957](#)). B) Synsedimentary Barremian
591 normal faults with Lower hemisphere stereograms projection of back-tilted striae. C) Synsedimentary
592 Aptian normal fault with lower hemisphere stereograms projection of back-tilted striae. D) Detail
593 photo within the synsedimentary fault plane.

594 **Figure 7.** A) Geologic map of Jebel Rihane ([Arfaoui, 2007](#)). B) Synsedimentary Albian right-lateral fault
595 with lower hemisphere stereograms projection of back-tilted striae. C). Detail photo within the
596 synsedimentary fault plane.

597 **Figure 8.** A) Detailed geologic map of Jebel Kechtilou ([Haggui, 2012](#)). B) Synsedimentary Albian normal
598 fault with lower hemisphere stereograms projection of back-tilted striae. C) Reworked Aptian
599 limestones in the Albian sequences. D) Inherited normal faulting controlling the sedimentation,
600 inverted later by tertiary compressions.

601 **Figure 9.** A) Detailed geologic map of Lansarine belt ([Masrouhi et al., 2013](#)). B) Synsedimentary Aptian-
602 Albian normal fault with lower hemisphere stereograms projection of back-tilted striae in Oued
603 Tazegga. C) Synsedimentary Cenomanian normal fault with lower hemisphere stereograms projection
604 of back-tilted striae in Glib El Abiod.

605 **Figure 10.** Structural data from Slump folds with inferred paleoslope direction in Jebel Oust. A)
 606 Berriasian: Field photo with rose diagram of direction of poles to axial planes and lower hemisphere
 607 equal area projection of back-tilted slump axis (red box) and poles to axial planes (Blue triangles) with
 608 respectively MAM (red arrow) and APM (Blue Arrow) methods. B) Valanginian: same caption as A. C).
 609 Barremian: same caption as A.

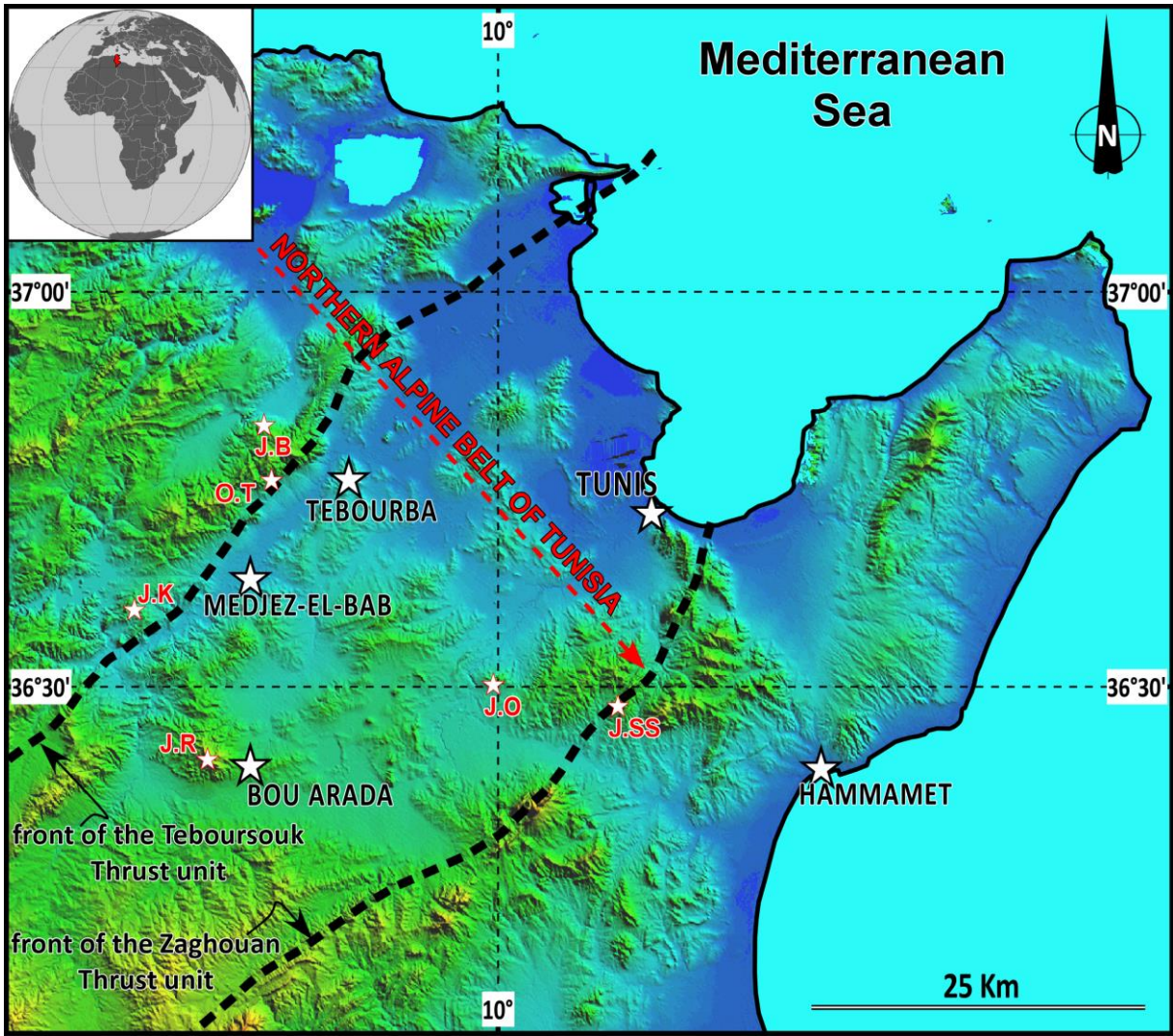
610 **Figure 11.** Structural data from Slump folds with inferred paleoslope direction in Jebel Sidi Salem–
 611 Messella and Jebel Rihane. A) Aptian–Albian (Jebel Sidi Salem): Field photo with rose diagram of
 612 direction of poles to axial planes and lower hemisphere, equal area projection of back-tilted slump axis
 613 (red box) and poles to axial planes (Blue triangles) with respectively MAM (red arrow) and APM (Blue
 614 Arrow) methods. B) Albian (Jebel Rihane): same caption as A. C). Aptian (Jebel Kechtilou): same caption
 615 as A.

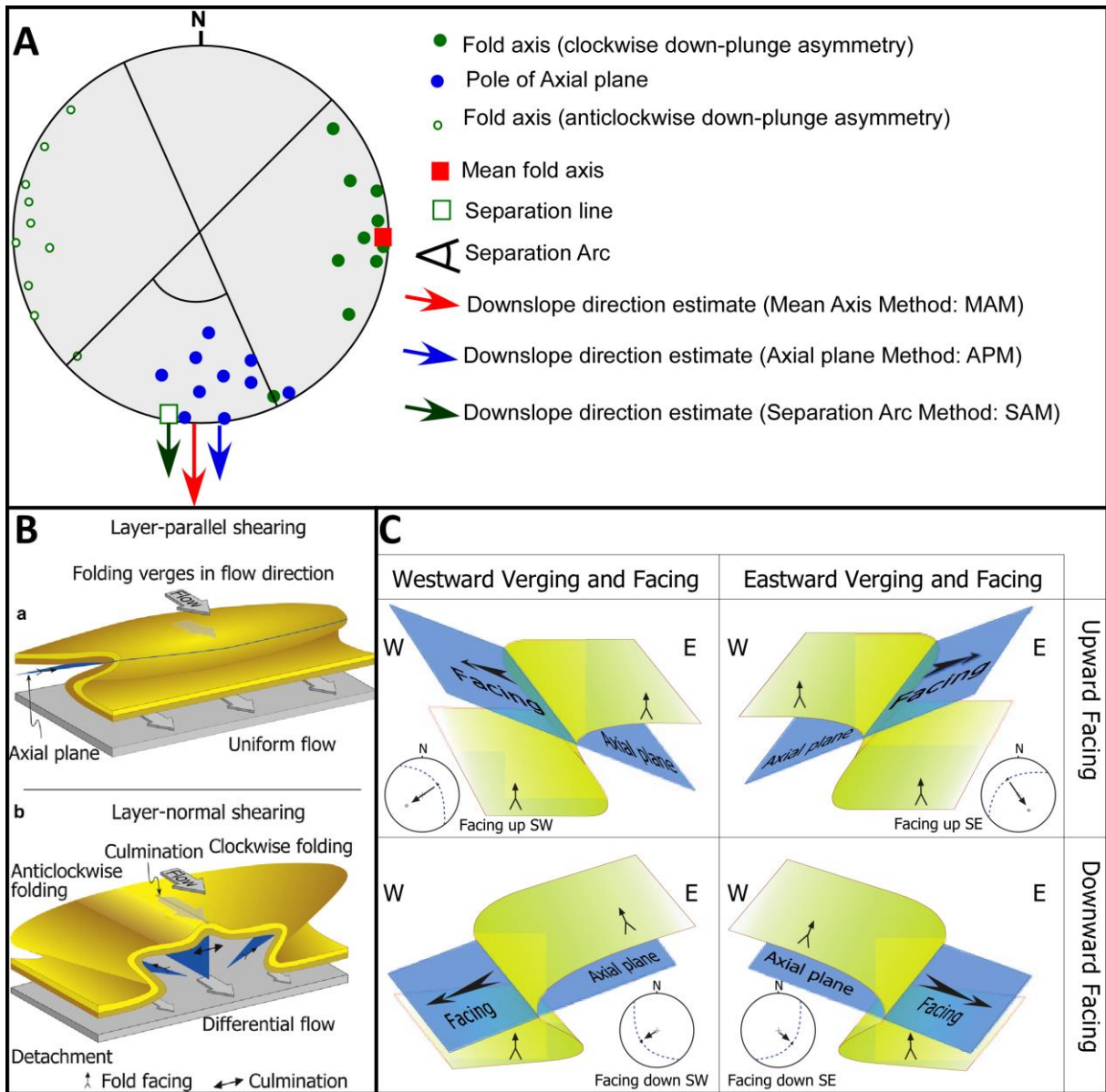
616 **Figure 12.** Structural data from Slump folds with inferred paleoslope direction in Lansarine belt: A)
 617 Aptian–Albian (Glib El Abiod): Field photo, with rose diagram of direction of poles to axial planes and
 618 lower hemisphere, equal area projection of back-tilted slump axis (red box) and poles to axial planes
 619 (Blue triangles) with respectively MAM (red arrow) and APM (Blue Arrow) methods. B) Aptian Albian
 620 (Oued Tazegga): same caption as A. C). Aptian (Jebel Boulahouadjeb): same caption as A.

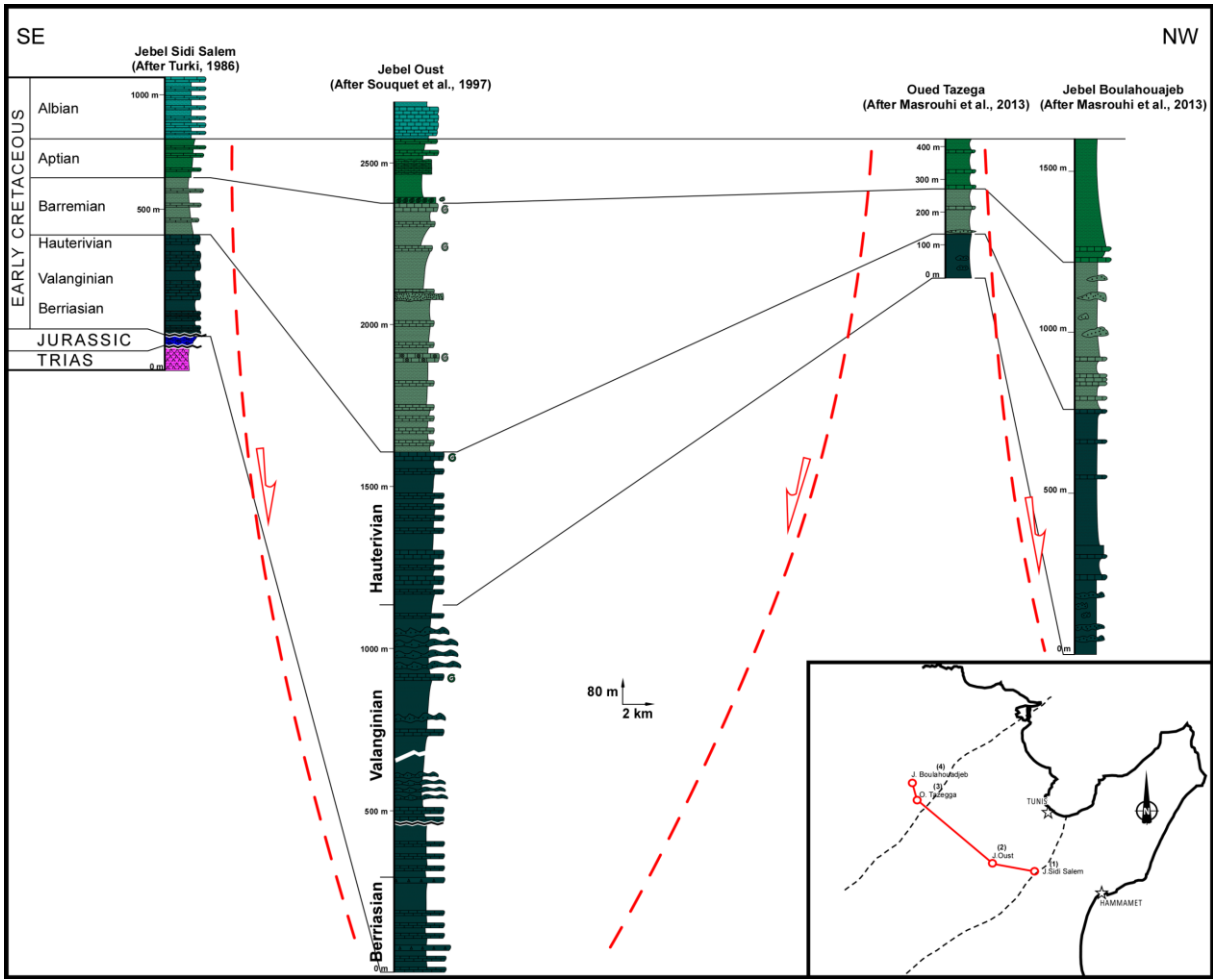
621 **Figure 13.** Structural data from slump folds in Upper Cretaceous series with inferred paleoslope
 622 direction near Jebel Sidi Salem-Messella, Jebel Kechtilou and Jebel Bechtat: A) Cenomanian
 623 (Jebel Sidi Salem): Field photo with rose diagram of direction of poles to axial planes and lower
 624 hemisphere equal area projection of back-tilted slump axis (red box) and poles to axial planes (Blue
 625 triangles) with respectively MAM (red arrow) and APM (Blue Arrow) methods. B) Cenomanian (Jebel
 626 Kechtilou): same caption as A. C). Cenomanian (Jebel Bechtat): same caption as A.

627 **Table 1:** Summary of paleoslope analysis results derived from MAM and APM methods.

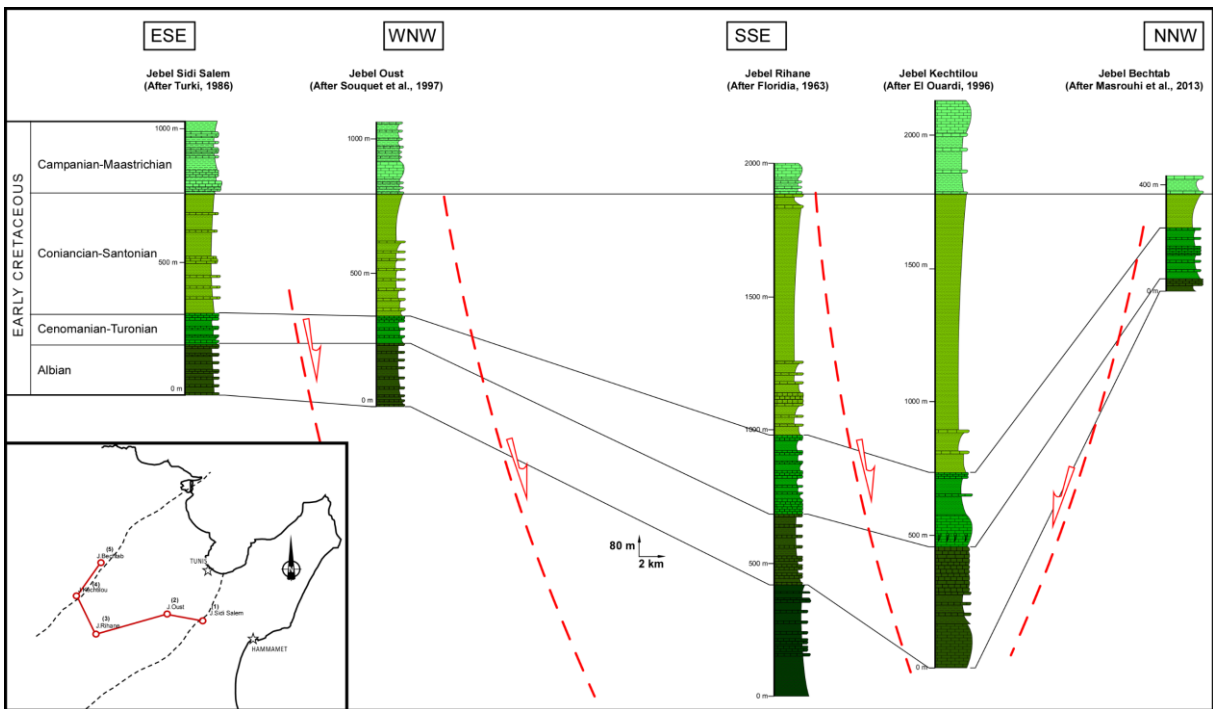
628 **Figure 14.** Simplified reconstruction of Northern Tunisia Early-to-Late Cretaceous basin. A)
 629 Paleogeographic reconstruction map of the western Tethyan domain during the Early Cretaceous
 630 (Valanginian, late Hauterivian, early Aptian) and Cenomanian–Turonian transition (modified After,
 631 [Soua, 2016](#)). Soua's reconstruction was made after [Barrier and Vrielynck \(2008\)](#) and [Tribovillard et al. \(2012\)](#). B) Paleogeographic reconstruction map of the Tunisian domain during the Early Cretaceous,
 633 with paleoslope assessment. C) Simplified reconstructed schematic cross-section (not to scale) of
 634 Northern Tunisia during Early–Late Cretaceous showing the structural configuration of the area with
 635 widespread soft-sediment deformation.



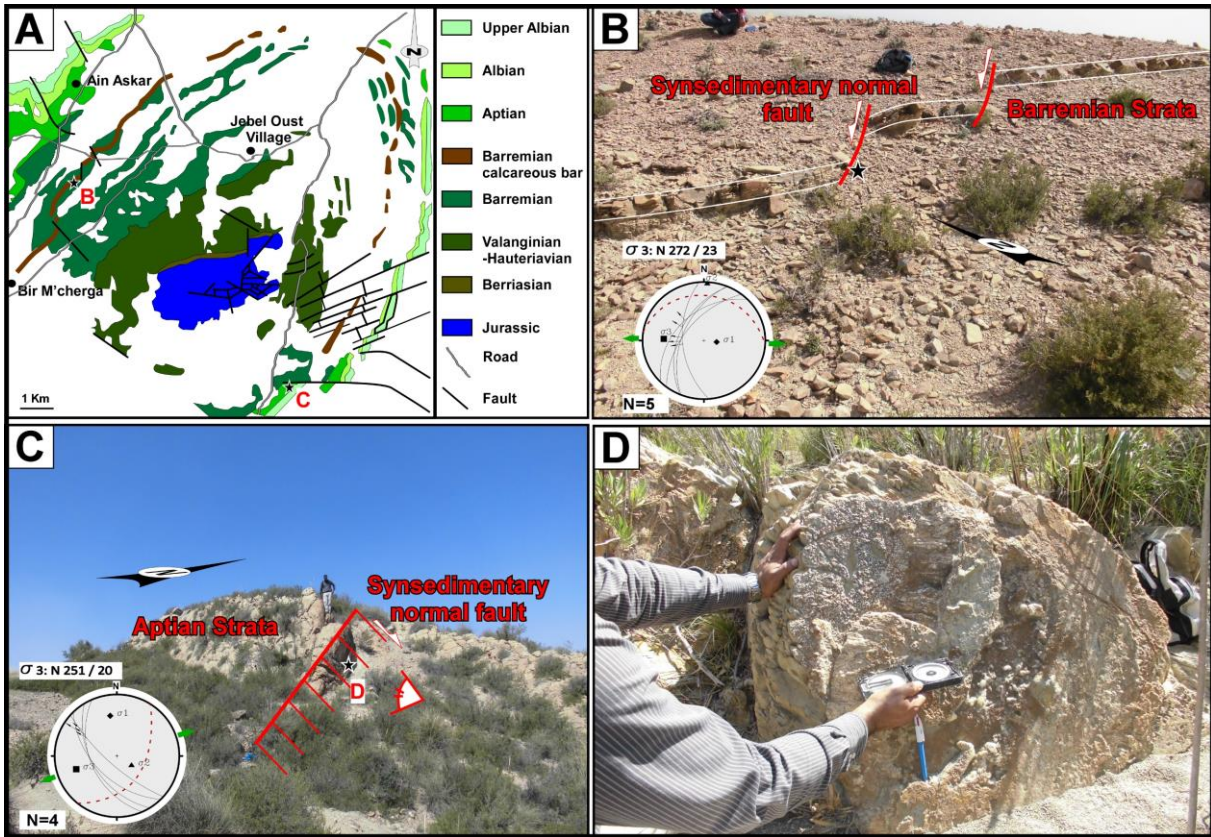




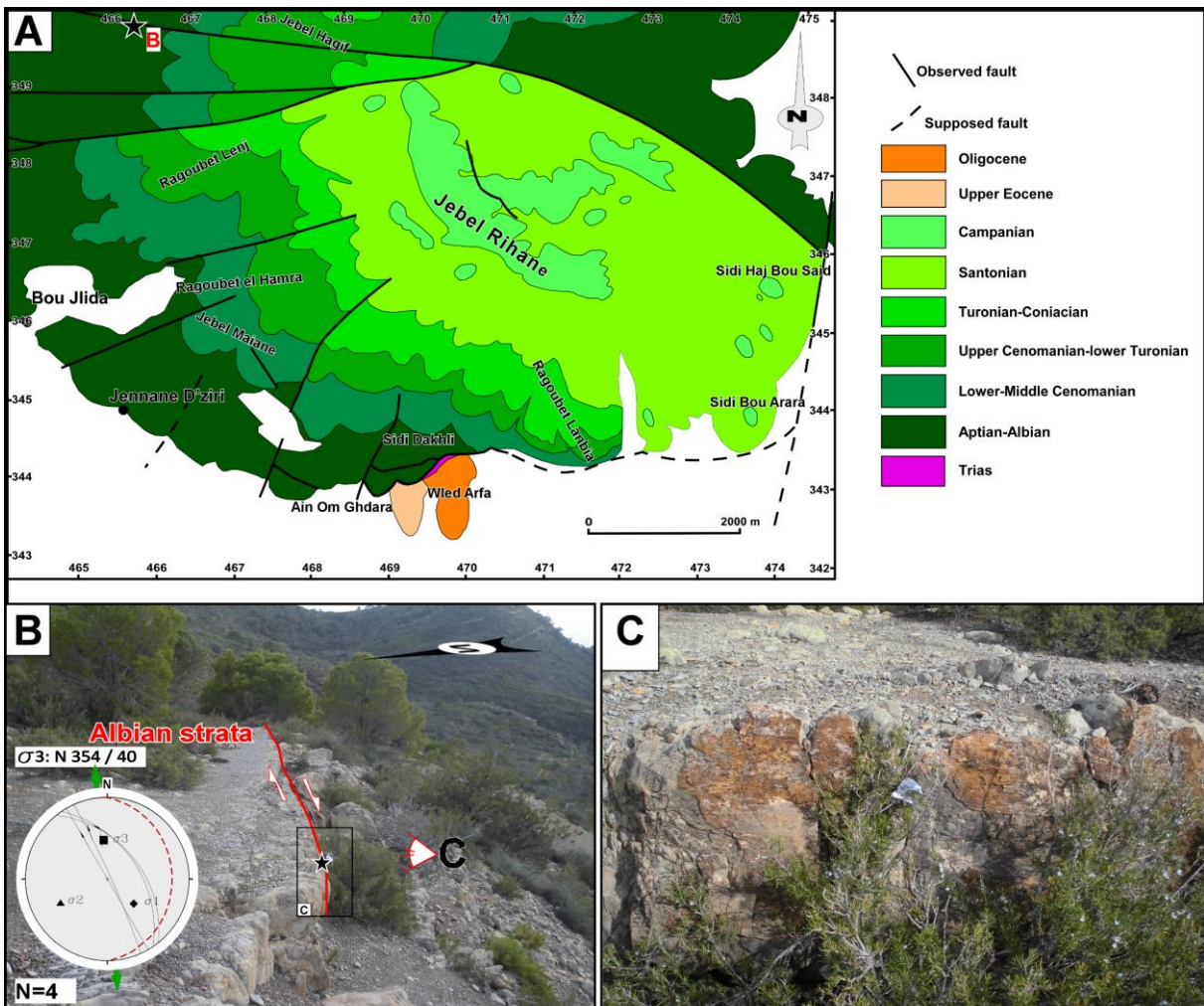
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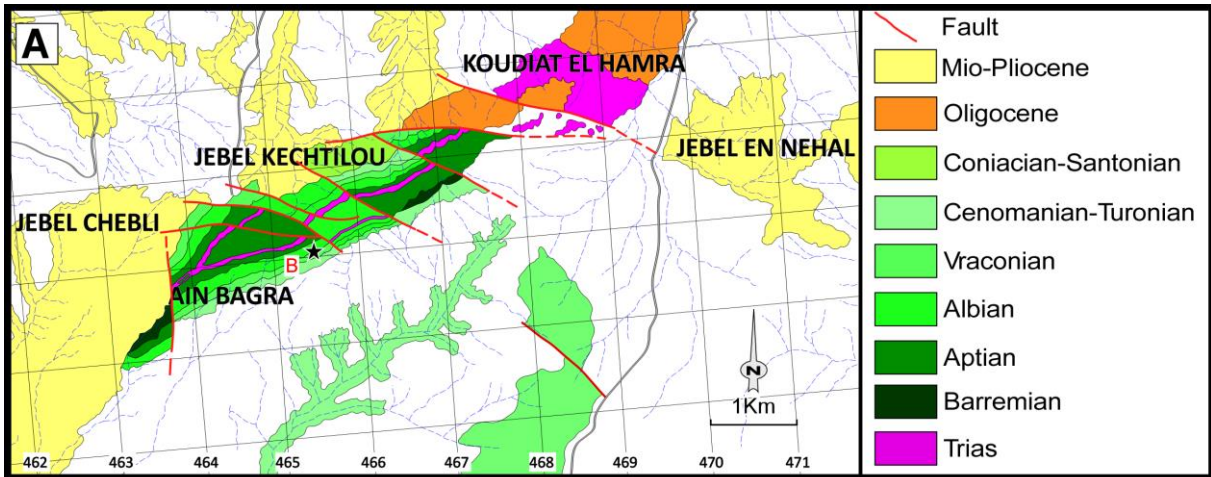
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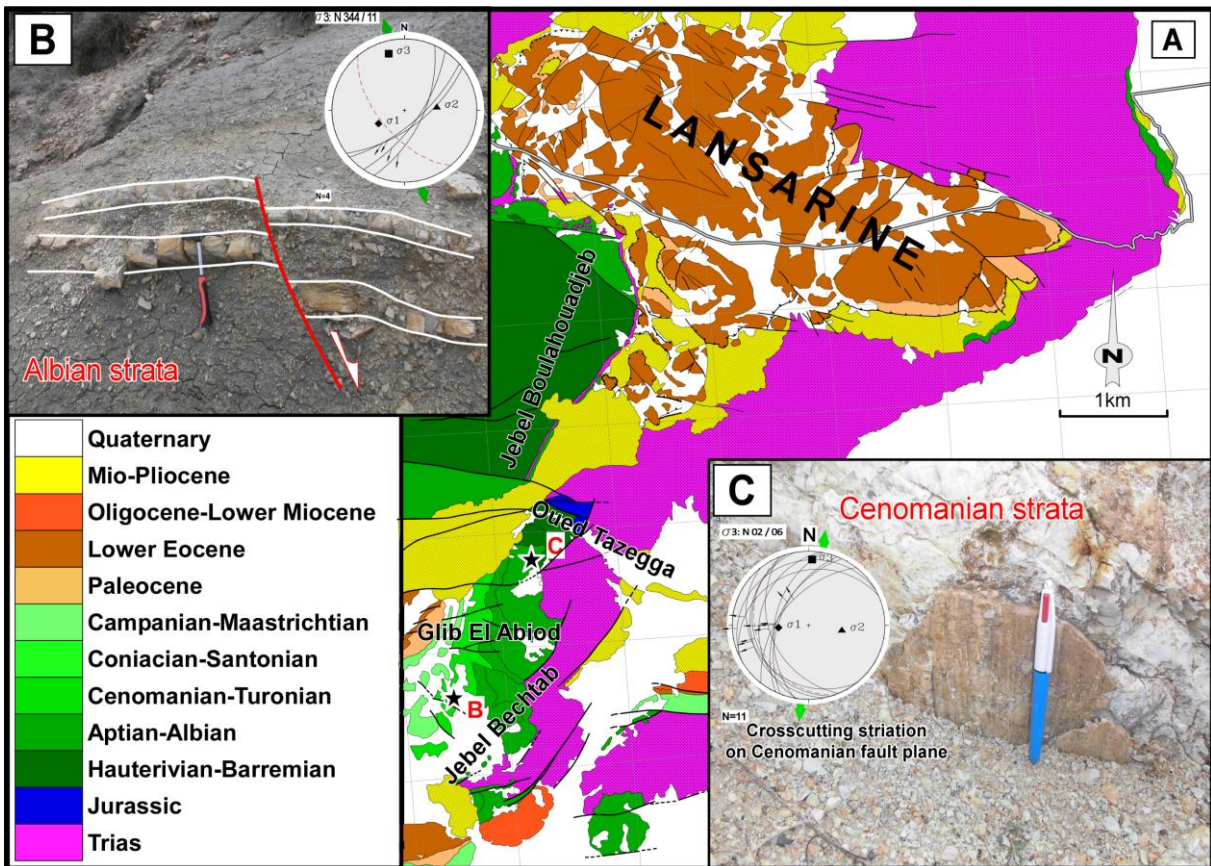
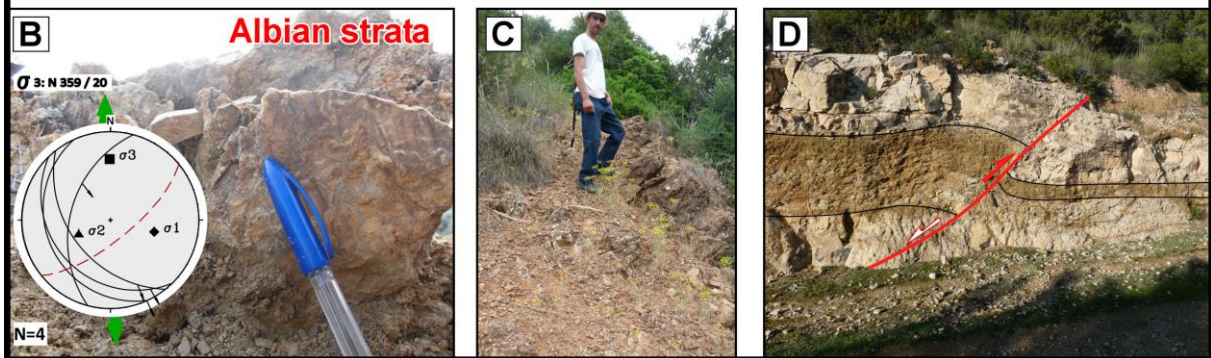
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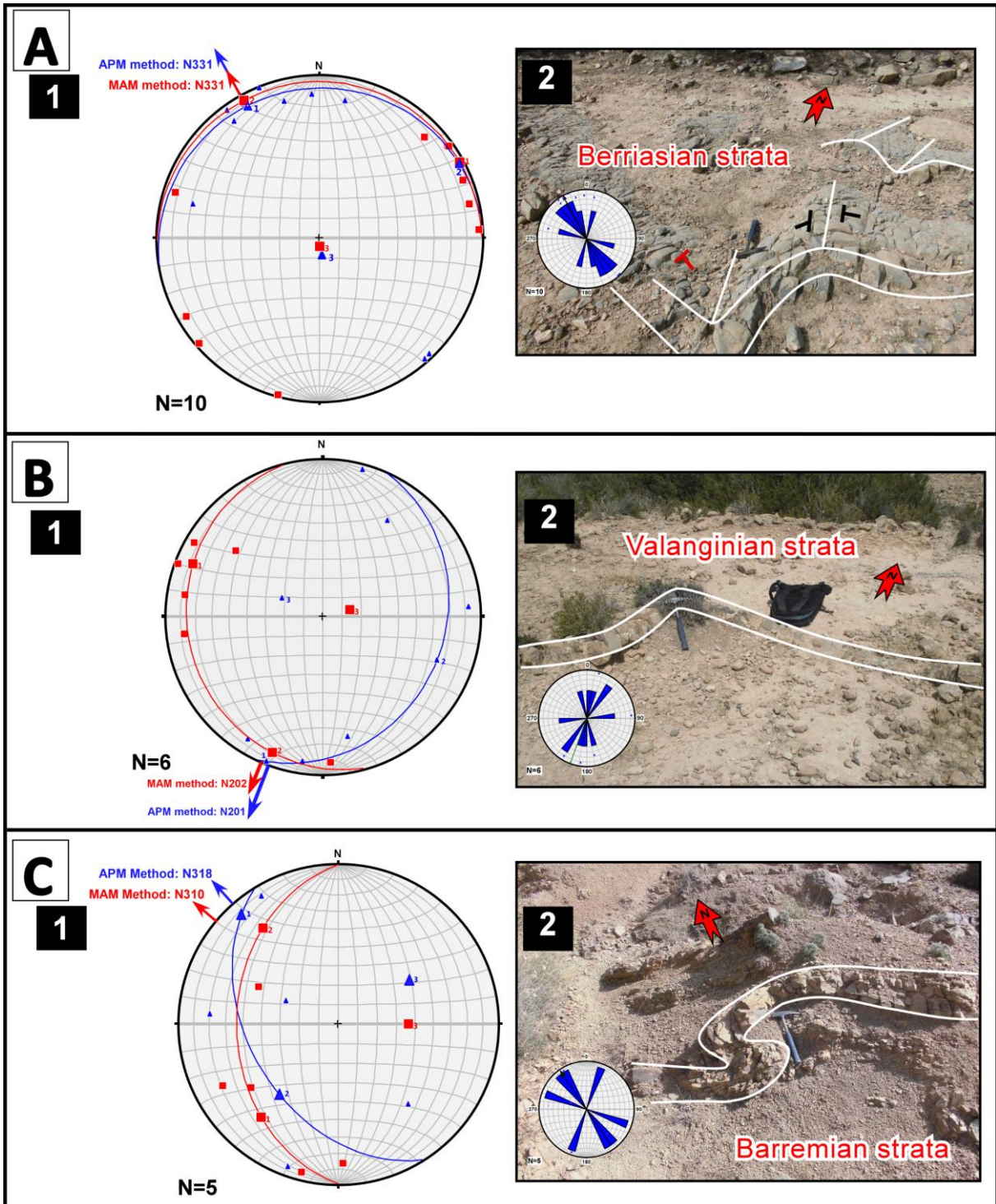
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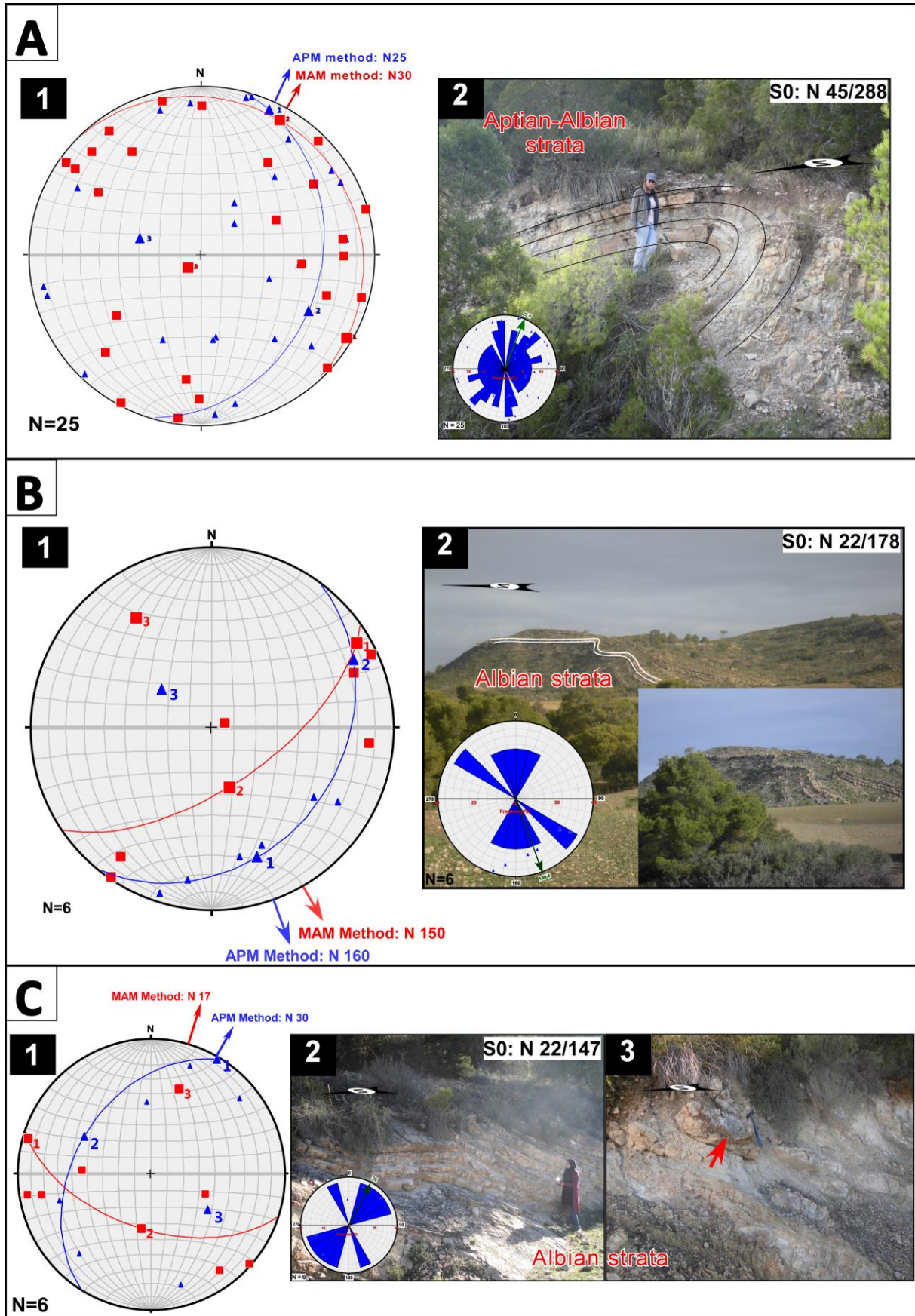


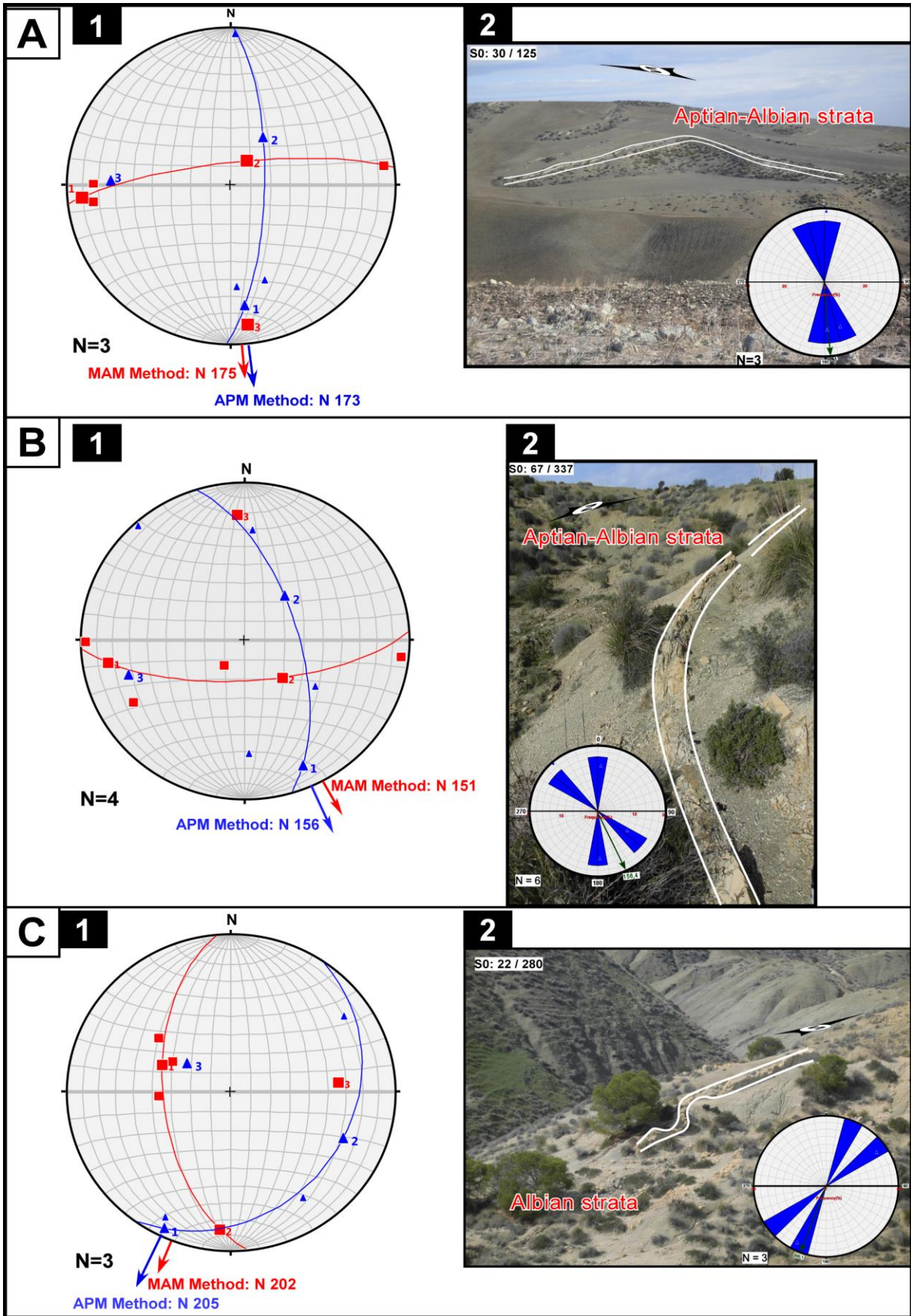
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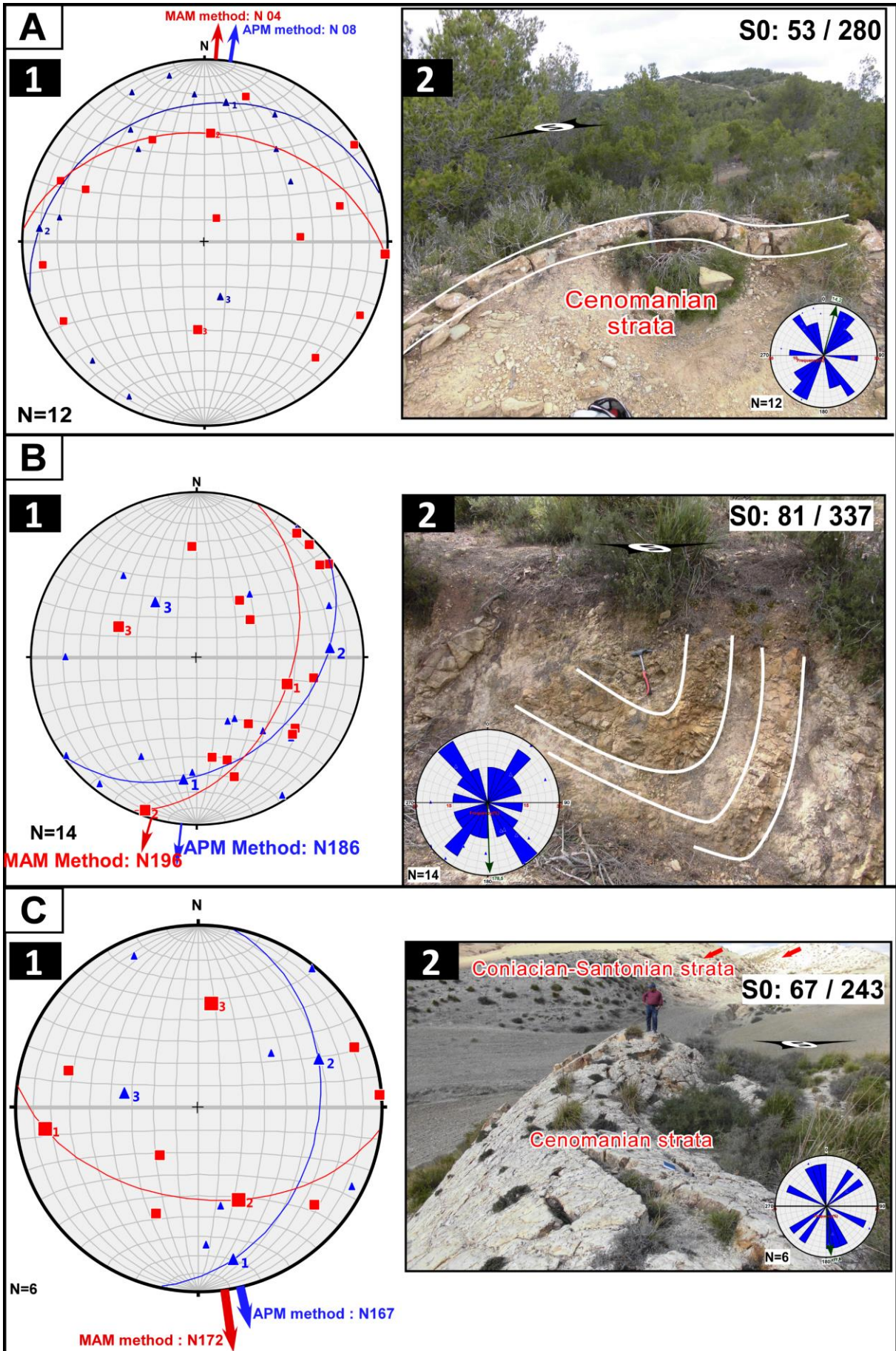


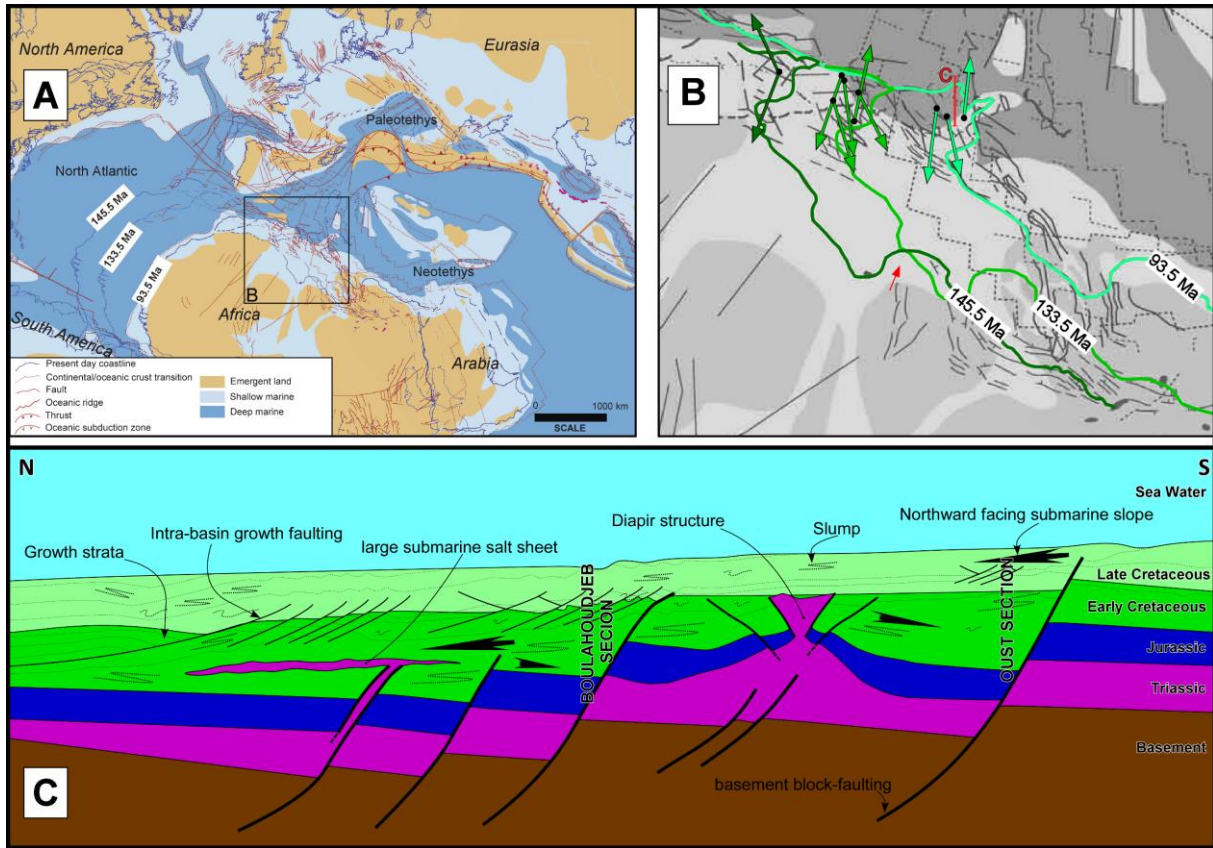
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