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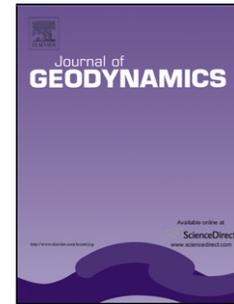


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Ikaria Island corresponds to a large-scale migmatite-cored MCC.

Thermal structure revealed by RSCM shows a drastic increase from top to bottom.

The MCC was exhumed by two detachments through the brittle-ductile transition.

Migmatites were dated to 15.7 ± 2 Ma by U-Th-Pb analysis on monazites.

A large-scale high temperature zone is proposed for the central part of the Aegean.

Accepted Manuscript

1 The Ikaria high-temperature Metamorphic Core Complex (Cyclades, Greece): Geometry,
2 kinematics and thermal structure.

3

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16

17

18 Abstract

19

20 This work attempted at clarifying the structure of Ikaria using primarily intensive
21 geological mapping combined with structural analysis and a geothermometry approach of
22 Raman spectrometry of carbonaceous material. Foliation over the whole island defines a
23 structural dome cored by high-grade to partially-molten rocks. Its exhumation was completed
24 by two top-to-the-N ductile extensional shear zones, operating in the ductile and then the
25 brittle fields, through a single extensional event coeval with progressive strain localization.

26 The thermal structure of the dome with regard to position of ductile shear zones was retrieved
27 using the Raman spectroscopy of carbonaceous material. Peak-metamorphic temperatures
28 range from 390 °C in the upper parts of the structure down to 625 °C in the core of the dome
29 in the vicinity of migmatites and S-type granite. Pioneer in situ U-Th-Pb analyses on monazite
30 performed on the leucosome parts of these rock yielded a 15.7 ± 0.2 Ma age. Ikaria Island
31 thus completes the series of Miocene migmatite-cored Metamorphic Core Complex in the
32 central part of the Aegean domain where a genuine high-temperature zone can be defined as
33 the central Aegean HT zone. There, the extreme stretching of the continental crust is
34 associated with dominantly top-to-the-N kinematics.

35

36 Keywords: Structural analysis; RSCM geothermometry; U-Th-Pb geochronology;
37 Metamorphic Core Complex; Ikaria; North Cycladic Detachment System.

38

39

40 1. Introduction

41

42 In the Mediterranean realm, the retreat of oceanic slabs triggered the initiation of back-
43 arc extension (i.e large-scale extension in the upper plate of a subduction zone), leading to the
44 collapse of previously thickened continental lithosphere (Le Pichon and Angelier, 1979;
45 Malinverno and Ryan, 1986; Dewey, 1988; Platt and Vissers, 1989; Royden, 1993; Jolivet
46 and Faccenna, 2000; Rosenbaum et al., 2002; Faccenna et al., 2004; Jolivet et al., 2008). This
47 post-orogenic evolution (i.e. crustal thinning by extension-related normal faulting after an
48 episode of crustal thickening) resulted in the formation of series of extensional domains or
49 wide-rift systems (Lister et al., 1984; Buick, 1991; Corti et al., 2003) such as the Alboran Sea,
50 the Tyrrhenian Sea, the Pannonian Basin and the Aegean Sea. Lateral evolution from the

51 central parts of the extensional domains to the bounding non-collapsed orogenic segments
52 implies drastic lateral gradients of finite extension and suggests highly non-cylindrical
53 structures. In the Aegean Sea, along with the drastic decrease in topography and crustal
54 thickness, the main orogenic structures of the Hellenic belt are increasingly reworked by
55 extension from continental Greece to Naxos, in the center of the Aegean Sea. Extensional
56 structures related to back-arc crustal stretching evolve from essentially brittle steep and
57 shallow-dipping normal faults in continental Greece to shallow-dipping ductile shear zones in
58 the center of the Aegean domain (Jolivet and Patriat, 1999; Jolivet et al., 2010). Furthermore,
59 a straightforward correlation between the degree of non-coaxiality of the back-arc domain and
60 the amount of stretching of the continental crust is observed at the scale of the entire Aegean
61 domain (e.g. Augier et al., 2015). Marginal areas that connect to non-collapsed orogenic
62 segments display symmetrically arranged detachment systems. In the western parts of the
63 Aegean domain, both the West Cycladic Detachment System (WCDS) and the North Cycladic
64 Detachment System (NCDS) exhume a horst-shaped domain, where orogenic features are still
65 nicely preserved, depicting bivergent extension (e.g. Jolivet et al., 2010; Grasemann et al.,
66 2012). Conversely, in the center of the Aegean domain, deformation remains highly
67 asymmetric from Mykonos in the north, all the way to Sikinos in the south (e.g. Gautier et al.,
68 1993; Kumerics et al., 2005; Denèle et al., 2011; Augier et al., 2015). There, orogenic features
69 are particularly overprinted or even locally erased by the combined effects of intense top-to-
70 the-N shearing and partial-melting. Migmatite-cored Metamorphic Core Complexes (MCCs),
71 associated with Miocene intrusions, roofed by major top-to-the-N crustal-scale detachments,
72 are described on Naxos, in the center (e.g. Lister et al., 1984; Urai et al., 1990; Buick, 1991;
73 Gautier and Brun, 1994; Jolivet et al., 2004a; Vanderhaeghe, 2004) and Mykonos in the north
74 of the Aegean domain (Lecomte et al., 2010; Denèle et al., 2011). Concentrating a large part
75 of the total amount of stretching, recognition of MCCs therefore appears of prime importance.

76 Similarly, asymmetry of crustal thinning at the regional-scale is another key-question for
77 understanding back-arc extension dynamics. However, the current understanding of back-arc
78 dynamics in the Aegean domain is hindered by the severe lack of knowledge for the bulk of
79 its eastern part.

80 One of the largest Aegean islands, displaying the largest intrusion of the Aegean Sea,
81 located between the northern Cyclades and western Turkey, Ikaria Island has been the focus
82 of several recent studies. However, the first order structural architecture of this island remains
83 conflicting and the existing geological maps of Ikaria present marked discrepancies. Besides,
84 the current knowledge of the metamorphic record and particularly the thermal structure of
85 Ikaria remain fragmentary. These problems were reconsidered after an extensive field survey,
86 including primarily new geological mapping and structural analysis. The position and
87 importance of the various tectonic contacts as well as the bulk thermal architecture of the
88 island were further constrained using the geothermometry approach of Raman Spectrometry
89 of Carbonaceous Material (RSCM). Following their recent discovery, the migmatites
90 described in the core of the structure were dated by U-Th-Pb LA-ICPMS analyses on
91 monazite.

92

93

94 2. Geological setting

95

96 2.1. Geodynamic context

97

98 The Aegean domain (Fig. 1) corresponds to the collapsed segment of the Hellenides-
99 Taurides belt, developed as the result of the convergence between Apulian and European
100 plates, in the eastern Mediterranean, since the Late Cretaceous (e.g. Aubouin and Dercourt,

101 1965; Brunn et al., 1976; Bonneau and Kienast, 1982; Bonneau, 1984; van Hinsbergen et al.,
102 2005; Jolivet and Brun, 2010, Ring et al., 2010). During this period, a south-verging crustal-
103 scale orogenic wedge was formed by subduction and accretion of several Apulia-derived
104 continental blocks separated by oceanic basins. Post-orogenic extension of the Hellenic
105 thickened crust started in the Early Oligocene (Jolivet and Faccenna 2000; Jolivet et al., 2008)
106 or the Early Miocene (Ring et al., 2010) by a combination of gravitational collapse and back-
107 arc extension during the southward retreat of the African slab (e.g. Jolivet and Faccenna,
108 2000; Jolivet and Brun, 2010). Intense crustal stretching of the upper plate leads to the
109 formation of a series of MCCs (Lister et al., 1984; Avigad and Garfunkel, 1989; 1991;
110 Gautier and Brun, 1994; Jolivet et al., 2004a). Despite the importance of the post-orogenic
111 overprint on the orogenic architecture, the original vertical superposition of tectonic units in
112 the nappe stack is often preserved at the scale of the whole Aegean domain (e.g. Bonneau,
113 1984; van Hinsbergen et al., 2005). Three main tectonometamorphic units are classically
114 recognized (e.g. Bonneau, 1984; Ring et al., 2010) (Fig. 1):

115 1) The Upper Cycladic unit corresponds to a lateral equivalent of the Pelagonian nappe
116 (e.g. Bonneau, 1984; Jolivet et al., 2004a) recognized in continental Greece. This unit
117 generally crops out as isolated klippe or rafts in the Cyclades, essentially made of ophiolitic
118 material such as in Andros, Tinos, Mykonos, Kea, Kythnos, Serifos and Samos (e.g. Ring et
119 al., 1999; Jolivet et al., 2010; Grasemann et al., 2012). Rocks preserve a Cretaceous *HT-LP*
120 metamorphic imprint but escaped both the Eocene *HP-LT* and the Oligocene-Miocene *HT-LP*
121 tectonometamorphic events (Katzir et al., 1996). Syn-tectonic detrital shallow-marine and
122 continental sediments locally form the uppermost unit on Mykonos, Paros, Naxos and Ikaria
123 (e.g. Angelier, 1976; Photiades, 2002a; Sánchez-Gómez et al., 2002; Kuhlemann et al., 2004;
124 Lecomte et al., 2010). Conglomerates mostly contain pebbles derived from the Upper

125 Cycladic nappe and reworked magmatic rocks as young as 10 Ma (e.g. Sánchez-Gómez et al.,
126 2002).

127 2) The Cycladic Blueschists unit crops out as a composite unit including locally a
128 significant component of metabasic rocks interleaved with metapelites and marbles, all
129 equilibrated in blueschist-facies conditions (e.g. Blake et al., 1981; Bonneau, 1984; Avigad
130 and Garfunkel, 1991; Keiter et al., 2004). This unit experienced a complex alpine
131 tectonometamorphic evolution, with an early burial in *HP-LT* conditions reaching blueschist
132 to eclogite-facies conditions during the Eocene, followed by a greenschist overprint of
133 variable intensity during the Oligocene and the Miocene (Altherr et al., 1979, 1982; Wijbrans
134 et al., 1990; Parra et al., 2002; Duchêne et al., 2006; Augier et al., 2015). On Ikaria, despite
135 the lack of reliable *HP* index minerals, the Messaria unit (Kumerics et al., 2005) was
136 correlated with the Ampelos nappe recognized on Samos that experienced metamorphic
137 conditions of the order of 15 kbar and 500 °C (Will et al., 1998).

138 3) The lower units crop out in tectonic windows. The Cycladic Basement unit crops
139 out in the central and southern Cyclades (i.e. on Naxos, Paros, Sikinos and Ios) (e.g.
140 Andriessen et al., 1987). It is composed of Variscan granitoids mantled by micaschists that
141 retain either metamorphic relics of amphibolite-facies assemblages or inherited radiometric
142 ages suggesting a complex prealpine history (e.g. Henjes-Kunst and Kreuzer, 1982;
143 Andriessen et al., 1987; Keay and Lister, 2002). It is sometimes covered by Mesozoic marbles
144 that may represent a *HP* equivalent of the Gavrovo unit cropping out in continental Greece
145 (Jolivet et al., 2004b). Just as the Cycladic Blueschists unit, the Cycladic Basement unit
146 shows a complex alpine tectonometamorphic evolution, with an initial subduction-related
147 burial in *HP-LT* conditions during the Eocene whose trace has been obscured by an
148 Oligocene-Miocene local overprint (van der Maar et al., 1981; Vandenberg and Lister, 1996;
149 Baldwin and Lister, 1998; Augier et al., 2015). On Naxos and Paros, it experienced a partial-

150 melting stage in amphibolite to granulite-facies conditions (i.e. Jansen and Schuiling, 1976;
151 Buick and Holland, 1989; Vanderhaeghe, 2004; Duchêne et al., 2006).

152 Rocks of the Cycladic Blueschists and the lower units were exhumed during two
153 distinctive stages in two contrasted geodynamic settings. The first stage occurred in the
154 Hellenic subduction context, during the Eocene, with burial and synorogenic exhumation (by
155 shortening-related normal faulting) of blueschist to eclogite-facies assemblages in an
156 extrusion wedge structure (e.g. Altherr et al., 1979; Wijbrans et al., 1990; Trotet et al., 2001a;
157 2001b; Groppo et al., 2009; Ring et al., 2007; Jolivet and Brun, 2010). The second stage
158 occurred in the Oligocene-Miocene with the post-orogenic exhumation in the back-arc
159 domain of the Cycladic Blueschists and the lower units as metamorphic domes or migmatite-
160 cored MCCs (Lister et al., 1984) below a series of detachments (e.g. Avigad and Garfunkel,
161 1989; Buick and Holland, 1989; Buick, 1991; Faure et al., 1991; Lee and Lister, 1992;
162 Gautier et al., 1993; Gautier and Brun, 1994; Jolivet and Patriat, 1999; Keay et al., 2001;
163 Vanderhaeghe, 2004; Kumerics et al., 2005; Mehl et al., 2005; 2007; Denèle et al., 2011;
164 Augier et al., 2015). In the northern Cyclades, these detachments were recently grouped in a
165 single large-scale top-to-the-N structure running over 130 km to form the NCDS (Fig. 1;
166 Jolivet et al., 2010). A similar set of detachments with opposed, top-to-the-S or SW
167 kinematics, was identified in the western Cyclades (Grasemann and Petrakakis, 2007; Iglseider
168 et al., 2009, 2011; Tschegg and Grasemann, 2009; Brichau et al., 2010) and recently
169 mechanically linked to form the WCDS (Grasemann et al., 2012). The Naxos-Paros
170 Detachment (NDP) completes those series of detachment systems in the center of the Aegean
171 domain (Buick and Holland, 1989; Buick, 1991; Gautier and Brun, 1994; Kruckenberg et al.,
172 2011). Another major shear zone over which top-to-the-N kinematics rework top-to-the-S
173 shear sense occurs in the southern Cyclades on the islands of Sikinos and Ios (e.g.
174 Vandenberg and Lister, 1996; Forster and Lister, 2009; Huet et al., 2009; Thomson et al.,

175 2009; Augier et al., 2015). Extension on this shear zone starts to exhume rocks near the Early
176 Miocene (Thomson et al., 2009). Here, top-to-the-S kinematics was first considered as
177 extensional (Vandenberg and Lister, 1996; Forster and Lister, 2009; Thomson et al., 2009)
178 and later correlated with the WCDS (Ring et al., 2011). Another study reinterpreted this shear
179 zone as a top-to-the-S thrust reworked by top-to-the-N extension (the South Cycladic Thrust,
180 SCT) (Huet et al., 2009). The same interpretation is made on Sikinos where the SCT crops out
181 (Augier et al., 2015). Late exhumation stages along extensional systems were accompanied by
182 the emplacement of syn-tectonic Miocene I and S-type granites (i.e. Tinos, Mykonos, Naxos,
183 Serifos and Ikaria; Fig. 1) (Faure et al., 1991; Lee and Lister, 1992; Altherr and Siebel, 2002;
184 Grasemann and Petrakakis, 2007; Ring, 2007; Iglseider et al., 2009; Bolhar et al., 2010;
185 Laurent et al., 2015). The activity of detachments is also constrained by syn-tectonic
186 deposition of sediments over the Upper Cycladic unit (e.g. Sanchez-Gomez et al., 2002;
187 Kuhlemann et al., 2004; Lecomte et al., 2010; Menant et al., 2013).

188 Although part of these features of the Cycladic geology are found further east within
189 the Menderes massif (Fig. 1), type, distribution and timing of metamorphism appear less
190 clear. The Cycladic Blueschists unit and the Lycian Nappe that experienced *HP-LT*
191 metamorphic conditions rest on top of the structure to the north and to the south of the massif
192 (Oberhänsli et al., 1998; Rimmelé et al., 2003a; 2003b; Pourteau et al., 2010). *HP-LT*
193 metamorphic conditions are dated from Late Cretaceous to Eocene (Oberhänsli et al., 1998;
194 Pourteau et al., 2013). Crustal stretching responsible for the final exhumation of the
195 metamorphic part of the Menderes massif under shallow-dipping shear zones (e.g. Rimmelé et
196 al., 2003b; Bozkurt, 2007; van Hinsbergen, 2010; Bozkurt et al., 2011) is very similar to the
197 Cyclades one. Detachments juxtapose Neogene syn-tectonic sediments on metamorphic rocks
198 that are strongly retrograded in the greenschist facies during the Miocene (Hetzl et al.,

199 1995a; Lips et al., 2001). Here again, crustal thinning is accompanied by the emplacement of
200 syn-tectonic granites (Ring and Collins, 2005; Glodny and Hetzel, 2007).

201

202

203 2.2. Geology of Ikaria

204

205 Ikaria is a 40 km-long island located in the eastern Aegean Sea between Mykonos and
206 Samos. A recent map coverage was performed by the Greek Institute of Geology and Mineral
207 Exploration (G-IGME) (Photiades, 2002b), complemented by independent geological maps
208 presenting highly conflicting interpretations (Papanikolaou, 1978; Kumerics et al., 2005;
209 Ring, 2007; Bolhar et al., 2010; Kokkalas and Aydin, 2013).

210 At first glance, the geology of Ikaria consists in an equal distribution of a metamorphic
211 domain to the east and a large-scale magmatic complex to the west, bounded by sedimentary
212 rocks (Fig. 2). The metamorphic part consists in a 1500 m-thick tectonometamorphic
213 succession made of metasediments including metapelites, metaquartzites, marbles and minor
214 metabasites occurrences passing upward to finely alternating metapelites and marbles
215 (Photiades, 2002a; 2002b; Kumerics et al., 2005). The metamorphic grade decreases upward
216 from widespread amphibolite-facies associations (i.e. staurolite-garnet-biotite in metapelites
217 and hornblende-plagioclase in metabasites) to greenschists-facies associations in the
218 uppermost parts of the succession (e.g. Altherr et al., 1982; Kumerics et al., 2005; Martin et
219 al., 2011). Peak-metamorphic conditions, as retrieved from pseudo-section approaches,
220 yielded 6-8 kbar for 600-650 °C conditions for the basal parts of the tectonometamorphic
221 succession (Kumerics et al., 2005; Martin et al., 2011), fringing partial-melting conditions
222 assuming water saturation conditions (Weinberg and Hasalova, 2015). While similar
223 quantitative P-T estimates are currently lacking for the upper parts of the succession, a

224 localized change in the metamorphic grade has however been proposed within the upper parts
225 (Papanikolaou, 1978; Altherr et al., 1982; Kumerics et al., 2005). Besides, it is noteworthy
226 that traces of an initial *HP* imprint is currently lacking on Ikaria, at variance with neighboring
227 islands (Altherr et al., 1982; Photiades, 2002a; Kumerics et al., 2005; Ring, 2007).

228 Three main magmatic intrusions were recognized on Ikaria (Papanikolaou, 1978;
229 Photiades, 2002b; Ring, 2007); two small-scale, less than 10 km² S-type two-mica
230 leucogranite intrusions (i.e. the Xylosyrtis and the Karkinagrion intrusions) and a large-scale
231 I-type intrusion (i.e. the Raches intrusion). Along with a related pervasive pegmatite dyke
232 array (Photiades, 2002b; Hezel et al., 2011), the Xylosyrtis pluton displays a clear intrusive
233 character within the metamorphic series. Conversely, the nature of the Raches granite contact
234 remains highly controversial and was mapped so far as intrusive (Papanikolaou, 1978;
235 Laurent et al., 2015), as a detachment (Kumerics et al., 2005) or as a thrust (Photiades, 2002a;
236 2002b; Kokkalas and Aydin, 2013). Emplacement of the main intrusions occurred in a narrow
237 15-13 Ma age-range (Bolhar et al., 2010).

238 Metamorphic and intrusive rocks experienced an intense top-to-the-N shearing in both
239 ductile and then brittle regimes (Kumerics et al., 2005; Ring, 2007). The Raches intrusion in
240 the west thus shows a top-to-the-N strain gradient toward the upper structural levels, from
241 proto-mylonites to ultra-mylonites, and finally cataclasites (Laurent et al., 2015). Prior to this
242 study, two main tectonic contacts, along which deformation is concentrated, were
243 distinguished: the Messaria and Fanari detachments (Kumerics et al., 2005). According to
244 Kumerics et al. (2005), the Messaria detachment corresponds to a mylonite zone later partly
245 overprinted by cataclastic deformation. At variance, the Fanari detachment is currently
246 regarded as a purely brittle contact roofing the metamorphic rocks and bounding the
247 sediments of the upper unit (Kumerics et al., 2005; Ring, 2007). This unit, regionally known
248 as the Fanari unit, consists in sandstones, siltites and conglomerates containing clasts of red

249 cherts and ophiolitic rocks of Early Cretaceous age and large-scale olistoliths of Triassic
250 recrystallized limestones (Papanikolaou, 1978) reminding typically rocks of Pelagonian
251 affinity (Pe-Piper and Photiades, 2006). Ages of those sediments spread from Oligocene to
252 Pliocene (Photiades, 2002a; 2002b). In the center of the island, the recrystallized limestone of
253 Kefala, inferred as Triassic, is sometimes described as a klippe of the upper unit of Ikaria
254 (Papanikolaou, 1978; Photiades, 2002a; 2002b; Pe-piper and Photiades, 2006). The main
255 argument for the presence of the Pelagonian unit relies on the description of a diorite intrusive
256 body that yielded Cretaceous K/Ar ages on hornblende (Altherr et al., 1994).

257 Precise age-constraints on peak-metamorphic conditions and the accurate timing for
258 the onset of extensional motions along main shear zones are currently lacking on Ikaria. First
259 record of cooling, below 550 °C (Villa, 1998), is scattered between 25 and 17 Ma (K/Ar on
260 hornblende; Altherr et al., 1982). Late exhumation stages are, in turn, well constrained after
261 15 Ma by varied thermochronological tools (Altherr et al., 1982; Kumerics et al., 2005). The
262 K/Ar and Ar/Ar ages on both fabric-forming white mica and biotite yield numerous 12 to 9
263 Ma ages. In metamorphic rocks, these ages are interpreted as both cooling or deformation
264 ages during exhumation while they are considered as cooling ages for granites (Altherr et al.,
265 1982; Kumerics et al., 2005). Fission-track (FT) analyses yielded 10.3 ± 0.3 to 7.1 ± 0.3 Ma
266 ages for zircon and 8.4 ± 0.8 to 5.9 ± 0.8 Ma ages for apatite (Kumerics et al., 2005). Final
267 cooling stages were constrained by U-Th/He analyses on apatite that yielded symmetrically
268 arranged 6 Ma ages for both flanks and a 3 Ma age for the core of the structure (Kumerics et
269 al., 2005). Cooling rates therefore exceeded 100 °C/Ma from the ductile to brittle transition
270 (i.e. 300-400 °C, Stöckhert et al., 1999; Imber et al., 2001) and temperatures as low as 70 °C
271 between 10 to 6 Ma.

272

273

274 3. A new geological map of Ikaria

275

276 An extensive field survey was carried out on Ikaria including new field mapping in
277 order to complement the existing G-IGME geological map (Photiades, 2002b). The result is
278 presented in Fig. 2; readers are referred to existing maps showing very conflicting
279 interpretations to appreciate the modifications (Papanikolaou, 1978; Photiades, 2002b;
280 Kumerics et al., 2005; Ring, 2007; Bolhar et al., 2010; Kokkalas and Aydin, 2013).

281 For clarity, the lithostratigraphic subdivisions of Photiades (2002b) were kept as much
282 as possible. Three main lithologies, including marbles, micaschists (including calcschists and
283 minor metaquartzites and metabasites occurrences) and granitic rocks derived from three main
284 intrusive bodies were mapped (Fig. 2). Definition of the main tectonic units changed as the
285 result of the modification and the reinterpretation of their boundaries, particularly the tectonic
286 and intrusive contacts. Three main tectonic units, from bottom to top: the Ikaria, Agios-
287 Kirykos and Fanari units, limited by two major shear zones, are now distinguished. The main
288 tectonic features having a map-scale expression were also reported in detail. These structures
289 consist primarily in the Agios-Kirykos and the Fanari ductile shear zones developed as ramp-
290 flat extensional structures either at shallow-angle to the compositional layering as décollement
291 zones or cutting down-section. The Fanari shear zone even presents evidence for subsequent
292 displacements in the brittle field superimposed on ductile features. On the map, the Fanari
293 detachment clearly cuts across the whole Agios-Kirykos unit and the Agios-Kirykos shear
294 zone. Other shear zones or brittle contacts put forward on previous maps were abandoned.
295 The basal contact below Kefala marble (e.g. Papanikolaou, 1978; Altherr et al., 1994;
296 Photiades, 2002a; 2002b; Pe-piper and Photiades, 2006) is considered less important than in
297 previous studies since no metamorphic gap is backed up by published data. Moreover,
298 detailed mapping of this zone indicate that the diorite is intrusive in Ikaria unit. However, a

299 panorama on the Kefala marble seen suggests the presence of a tectonic contact at its base
300 (Papanikolaou, 1978). Although probably minor, this contact might be a late brittle expression
301 of the system of detachments. Similarly, the nature of the eastern contact of the Raches
302 granite considered either as the lateral equivalent of the Messaria detachment (Kumerics et al.,
303 2005; Ring, 2007) or a major thrust contact (Photiades, 2002a; 2002b; Kokkalas and Aydin,
304 2013) was re-evaluated. Field work in the vicinity of the contact of the Raches granite
305 unambiguously shows, despite ductile deformation, the clear intrusive character of the granite
306 within Ikaria unit (Fig. 3a) as suggested in earlier studies (e.g. Papanikolaou, 1978; Laurent et
307 al., 2015). Along with the Fanari detachment, the Fanari sedimentary unit of Kumerics et al.
308 (2005) was extended further southwest (Fig. 3b). This sedimentary unit was correlated to
309 sediments recognized on the northern part of Ikaria at Gialiskari (Fig. 3c) as initially proposed
310 by Photiades (2002a; 2002b) and recently reinforced by Laurent et al. (2015). There, the
311 Fanari sedimentary unit lies on top of a thick cataclastic body (Figs. 3c and 3d) superimposed
312 over a 300-500 m-thick ductile strain gradient developed within both the Raches and
313 Karkinagrion granites (Laurent et al., 2015). Besides, the initial geological outline of the
314 Karkinagrion granite (Ring, 2007) was significantly modified and extended toward the north
315 (Fig. 2). Detailed field work within and around this granite massif allows the recognition of a
316 large-scale migmatite complex closely associated with this S-type granite (see recent
317 description in Laurent et al. (2015)) (Fig. 3e).

318

319

320 4. Structural analysis of the ductile deformation in metamorphic rocks

321

322 4.1. Main planar fabrics

323

324 All metamorphic rocks and most of magmatic lithologies from Ikaria are pervasively
325 foliated. In most cases, compositional layering in the metasediments has been transposed into
326 a main, generally shallow-angle foliation.

327 221 foliation planes were measured in all lithologies of the two metamorphic units.
328 Besides, bedding was measured in Fanari unit. Measurements, statistical analysis and foliation
329 trajectories are reported on Figs. 4a, b and c, respectively. The dip of foliation planes displays
330 a large range of variation between 10° and 50° for a mean value at $20-25^{\circ}$ (Fig. 4b).
331 Smoothing out these small-scale dip variations, the main foliation planes commonly dip away
332 from the long axis of the island. The strike of foliation therefore shows a fairly concentric
333 pattern depicting a NE-SW elongated structural dome (Fig. 4c), which ends in the northeast.
334 The dome axis, issued from the inversion of field measurements, trends $N037^{\circ}E$, which
335 corresponds to the main orientation of regional foliation (Figs. 4b and 4c). Additionally, this
336 dome presents a marked asymmetry with a steeper, $30-40^{\circ}$ dipping southeastern flank (Fig.
337 4b). The deeper parts of the dome crop out in the central part of the southern coast of the
338 island, close to the contact with the Raches granite. Toward the southwest, the axis extends
339 seaward and turns to a more E-W orientation to finally show up again across the main
340 occurrence of migmatites (Fig. 4c). Strike and dip of the main foliation are generally
341 discontinuous across intrusive contact of the Raches intrusion, particularly to the south.
342 Conversely, foliation trajectories are more continuous to the north where both the granite and
343 the wall-rocks locally present an ultramylonitic fabric (Fig. 4c).

344

345

346 4.2. Stretching lineation

347

348 Stretching lineation is carried by the foliation plane in most metamorphic rocks, while
349 it may sometimes occur as the unique strain marker in the intrusive rocks. It is marked by
350 various indicators, depending on lithology, metamorphic grade and strain intensity. In calcite
351 and dolomite marbles, it is defined by very fine-grained mica slates and the elongation of
352 graphite-rich inclusions. In metapelites, it is generally defined by elongated quartz rods and
353 phyllosilicate aggregates. It is also sometimes marked by the elongation and the truncation of
354 epidote or tourmaline in granitic rocks. Evidence for stretching is also recorded by the
355 preferred elongation and the brittle truncation of clasts in conglomeratic layers at the base of
356 Fanari unit, just above the Fanari detachment.

357 202 stretching lineations have been measured in the field in all metamorphic rocks.
358 Data are all reported in Fig. 5 together with some first-order statistics. At first glance, the
359 trend of lineation shows very little dispersion. It is centered on an average value of N008°E
360 ranging from N160°E and N020°E (Fig. 5b). Wrapped around the dome, the lineation plunges
361 to the north in the northwestern flank of the island and to the south in the southeastern flank.
362 At large-scale, the trend of stretching lineation describes slightly curved patterns from N-S to
363 more NNE-SSW directions. The spatial rotation of the stretching lineation can be correlated
364 with the relative structural position; NNE-SSW orientations are observed in the uppermost
365 parts of the metamorphic succession and particularly in the vicinity of the Fanari shear zone,
366 in the northeast of Ikaria or at Gialiskari (Fig. 5a).

367

368

369 4.3. Asymmetry of ductile deformation

370

371 The whole volume of the metamorphic succession and a great part of the magmatic
372 intrusions of Ikaria are pervasively affected by a top-to-the-N to -NNE ductile deformation

373 (Fig. 5a). Kinematics indicators of top-to-the-N deformation are very common and often
374 unambiguous, particularly toward the top of Ikaria unit and in the bulk of Agios-Kirykos unit.
375 Differing in terms of style, asymmetry and physical conditions of deformation, descriptions in
376 Ikaria and Agios-Kirykos units are presented separately.

377

378

379 4.3.1. Top-to-the-N shearing gradient in Ikaria unit

380

381 Shear bands accompanied by asymmetric boudinage are the most common kinematic
382 indicators in Ikaria unit (Fig. 6). Top-to-the-N shear bands are particularly abundant in
383 metapelite layers interleaved with more competent rocks such as marbles, metaquartzites or
384 metabasites. Due to asymmetric dome-shaped architecture of the island, shear bands display,
385 in their present position, gentle to moderate north-westward dips and normal-sense
386 displacements on the northwestern flank of the island while they often present flat or even
387 “reverse” geometry on the southeastern flank. Boudinage occurs at various scales in
388 alternating lithologies such as metapelites interleaved with marbles, metaquartzites or
389 metabasites. Boudins frequently show asymmetric shapes consistent with a northward
390 asymmetry. However, antithetic bookshelf structures compatible with a top-to-the-N sense of
391 shear can be developed within marble at different scales as observed by Ring (2007).

392 Top-to-the-N ductile deformation appears unevenly distributed within Ikaria unit,
393 primarily controlled by the relative structural position. Evolution of its asymmetry can be
394 studied along a composite cross-section from the deepest parts to the top of this unit from
395 Plagia to Evdilos. The deepest parts crop out on the southeastern coast between Chrisostomos
396 and Plagia. There, the most typical structural feature is an intense folding of the primary
397 compositional layering (Fig. 6a). Folds display a wide range of morphologies but all share

398 common subhorizontal axial planes consistent with vertical flattening. A N-S stretching
399 lineation is only developed in metapelitic or sometimes quartzitic layers while other
400 lithologies display more randomly oriented intersection lineations. Sense of shear is often
401 ambiguous with the presence of both top-to-the-N and top-to-the-S shear criteria suggesting a
402 strong flattening component. Upward, the development of meter-scale decimeter-thick shear
403 bands marks the appearance of a clear asymmetry. Shear bands delimit asymmetric boudins
404 that preserve more ductile deformation as tight recumbent to isoclinal folds, or only as
405 detached fold hinges (Fig. 6b). However, at variance with deeper levels of the unit, fold axes
406 trend parallel to the stretching lineation arguing for a significant component of shearing in the
407 direction of stretching (Fig. 6c). Upward, along with the multiplication of shear bands, the
408 decreasing size of shear domains and boudins, metapelitic rocks become more homogeneous.
409 Lenses of metabasites, dolomitic marbles and remains of metaquartzites occur as sigma or
410 more rarely delta-type porphyroclasts systems consistent with an overall top-to-the-N sense of
411 shear (Fig. 6d). Intense shear strain has turned the rocks into fine-grained mylonites (Fig. 6e).
412 The resulting cross-section shows a first-order strain gradient, where asymmetric stretching is
413 more and more systematic toward the highest structural levels of the unit (Fig. 6f).

414 Metamorphic conditions that prevailed during deformation also depend on structural
415 position. Amphibolite-facies assemblages and gneisses are generally well preserved in the
416 deepest parts of Ikaria unit. There, first retrogression stages are clearly synkinematic, as
417 exemplified by the crystallization of biotite around stable garnets forming sigma-type
418 porphyroclasts system (Fig. 7a). Upward, developed during garnet breakdown, asymmetric
419 strain shadows around garnet contain chlorite and white-micas. Biotite, still metastable in the
420 bulk of the rock is retrograded to chlorite in shear bands. Associated with the crystallization
421 of large amounts of synkinematic chlorite and albite, top-to-the-N shearing in the highest
422 levels was clearly recorded within greenschist-facies metamorphic conditions (Fig. 7b).

423

424

425 4.3.2. Deformation in Agios-Kirykos unit and along Fanari shear zone

426

427 The Agios-Kirykos unit can be studied along a series of foliation-orthogonal valleys
428 from Therma to Agios Kiriaki (see location on Fig. 2). It consists in the alternation of marble
429 and dark metapelites layers reaching 150-200 m of structural thickness. On most outcrops,
430 top-to-the-NNE ductile shearing is clear. Greenschist-facies shear bands are abundant and
431 display a clear upward evolution to more localized or even cataclastic flow (compare Fig. 8a
432 and Fig. 8b). The core of Fanari shear zone, developed at the expense of the uppermost parts
433 of Ikaria unit, is well exposed in the western part of Agios Kiriaki harbor (Fig. 8c). While the
434 contact itself is hidden by the airport runway, Fanari unit crops out directly to the northeast,
435 100 m apart (Fig. 8c). There, rocks present a northeast-dipping mylonitic fabric and carry a
436 N035°E stretching lineation marked by numerous quartz rods and boudins of all sizes from a
437 few centimeters to several meters. Marble layers are stretched and boudinaged within the
438 metapelitic matrix. The internal deformation of marbles displays successive stages of
439 symmetric boudinage, evolving from ductile to brittle (Fig. 8d). Metapelite levels show a
440 clear non-coaxial component of shearing consistent with an overall top-to-the-NNE
441 kinematics (Fig. 8e). These layers, generally thinned to 1 m or even less, display a single and
442 penetrative set of shear bands locally obliterating the main foliation. Spacing between shear
443 bands, controlled by the presence of quartz lenses is locally as small as 1-3 cm (Fig. 8e).
444 Chlorite is abundant within these rocks and adopts a clear synkinematic character in the
445 vicinity of shear bands. The brittle expression of the Fanari shear zone is not exposed there
446 but occurs spectacularly near Fanari, 1 km further south along the coast (Fig. 9). The Fanari
447 detachment is described in the next section.

448

449

450 5. Brittle deformation analysis

451

452 Ikaria presents either large to meso-scale brittle structures or locally pervasive
453 networks of small-scale brittle faults systems developed from the ductile-brittle transition to
454 the brittle field. Brittle structures of all scales are first described. Results of the paleostress
455 analysis are then presented. It must be noted here that some NNE-SSW lineaments, visible on
456 satellite images, do not have any particular expression in the field.

457

458

459 5.1. Description of brittle structures

460

461 The Fanari detachment is the only large-scale structure that was active under both
462 ductile and subsequent brittle conditions. Displacement in the brittle regime over the Fanari
463 detachment is attested by the development of cataclasites particularly well exposed along the
464 southeastern coast of the island as a series of several hundreds of meters of continuous
465 outcrops (Fig. 9). There, the detachment plane appears stripped of its sedimentary cover over
466 a large surface revealing large-scale corrugations and crescent-shaped structures that remind
467 the Cape Evros outcrop on Mykonos (Lecomte et al., 2010; Menant et al., 2013) or the Platy
468 Gialos outcrop on Serifos (Grasemann and Petrakakis, 2007). Breccias, locally reaching more
469 than 2 m, are mainly developed at the expense of Agios-Kirykos unit but also from the very
470 first decimeters of Fanari unit. Cataclastic rocks often show typical clasts size ranging from 1
471 to 20 cm embedded in reddish-brownish cement supposed to be composed of Fe-rich oxy-
472 hydroxides and carbonates. The detachment plane displays a NE-SW strike and a 60°

473 southeast-dip. It carries shallow-dipping, southwest-plunging, large-scale corrugations
474 consistent with the opening of perpendicular cracks. Variable types of kinematics indicators
475 of top-to-the-NNE brittle deformation in the detachment footwall unambiguously ascribe a
476 left-lateral reverse sense of displacement. A secondary discrete striae indicates locally a
477 reverse sense of displacement. At the scale of the map, kinematics of brittle motion over
478 Fanari detachment indicates northeast-directed displacement consistent with the kinematics of
479 the last increments of ductile deformation (compare kinematics given on Fig. 9 with the last
480 ductile stretching lineation on Fig. 5). In its current position, the detachment plane is offset by
481 a set of NW-SE vertical faults carrying oblique striations.

482 Cataclastic rocks also occur on the northern part of Ikaria at Gialiskari (Figs. 3c and
483 3d). They overprint the uppermost parts of a 500 m-thick strain gradient developed
484 exclusively at the expense of the underlying Raches granite (Laurent et al., 2015). Here, they
485 bound again a tectonic unit considered as the lateral equivalent of Fanari unit exposed further
486 east. Sediments, consisting on alternating grey sandstones, conglomerates and brownish
487 limestones are heavily affected by extensional features. WNW-ESE normal faults are
488 organized in conjugate sets accompanied by a dense array of subvertical WNW-ESE veins
489 that both overprint a gently south-dipping bedding.

490 Beside this major brittle structure, small-scale brittle features mostly correspond to late
491 W-E to NW-SE conjugate faults systems developed in all lithologies (Figs. 10 and 11).
492 Displacement on faults ranges typically from a few centimeters to a few meters and can
493 generally be constrained in the field using marker-levels. In the entire studied area, the fault
494 population is dominated by dihedral orientations arguing for the formation of newly formed
495 conjugate sets of faults rather than reactivated shear planes. Normal faults are often
496 accompanied by subvertical joints and tension gashes filled with quartz, chlorite, iron oxy-
497 hydroxides in the metamorphic rocks, quartz, chlorite or even epidote and tourmaline in

498 granites and mostly calcite in the sediments. The close association between faults and joints
499 often displays contradictory intersection relationships and thus argues for a contemporaneous
500 development.

501

502

503 5.2. Inversion of fault-slip data

504

505 Paleostress orientation patterns were evaluated by the Win-Tensor computer-aided
506 inversion software (Delvaux and Sperner, 2003). The reduced paleostress tensor consists in
507 the identification of orientation of the three principal stress axes and the axial ratio of stress
508 ellipsoid. Determination of the paleostress axes was completed by the analysis of
509 accompanying brittle structures such as joints, tension gashes or stretched pebbles. Brittle
510 structure analysis was conducted over 9 main sites scattered over the island (see Fig. 5 for
511 location). Results are presented in the next two sections.

512

513

514 5.2.1. Using microstructures to unfold the Fanari detachment

515

516 Although the Fanari detachment consistently exhibits top-to-the-NE kinematics, it
517 currently crops out with the geometry of southeast-dipping left-lateral reverse fault zone, a
518 quite rare configuration in the Aegean domain (Fig. 9). A detailed study of geometrical
519 relationships between detachment, sediments of Fanari unit and attitude of small-scale brittle
520 structures affecting either the sediments or the detachment plane, allows proposing a
521 restoration scenario of the initial geometry of the Fanari detachment (Fig. 10a). Two reference
522 outcrops (6 and 7) were selected 2 km apart within sediments of Fanari unit along the

523 detachment (see position on Fig. 5) with contrasting spatial relationships between structures.
524 Outcrop of station 7 (Fig. 10b), which presents almost vertical bedding, displays a conjugate
525 set of subvertical faults affecting alternating sandstones and conglomeratic layers. N030°E to
526 N160°E faults present consistent sinistral kinematics while N070°E to N130°E faults present
527 dextral kinematics. Part of these faults even cuts across the Fanari detachment causing 1-5 m
528 offsets of the plane. Outcrop of station 6 also presents two sets of faults that affect a 40°
529 dipping bedding further southeastward. The dominant set of faults displays subvertical fault
530 planes carrying both normal and reverse kinematics. This set is accompanied by a subordinate
531 set of flat to gently-dipping fault planes that also present both normal and reverse kinematics
532 (Fig. 10b). In their present geometry, this heterogeneous fault set requires the superimposition
533 of two distinct stress regimes. Back-tilting of those systems of faults about a horizontal axis,
534 in order to obtain a horizontal bedding, permits to get a coherent stress regimes for both sites
535 6 and 7. Maximum principal stress axis becomes vertical and the two others become
536 horizontal, compatibly with a NE-SW extension for both sites. It strongly suggests that faults
537 form before an unequal tilting of the different outcrops. This assumption is confirmed further
538 to the southwest where both the detachment plane and the bedding of sediments are
539 subhorizontal (i.e. less than 20° toward the southwest) while tension-gashes are now
540 subvertical and all faults appear as a single conjugate normal fault set preserved in its initial
541 attitude. Importantly, back-tilting of the whole system, including the Fanari detachment,
542 results in a system where the detachment plane operates at shallow angle, 10-15° to the
543 northwest, with normal-dextral-sense kinematics and cuts the sediments down-section as
544 observed in the neighboring island of Mykonos (e.g. Lecomte et al., 2010; Menant et al.,
545 2013). In this restored position (Fig. 10a), sediments are pervasively affected by normal
546 faulting consistent with a unique and common NNE-SSW to NE-SW extension (Fig. 10b).
547 This paleostress solution is consistent with the paleostress field deduced from other stations

548 throughout the island and previous studies (e.g. Kumerics et al., 2005). This point is discussed
549 below. The causes of the Fanari detachment tilting and the more general arching of the ductile
550 to brittle fabrics at the scale of the whole island are addressed in the discussion section.

551

552

553 5.2.2. Large-scale consistency of the paleostress inversions

554

555 Despite the local coexistence of both N80-110°E and N130-160°E trending normal
556 faults, results present rather small internal dispersion (Fig. 11). At most stations, stress tensor
557 analysis shows a consistent subvertical orientation for the maximum principal stress axis (σ_1).
558 In other stations (e.g. sites 6 and 7 on Fig. 10b), σ_1 was restored using horizontal-axis back-
559 tilting rotation. The minimum principal stress axis (σ_3) is horizontal or very gently dipping
560 with a consistent NNE-SSW to more NE-SW direction of extension. In turn, the overall
561 consistency of the direction of σ_3 prevents from significant vertical-axis large-scale block
562 rotations. Planes of joint and tension gashes correspond to the calculated σ_1 - σ_2 plane, and
563 their poles therefore appear scattered around the σ_3 axis. Brittle structures recorded during the
564 last exhumation stages present a marked consistency of a NE-SW stretching direction
565 occurring in an extensional regime.

566

567

568 6. Thermal structure

569

570 The RSCM method is based on the quantitative study of the degree of organization of
571 carbonaceous material (CM), which is a reliable indicator of metamorphic temperature (e.g.
572 Pasteris and Wopenka, 1991; Beyssac et al., 2002; Lahfid et al., 2010). Because of the

573 irreversible character of graphitization, CM structure is not sensitive to the retrograde
574 reactions and permits to determine peak temperature conditions (T_{\max}) reached during
575 metamorphism, using an area ratio (R2 ratio) of different peak of Raman spectra (Beysac et
576 al., 2002). T_{\max} can be determined in the 300-640 °C range with an accuracy of ± 50 °C
577 related with the precision and the dispersion of petrological data used for the method
578 calibration (Beysac et al., 2002). Relative uncertainties on T_{\max} are however much smaller,
579 around 10-15 °C and relative variations of that order of magnitude can be detected (e.g.
580 Gabalda et al., 2009; Vitale Brovarone et al., 2013; Augier et al., 2015). Sampling was
581 performed in order to quantitatively describe the large-scale thermal structure of Ikaria to
582 complement the few existing, punctual P-T estimates (Kumerics et al., 2005; Martin et al.,
583 2011). Sampling was therefore regularly distributed within the tectonometamorphic
584 succession of Ikaria and a more systematic sampling was carried out in the vicinity of possible
585 second-order thermal effects such as intrusions and tectonic contacts. 35 samples consisting of
586 CM-bearing metasediments were collected (see Fig. 12 for location). In order to bring out and
587 possibly smooth out the inner structural heterogeneity of CM within samples, 13 to 20 spectra
588 were recorded for each sample. Detailed results, including R2 ratio, number of spectra, T_{\max}
589 and standard deviation are presented on Table 1. In addition, RSCM temperatures are all
590 reported on Fig. 12.

591 RSCM results embrace a wide range of temperature from 391 to 625 °C (Table 1). At
592 first glance, T_{\max} presents a correlation with the relative structural position. 625 °C was
593 indeed recorded for the deepest parts of the structural dome with gneisses while 390 °C was
594 retrieved for the uppermost parts of the metamorphic succession where low-crystallinity
595 micaschists crop out. Ikaria therefore appears as a metamorphic dome in which temperature
596 increases down-section (Figs. 12a and 12b). Samples from Ikaria unit show rather high T_{\max} ,
597 from 625 to 500 °C, presenting an important scatter. The 550 °C temperatures for the volume

598 of Ikaria unit are moreover consistent with the amphibolite-facies metamorphic associations.
599 Samples from Agios-Kirykos unit display lower temperature ranging between 450 and 390 °C
600 (Table 1). Two main sets of temperatures can then be identified separated by a 50 °C
601 temperature gap (Figs. 12a and 12b). This gap of temperature corresponds to the position of
602 the Agios-Kirykos shear zone, independently recognized on the basis of structural criteria.
603 This particular feature was studied in the east of Ikaria, and particularly along a foliation-
604 orthogonal cross-section upstream of Therma valley (Fig. 12b). There, T_{\max} shows a regular
605 temperature decrease from 600 to 400 °C accompanied by a 50 °C normal-sense metamorphic
606 gap. Among the causes of temperature variation in Ikaria unit, local heating by intrusive rocks
607 has been tested in the vicinity of the Raches granite (Fig. 12c). Samples were picked at
608 decreasing distance from the intrusive contact for almost the same relative structural position.
609 Results show regularly decreasing temperatures from 580 °C to 520 °C, in accordance with
610 the intrusive character of the Raches granite (Fig. 3a). Accordingly, smaller intrusive bodies
611 as the Kefala diorite or small-scale pegmatite dykes that are only partly mapped (Fig. 2) may
612 be responsible of erratic, isolated high T_{\max} results.

613

614

615 7. Age constrains on the partial melting event

616

617 7.1. Analytical methods

618

619 Analyses were performed in thin section by laser ablation inductively coupled plasma
620 spectrometry (LA-ICPMS) at the Laboratoire Magmas et Volcans (LMV), Clermont-Ferrand
621 (France). The ablation is performed using a Resonetics Resolution M-50E system equipped
622 with an ultra-short pulse (<4ns) ATL excimer 193 nm wavelength laser. This laser system is

623 coupled with Agilent 7500 cs ICP-MS equipped with a pumping system to enhance the
624 sensitivity. Spot diameter of 9 μm was used with a 1 Hz repetition rates. Ablated material is
625 transported using a helium flux, and then mixed with nitrogen and argon before being injected
626 into the plasma source. Analytical procedures for monazite dating are reported in details in
627 Didier et al. (2013), Didier et al. (2014) and Paquette and Tiepolo (2007). Following isotopes
628 $^{204}(\text{Pb}+\text{Hg})$, ^{206}Pb , ^{207}Pb , ^{208}Pb , ^{232}Th and ^{238}U were acquired. Data disturbed by inclusions,
629 fractures or age mixing (between different areas of a single grain) were not taken into account
630 for calculation. The occurrence of initial Pb in the sample can be monitored by the evolution
631 of the $^{204}(\text{Pb}+\text{Hg})$ signal intensity, but no Pb correction was applied. Elemental fractionation
632 and mass bias were corrected making repeated analyses on Trebilcock monazite (272 ± 2 Ma,
633 Tomascak et al., 1996). Analyses on the Moacyr monazite (Cruz et al. 1996; Seydoux-
634 Guillaume et al. 2002; Gasquet et al. 2010; Fletcher et al. 2010) at the beginning and at the
635 end of each session, treated as unknowns, verify the reproducibility, especially for the
636 $^{208}\text{Pb}/^{232}\text{Th}$ ages, and the accuracy of the corrections. Data reduction was carried out with
637 GLITTER® software package (van Achterbergh et al. 2001; Jackson et al. 2004). Calculated
638 ratios were exported and age diagrams were generated using Isoplot software package by
639 Ludwig (2001).

640 In this study, $^{208}\text{Pb}/^{232}\text{Th}$ ages are preferably used because: 1) U decay series could be
641 in disequilibrium in young monazites (Schärer, 1984), resulting in overestimated $^{206}\text{Pb}/^{238}\text{U}$
642 ages; 2) ^{232}Th is so abundant that ^{208}Pb originating from initial Pb is negligible compared to
643 radiogenic ^{208}Pb (Janots et al. 2012; Didier et al. 2013). However, in this study, $^{206}\text{Pb}/^{238}\text{U}$
644 ages are fully consistent with the $^{208}\text{Pb}/^{232}\text{Th}$ ages suggesting the absence of disequilibrium in
645 the U decay series. Slight common Pb contamination is suggested by the Tera Wasserburg
646 diagram, enhanced by the large uncertainty of the $^{207}\text{Pb}/^{235}\text{U}$ ages due to very low ^{207}Pb
647 content in young monazite ages.

648

649

650 7.2. Results

651

652 Fourteen monazites have been analyzed for U-Th-Pb in situ dating in thin section from
653 one migmatite sample (IK01A; see Fig. 2 for localization; Qz, Pl, Afs, Bt, Sil and Zrn).
654 Monazites often appear as inclusions in biotite or set at grain boundaries and occur as large
655 pristine crystals (Fig. 13a; between 60 and 100 μm) showing in few cases irregular
656 boundaries. Besides, most of the crystals present a clear core and rim texture (Fig. 13b) which
657 represents variable Y and U contents : the core is Y and U-poor (Y_2O_3 between 0.3 and 1.6
658 wt% and UO_2 between 0.2 and 0.8 wt%) whereas the rim is Y and U-rich (Y_2O_3 between 1.7
659 and 4.2 wt% and UO_2 up to 3.7 wt %).

660 Thirty two analyses have been performed in the different domains of the monazite
661 grains. $^{208}\text{Pb}/^{232}\text{Th}$ analyses yielded ages between 14.9 ± 0.5 and 16.4 ± 0.5 Ma, with a
662 weighted mean age at 15.7 ± 0.2 Ma (MSWD = 4,1; N=32) . All ages appear always
663 concordant with the $^{206}\text{Pb}/^{238}\text{U}$ ages (Fig. 13c). The U-Pb ages are mostly discordant and
664 define a linear trend crosscutting the Concordia at 15.1 ± 0.2 Ma in a Tera Wasserburg
665 diagram (Fig. 13d). This suggests a slight common Pb contamination and the large uncertainty
666 of the $^{207}\text{Pb}/^{235}\text{U}$ ages due to the very low ^{207}Pb content in young monazite. For this reason,
667 only $^{208}\text{Pb}/^{232}\text{Th}$ ages are considered here (see details in analytical methods).

668 The scattering of the $^{208}\text{Pb}/^{232}\text{Th}$ ages (high MSWD) is broadly correlated to the core
669 to rim zonation, the Y and U-poor cores yielding older ages than the Y and U-rich rims (Fig.
670 13b). However, the small difference between the ages measured in the cores and in the rim of
671 the monazite grains (less than 1.5 Ma) does not enable a clear distinction between two age
672 groups. Because i) the monazite is well known to be an efficient chronometer in dating the

673 high temperature metamorphic processes, and ii) no inheritance (suggesting the recording of
674 older metamorphic event) has been observed, these results clearly establish a Langhian age
675 for the partial melting event that affected the infrastructure of Ikaria Island.

676

677

678 8. Discussion

679

680 8.1. An overall top-to-the-N sense of shear as the main deformation record

681

682 The whole tectonometamorphic succession of Ikaria is pervasively affected by top-to-
683 the-N to -NNE ductile deformation whose intensity increases toward the top of the structure
684 (Figs. 6f and 14a). While the deepest parts of the tectonometamorphic succession exhibits
685 rather symmetrical deformation dominated by flattening, top-to-the-N shearing deformation
686 shows a large-scale strain gradient toward the Fanari shear zone along which most of the
687 strain is concentrated, with an upward evolution toward a more top-to-the-NNE kinematics
688 (Fig. 5). Characterized by a subordinate top-to-the-NNE strain gradient, the Agios-Kirykos
689 shear zone is furthermore responsible for a 50 °C gap of normal sense, as retrieved by RSCM
690 analyses (Fig. 12). These embedded gradients are accompanied by greenschist-facies
691 retrogression gradients quite obvious at the scale of the outcrop or hand-specimen (Fig. 7).
692 Crossing the ductile to brittle transition, the Fanari shear zone has even recorded increments
693 of motion in brittle conditions responsible for the development of thick cataclasite bodies
694 (Figs. 3 and 9). The direction of ductile stretching and markers of later brittle extensional
695 motions on the detachment planes show a continuum of NNE-SSW stretching and top-to-the-
696 NNE shearing, consistent, in a broad sense, with extensional direction in both metamorphic
697 and sedimentary rocks (Fig. 11). This evolution is interpreted as an evidence of continuous

698 stretching from middle to upper crustal levels through the ductile-brittle transition and
699 continuous shearing along a major detachment localizing deformation.

700 Low-temperature geochronology pointed out a clear northward younging of ZFT ages
701 (10.3 ± 0.3 to 7.1 ± 0.3 Ma) and AFT ages (8.4 ± 0.8 to 5.9 ± 0.8 Ma) consistent with an
702 overall top-to-the-N unroofing with apparent slip-rates over the Fanari detachment of 10-8
703 km/Ma (Fig. 14b; Kumerics et al., 2005). U-Th/He analyses on apatite for both flanks of the
704 dome yielded similar 6 Ma ages. Motion over the Fanari detachment seems stop near 6 Ma.
705 These ages constrain, in turn, the maximum age for the onset of passive “folding” of the
706 Fanari detachment plane around the dome, which postdates the last brittle motions of the
707 detachment and even the brittle normal faulting that affects sediments.

708 A progressive rotation of the stretching and shearing direction is observed from the
709 core to the outer flanks of the domes (Fig. 5). As suggested by the structural study, the
710 shearing strain has progressively localized through time upward from a distributed flow,
711 recorded in the amphibolite-facies conditions, to mylonites, ultramylonites and then
712 cataclasites in the vicinity of Fanari detachment (Figs. 3 and 9). Accordingly, the N-S
713 trending lineation preserved in the core is older than the NNE-SSW trending one at the top
714 along the main shear zones. The core of the dome has therefore rotated 35° counterclockwise
715 during its exhumation before the final doming that seems to be the latest event recorded on
716 Ikaria. This late event tends to confer to the dome the geometry of an a-type dome where the
717 stretching direction is parallel to the dome axis (Jolivet et al., 2004a) and which can suggests
718 an oblique component of shearing during doming (Le Pourhiet et al., 2012). A significant
719 component of E-W shortening is currently recorded in the deformation of the northern Aegean
720 domain as a result of westward motion of Anatolia along the North Anatolian Fault (NAF)
721 (e.g. Le Pichon and Kreemer, 2010). If the NAF has reached the northern Aegean some 6 Ma
722 ago as a localized crustal-scale fault zone (Armijo et al., 1999; Melinte-Dobrinescu et al.,

723 2009), it has been suggested that dextral movements has been active since 12 Ma (Sengör et
724 al., 2005). In the field, consequences of an E-W shortening are clearly recorded in the
725 northern Cyclades, all the way to western Turkey (Angelier, 1976; Buick, 1991; Bozkurt and
726 Park, 1997; Ring et al., 1999; Avigad et al., 2001; Menant et al., 2013). This context, showing
727 a combination of exhumation, strike-slip and contemporaneous E-W shortening and N-S
728 crustal stretching, suggests a component of transtension during the last evolution stages (Le
729 Pourhiet et al., 2012; Fossen et al., 2013). This oblique component fits the hypothesis of a
730 left-lateral transfer zone from the Menderes massif to the Cyclades proposed by Ring et al.
731 (1999) and recently reassessed by Gessner et al. (2013). Detailed studies on the timing of the
732 formation of stretching directions across the dome coupled with the timing of doming are
733 necessary to go any further.

734

735

736 8.2. Eastern extension of the North Cycladic Detachment System

737

738 Obvious structural and metamorphic similarities of the series of detachments from
739 Andros to Mykonos, with coeval top-to-the-N to -NNE kinematics, led Jolivet et al. (2010) to
740 mechanically merge them in a single crustal-scale extensional structure, called NCDS. A
741 probable correlation of the Fanari detachment of Kumerics et al. (2005) with the NCDS was
742 even proposed (Jolivet et al., 2010). The existence on Ikaria of two detachments, one purely
743 ductile and the other evolving from ductile to brittle conditions and the presence of the
744 synkinematic granite piercing through the dome are indeed reminiscent of the anatomy of the
745 NCDS as described on Andros, Tinos and Mykonos (Jolivet et al., 2010). On Tinos (e.g.
746 Jolivet and Patriat, 1999; Jolivet et al., 2004a; Brichau et al., 2007) as well as on Mykonos
747 (e.g. Faure et al., 1991; Lecomte et al., 2010; Denèle et al., 2011; Menant et al., 2013),

748 increments of ductile and then brittle deformation are recorded by synkinematic intrusions
749 emplaced into the main extensional shear zones. The Tinos Detachment is mostly ductile and
750 is pierced by the Tinos intrusion. Conversely, the Livada Detachment shows ductile then
751 brittle deformation and affects the Tinos intrusion (Brichau et al., 2007). This last detachment
752 is then pierced by the Mykonos intrusion. The more superficial Mykonos Detachment that
753 clearly post-dates all intrusions, operated only in purely brittle conditions and carries syn-
754 extension sediments (e.g. Lecomte et al., 2010; Menant et al., 2013). These three detachments
755 that operate sequentially form the NCDS. A similar evolution is observed on Ikaria.
756 Exclusively operating in the ductile field, the Agios-Kirykos shear zone is later cut down-
757 section by the Fanari detachment that evolves from ductile to brittle and even controls the
758 deposition of syn-tectonic sediments in the hangingwall.

759 Along with the current lack of precise time-constraints on synkinematic minerals,
760 timing of extensional deformation onset is still unclear on Ikaria. If partial melting occurred at
761 ca. 15.5 Ma, as showed in this study, the signification of the 25-17 K/Ar ages on hornblende
762 (Altherr et al., 1982) are questioned. These ages may reflect either Ar inheritance from an
763 earlier tectonometamorphic event or simply excess argon (i.e. extraneous argon), as suggested
764 by the strong scattering of these ages. Then, onset of extension might starts at or just after ca.
765 15.5 Ma. The 12-9 Ma K/Ar and Ar/Ar ages obtained on fabric-forming white-micas and
766 biotite are believed to record either a fast cooling or last recrystallizations during
767 mylonitization (see compilation on Fig. 14b; Altherr et al., 1982; Kumerics et al., 2005). In
768 parallel, the three intrusions (Raches, Karkinagrion and Xylosyrtis) that crystallized at 14-13
769 Ma therefore synkinematically emplace within the Fanari shear zone (U-Pb on zircon; Bolhar
770 et al., 2010; Laurent et al., 2015). Intrusions all underwent a common fast cooling to
771 temperatures close to the ductile to brittle transition (Stöckhert et al., 1999; Imber et al., 2001)
772 at ca. 10 Ma (K/Ar and Ar/Ar on micas; Altherr et al., 1982; Kumerics et al., 2005). Being

773 only active in the ductile field, the Agios-Kirykos therefore probably ceased to be active some
774 10 Ma ago. Displacement and further strain localization were thus transferred to the Fanari
775 shear zone, responsible for the final exhumation, in the ductile and then the brittle field. The
776 last exhumation stages were characterized by fast (100-80 °C/Ma) cooling rates and ended at
777 around 6 Ma (Fig. 14b; FT and U-Th/He on zircon and apatite; Kumerics et al., 2005). As a
778 comparison, crystallization and cooling of the Mykonos pluton occurred between 13.5 Ma and
779 9 Ma as a result of the fast exhumation related to the activity of the Livada and Mykonos
780 detachments (e.g. Brichau et al., 2008; Lecomte et al., 2010; Jolivet et al., 2010). Low-
781 temperature ages even suggest that motions over the Fanari detachment continued until 6 Ma
782 and are thus the last top-to-the-N to -NNE detachment active in the Aegean domain. These
783 similarities in terms of geometries, kinematics and timing of the Ikaria detachments and the
784 various branches of the NCDS confirm the proposed eastward extension of the NCDS all the
785 way to the eastern end of Ikaria in the Agios-Kirykos and Fanari shear zones (Fig. 15).

786 Similarly to Mykonos, the detachment system of Ikaria, does not exhume *HP-LT*
787 metamorphic rocks belonging to the Cycladic Blueschists unit, at variance with Tinos and
788 Andros. Indeed, along with the structural position and the lithostratigraphic succession, the
789 apparent lack of any *HP-LT* imprint either in Ikaria and Agios-Kirykos units precludes
790 correlations with the Cycladic Blueschists unit. The Ikaria and Agios-Kirykos units are
791 characterized by the same monotonous lithologies and a similar 80-100 °C/km thermal field
792 gradient, it is proposed that both units were equilibrated along a single, warm gradient rooting
793 in partially-molten rocks.

794 Correlated with the Upper Cycladic unit, in a broad sense, the Fanari unit reworks
795 large amount of Pelagonian detritus shed into the extensional basins (Photiades, 2002a).
796 While deposition of sediments on the Upper Cycladic unit started around 23 Ma throughout
797 the Aegean domain (Angelier et al., 1976; Sánchez-Gómez et al., 2002; Kuhlemann et al.,

798 2004), sediments preserved on Ikaria are attributed to Late Oligocene to Early Pliocene
799 (Papanikolaou, 1978; Photiades, 2002a; 2002b). This rather recent deposition age is consistent
800 with last denudation ages ascribed to last extensional motions over the Fanari detachment
801 (Kumerics et al., 2005), which may be responsible for maintaining a significant tectonic
802 subsidence.

803

804

805 8.3. Definition of the Ikaria MCC and the central Aegean HT zone

806

807 The structure of Ikaria is composed of two main parts separated by a detachment
808 system consisting in shallow-dipping extensional ductile to brittle shear zones connecting
809 surficial to mid-crustal levels (e.g. Fig. 2). The bulk finite architecture of the footwall consists
810 in an elongated structural dome roofed by a partly eroded mylonite, ultramylonite and
811 cataclasite carapace (Fig. 4). This geometrical dome is paired with a fairly concentric
812 distribution metamorphic zones of decreasing grade from migmatites and 625 °C in the core
813 to 390 °C in the external parts (Fig. 12). The dome flanks display 500-m thick strain gradient
814 evolving upward to mylonites characterized by the uniformity of shear sense from one limb to
815 the other. They are partially overprinted by cataclasites and fault-rocks in the Fanari
816 detachment zone (Figs. 3 and 9). The detachment itself is marked by a clear-cut fault plane
817 carrying evidences for large-scale displacements in the brittle field (Fig. 9). The hangingwall
818 unit, exclusively composed sediments, is only characterized by brittle deformation (Fig. 10).
819 Following this description, Ikaria dome shares all attributes of a typical migmatite-cored
820 MCC (see Platt et al., 2014, for review) firstly described in the Basin and Range of western
821 United States (Davis and Coney, 1979; Crittenden et al., 1980; Wernicke, 1981, 1985), and
822 then in other Alpine (e.g. Lister et al., 1984; Dewey, 1988; Gautier et al., 1993; Gautier and

823 Brun, 1994; Jolivet and Patriat, 1999; Vanderhaeghe, 2004) and Variscan or Caledonian
824 orogenic belts (e.g. Norton, 1986; Andersen et al., 1991).

825 Partial melting is a key factor that controls the strength of the continental crust. The
826 recent discovery of migmatites in the infrastructure of the Ikaria MCC dome may have
827 important implications for the behavior of the Hellenic continental crust during Oligo-
828 Miocene extension. Several high-grade gneiss and migmatite massifs have already been
829 described in the Aegean realm. However, most of these occurrences of HT rocks are basement
830 rocks inherited from ancient tectonometamorphic events. An early Paleozoic HT event is
831 known in the Menderes (e.g. Gessner et al., 2004) and in the East of the Aegean Sea,
832 particularly on Ikaria (Kumerics et al., 2005; Ring et al., 2007). There, a ca. 460 Ma age
833 retrieved from a pegmatite dyke may reflect a minimum age for HT metamorphism (Kumerics
834 et al., 2005). A Variscan inheritance is also well described in the Cycladic basement unit in
835 the Southern Cyclades (e.g. Henjes-Kunst and Kreuzer, 1982; Andriessen et al., 1987; Keay
836 and Lister, 2002). On these islands, a ca. 320 Ma granite and amphibolite facies gneisses are
837 partially overprinted by the Eocene HP imprint (e.g. Henjes-Kunst and Kreuzer, 1982;
838 Andriessen et al., 1987; Keay and Lister, 2002; Huet et al., 2009; Augier et al., 2015). The
839 only Oligo-Miocene migmatites crop out in the core of the Naxos MCC, the first that has been
840 described in the Aegean domain (e. g. Lister et al., 1984). There, the age of the partial melting
841 is quite constrained by U-Pb analyses on zircon yielding series of ages ranging from 21 to 17
842 Ma (Keay et al., 2001). The age of migmatites recently discovered on Ikaria is therefore of
843 prime importance. In this study, pioneer U-Th-Pb analyses on monazites from a leucosome of
844 the migmatites yielded an age of 15.7 ± 0.2 Ma (Fig. 13). This result falls in the same age-
845 range than the recent U-Pb age on zircon ascribed to the Karkinagrion S-type granite (Bolhar
846 et al., 2010) together with closely associated migmatites (Laurent et al., 2015). Clearly
847 associated with the same large scale partial melting event that affected the Hellenic

848 continental crust, anatexis at the latitude of Ikaria appears slightly younger than further south
849 on Naxos. Along with Naxos, Mykonos and Ikaria, HT MCCs all cluster in the central part of
850 the Aegean domain. It is here proposed to define the central Aegean HT zone as a large-scale
851 extensional tectonic window of high-grade, partially-molten rocks (Fig. 15), of Early-Middle
852 Miocene age (Keay et al., 2001; this study) (Fig. 15). There, intense stretching leads to the
853 complete tectonic omission of the Cycladic Blueschists unit that is observed in colder MCCs
854 further west in Syros, Andros or Tinos (e.g. Trotet et al., 2001a; Mehl et al., 2007), to the east
855 in Samos (e.g. Ring et al., 1999) or to the south in Ios, Folegandros and Sikinos (e.g. Huet et
856 al., 2009; Augier et al., 2015). This gradient of finite stretching toward the center and east of
857 the Cyclades is attested by the joint evolution of the topography, the crustal thickness and the
858 first order transition from brittle to ductile deformation from marginal areas to the center of
859 the Aegean domain (e.g. Jolivet et al., 2010). Uprise of partially-molten materials in-between
860 less-extended Cycladic Blueschists unit evokes crustal-scale equivalents of the scar folds that
861 result from “boudins” separation in the process of boudinage (Tirel et al., 2008). This simple
862 model should however be adapted as three distinctive detachment bounding smaller-scale
863 MCCs are currently observed (Fig. 15). The NCDS (Jolivet et al., 2010) exhumes Andros,
864 Tinos, Mykonos and Ikaria in the north while the NDP exhumes Naxos and Paros in the
865 center (Buick, 1991; Gautier et al., 1993; Vanderhaeghe, 2004; Duchêne et al., 2006;
866 Kruckenberg et al., 2011). Both are associated with top-to-the-N shear. Finally, further
867 southwest, the alignment of Kea, Kythnos and Serifos shows another row of small-scale
868 domes exhumed by the top-to-the-SW WCDS (Grasemann et al., 2012). If top-to-the-S
869 criteria on Ios and Sikinos are related to a thrust (Huet et al., 2009; Augier et al., 2015) and
870 not to a detachment (Vandenberg and Lister, 1996; Forster and Lister, 1999; Thomson et al.,
871 2009; Ring et al., 2011) then extensional top-to-the-N shearing is systematically observed
872 from north to south of the central Aegean (Fig. 15). This result contrasts with the more

873 symmetric deformation recognized in marginal areas, where symmetrically arranged
874 detachment systems depict more bivergent extensional systems either to the west (Grasemann
875 et al., 2012) or east in Samos (Ring et al., 1999) or in the Menderes massif (Hetzel et al.,
876 1995b; Gessner et al., 2001). A correlation between the amount of stretching of the crust and
877 the degree of non-coaxiality is therefore proposed. Just as the Tyrrhenian or the Alboran back-
878 arc systems, the causes of this regional-scale non-coaxial extension in the Aegean back-arc
879 domain is a first-order question that is outside the focus of this paper but may pertain to the
880 interactions between the upper plate, the retreating subducting slab and the flowing
881 asthenospheric mantle (Jolivet et al., 2009; Sternai et al., 2014).

882

883

884 9. Conclusions

885

886 Although Ikaria is one of the largest Aegean islands and paves the way between the
887 Cyclades and the Menderes massif, its geology and tectonic evolution remain poorly
888 understood since recent studies lead to very conflicting maps and interpretations. This paper
889 reports a detailed 3D study of the geometry and the thermal structure of Ikaria. The structural
890 study shows that the HT-LP foliation is arched, forming a NE-SW trending structural dome
891 cored by partially molten rocks and intruded by late intrusive granitic bodies. Lineation shows
892 a N-S to NNE-SSW ductile stretching associated with an overall top-to-the-N to -NNE sense
893 of shear. Final exhumation of the dome was completed by Miocene extensional system made
894 of two main top-to-the-N to -NNE shear zones, operating in the ductile and then the brittle
895 fields. The thermal structure revealed by the RSCM approach strengthened the subdivision of
896 the metamorphic succession in two main metamorphic units (Ikaria and Agios-Kirykos units)
897 with an upward decrease of maximum temperature, separated by the Agios-Kirykos shear

898 zone. The Fanari detachment permits to juxtapose the sedimentary Fanari unit directly on
899 metamorphic rocks. The distribution of RSCM temperatures within the dome and the presence
900 of migmatites of ca. 15.5 Ma age in the western part of the island fit the description of a *HT*
901 MCC such as Naxos or Mykonos. The proposed tectono-metamorphic evolution of the dome
902 is consistent with the evolution of the northern Aegean area controlled by the eastern
903 extension of the NCDS. Definition of Ikaria as a *HT* MCC allows the definition of the
904 geographic extent of the central Aegean *HT* zone associated with strictly asymmetric top-to-
905 the-N ductile shearing and an Early-Middle Miocene partial melting event. A correlation
906 between the amount of stretching of the crust and the degree of non-coaxiality is therefore
907 proposed.

908

909

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1510

1511 Table caption

1512

1513 Table 1: RSCM peak temperature results

1514 RSCM results classified by increasing temperatures. For each sample are indicated the GPS
1515 position, the total number of Raman spectra (n) performed, the mean R2 ratio and temperature
1516 calculated, both associated with their standard deviation (SD) related to the intra-sample
1517 heterogeneity. Location of samples and T_{max} results are given on Fig. 12.

1518

1519

1520 Figure captions

1521

1522 Figure 1: Tectonic map of the Aegean domain

1523 Tectonic map of the Aegean domain showing the main geological units and structures related
1524 to both synorogenic and postorogenic episodes, modified after Jolivet et al. (2013) and
1525 references therein. Original map has been modified incorporating recent works (Ring et al.,
1526 1999; Kumerics et al., 2005; Rosenbaum et al., 2007; Huet et al., 2009; Grasemann et al.,
1527 2012; Augier et al., 2015). NCDS: North Cycladic Detachment System; WCDS: West
1528 Cycladic Detachment System; NDP: Naxos-Paros Detachment; SCT: South Cycladic Thrust.

1529

1530 Figure 2: Geological map of Ikaria

1531 (a) New geological map of Ikaria proposed in this study. Lithologic outlines correspond to
1532 new field observations and a compilation of existing maps (e.g. Papanikolaou, 1978;

1533 Photiades, 2002b; Kumerics et al., 2005; Ring, 2007). Also indicated are two representative
1534 tectonometamorphic piles for the eastern and the western parts of Ikaria.

1535

1536 Figure 3: New geological features and precisions about nature of contacts

1537 (a) The intrusive contact between Raches granite and marbles of Ikaria unit is clearly marked
1538 by the presence of granitic dykes within marbles. (b) Southward extension of the Fanari
1539 detachment at Therma, separating conglomerates of Fanari unit and metasediments of Agios-
1540 Kirykos unit. (c) Panorama of the Fanari detachment at Gialiskari between sediments of
1541 Fanari unit and the Raches granite. (d) Close-up view of granite-derived cataclasites beneath
1542 the Fanari detachment at Gialiskari. (e) Details of migmatites from the lower parts of Ikaria
1543 unit preserved in the south of Raches granite. See Fig. 2 for location.

1544

1545 Figure 4: Main planar fabrics

1546 (a) Foliation-map of Ikaria. Geometry of sedimentary bedding in Fanari unit is also showed.
1547 (b) Statistics of the main foliation geometry. Poles of foliation are presented in Schmidt's
1548 lower hemisphere equal-area projection and preferred orientations of foliation is given by the
1549 rose-diagram. The elongation of the cloud allows retrieving the geometry of the dome axis.
1550 Also note that the asymmetry of the cloud calls for the asymmetry of the dome with a steeper
1551 southeast flank. (c) Simplified geological map showing foliation trajectories and traces of the
1552 main shear zones.

1553

1554 Figure 5: Stretching lineation

1555 (a) Stretching lineation-map of Ikaria. Note that the trend of the lineation shows very little
1556 dispersion. It is noteworthy that the strike of the stretching lineation describes slightly curved
1557 patterns from N-S to more NNE-SSW directions. Indicated are the results of the fault-slip data

1558 inversion. (b) Statistics on the preferred orientation of stretching lineation presented in
1559 Schmidt's lower hemisphere equal-area projection.

1560

1561 Figure 6: Upward strain gradient within Ikaria unit

1562 (a) Incomplete transposition of the compositional layering into a main, flat-lying penetrative
1563 foliation within the deepest parts of the structure. (b) More advanced transposition of the
1564 compositional layering into the flat-lying penetrative foliation. Note that boudin-shaped
1565 volume of rocks preserved isoclinal folds between decameter-scale top-to-the-N shear bands.
1566 (c) Close-up view of the cross-cutting relationships compositional layering and the main
1567 foliation. Note that fold axes are now mostly parallel to the stretching direction. (d) Mylonitic
1568 deformation of the upper parts of Ikaria unit. The foliation dips gently toward the north.
1569 Metabasites form asymmetric pinch-and-swell boudinage indicating top-to-the-N kinematics.
1570 (e) Typical mylonites from the uppermost parts of Ikaria unit, near Evidilos. Quartz veins have
1571 been transposed and stretched into the shear direction and forming asymmetric sigmoids
1572 indicating top-to-the-NNE kinematics. (f) Composite cross section showing the first-order
1573 strain gradient and the structural position of outcrops of Fig. 6. Trace is represented on Fig. 2.

1574

1575 Figure 7: Physical conditions of the deformation within Ikaria unit

1576 (a) Small-scale top-to-the-S shear bands operating in the lower parts of Ikaria unit. The
1577 volume of rocks involved in the deformation is quite large and the amphibolite-facies
1578 associations are quite well preserved. Note that in these incipiently deformed rocks,
1579 deformation is rather symmetrical and local top-to-the-S kinematics are sometimes observed.
1580 (b) Close-up view of syn-greenschist-facies mylonites from the core of the Agios-Kirykos
1581 shear zone in the upper parts of Ikaria unit.

1582

1583 Figure 8: Deformation in Agios-Kirykos unit

1584 Outcrop pictures of rather weak (a) and intense (b) top-to-the-NE asymmetric deformation in
1585 Agios-Kirykos unit. Note that shearing, developed in the greenschist-facies conditions,
1586 displays a clear evolution to more localized or even cataclastic flow. (c) Large-scale view of
1587 the core of the Fanari shear zone at Agios Kiriaki. Exposed are the mylonites of the
1588 uppermost parts of Ikaria unit, on the first plane, while sediments of Fanari unit crop out in
1589 the background. Close-up views of the deformation of (d) marble and (e) metapelite layers.
1590 Marble layers display successive stages of symmetric boudinage, ranging from ductile to
1591 strictly brittle while metapelite levels show a strong non-coaxial component consistent with
1592 an overall top-to-the-NE kinematics.

1593

1594 Figure 9: The Fanari detachment plane

1595 Large-scale representative view of the Fanari detachment plane at Fanari. The detachment
1596 plane is stripped of its sedimentary cover over several hundreds of square meters. Sediments
1597 that are almost vertical are preserved in the incised gullies and all along the coast from Fanari
1598 to Agios Kirykos. The plane carries series of large-scale corrugations and crescent-shaped
1599 structures indicating a consistent top-to-the-NNE kinematics. Note that the detachment plane
1600 is cut across by series of high-angle faults carrying sub-horizontal striations. Also are
1601 represented stereographic projections of striations and kinematics of both the detachment
1602 plane and the faults that offset the plane.

1603

1604 Figure 10: Using microstructures to unfold the Fanari detachment

1605 (a) Sketch depicting the probable geometry of the Fanari detachment prior to its tilting
1606 together with the whole system (see Fig. 9). The result is a system where the detachment
1607 plane operates at shallow angle with dextral-normal-sense. In the restored position, sediments

1608 appear pervasively affected by a single set of conjugate normal faults. (b) Outcrop pictures
1609 and stereographic projections of fault systems before and after back-tilting rotation for sites 6
1610 and 7 (see Fig. 5 for location). In this restored position, paleostress solutions are all consistent
1611 with a unique and common NNE-SSW to NE-SW extensional stretching. This orientation is
1612 consistent with the paleostress field deduced from other stations throughout the island (see
1613 Fig. 11). Note that, conversely, three successive stress states are required to account for the
1614 heterogeneous fault populations present in their current position.

1615

1616 Figure 11: Small-scale brittle structures

1617 Detailed results of the fault-slip data inversion. Fault planes, associated striae and results of
1618 inversion were plotted using WinTensor software in Schmidt's lower hemisphere equal-area
1619 projection (Delvaux and Sperner, 2003). Also is presented a representative outcrop recognized
1620 as demonstrative of a brittle stage subsequently developed after the ductile one (site 1; see
1621 location on Fig. 5). Faults occur as a dense array of rather steep normal faults. Note the
1622 consistency of the NNE-SSW to NE-SW direction of extension all over the island.

1623

1624 Figure 12: RSCM peak temperature results

1625 (a) Sampling map used for the RSCM study. In the bottom right corner, T_{\max} are sorted by
1626 increasing temperature, where errors brackets correspond to intra-sample heterogeneity
1627 standard deviation. Note that temperature is comparable for two samples, positioned side by
1628 side in the chart, for Ikaria unit on one hand, and for Agios Kirykos unit on the other hand.
1629 But the two units are characterized by clearly differentiable bulk temperature. (b) T_{\max} vs
1630 structural position for the northeastern part (samples used for (b) are written in bold on the
1631 map). Note the gap of temperature near the Agios-Kirykos shear zone, quite clear on the
1632 Therma cross section. (c) Increasing temperature approaching the Raches granite.

1633

1634 Figure 13: U-Th-Pb analyses on monazite from migmatite sample IK01A

1635 (a) Typical textural relationships between monazite (Mnz) crystals and the magmatic
1636 paragenesis as explored by BSE mean. Mnz 3 is included into biotite (Bt). (b) Details of the
1637 internal texture of Mnz 3 and Mnz 4 monazite crystals. BSE image reveal a clear core-rim
1638 textures. Is also shown the location of ICPMS laser ablation analysis (9 μm) and their
1639 corresponding $^{208}\text{Pb}/^{232}\text{Th}$ ages (2σ level). Note the correlation between ages and the core-rim
1640 textures. (c) $^{206}\text{Pb}/^{238}\text{U}$ vs $^{208}\text{Pb}/^{232}\text{Th}$ diagram for all data showing a $^{208}\text{Pb}/^{232}\text{Th}$ age of $15.7 \pm$
1641 0.2 Ma concordant with the $^{206}\text{Pb}/^{238}\text{U}$ ages. (d) Tera-Wasserburg diagram for all analyses,
1642 intercepting the Concordia at 15.1 ± 0.2 Ma. Discordant U/Pb ages suggest a slight common
1643 Pb contamination and the uncertainty of the $^{207}\text{Pb}/^{235}\text{U}$ ages due to the low ^{207}Pb content of
1644 young monazites. In this study, only the $^{208}\text{Pb}/^{232}\text{Th}$ ages were considered (inset c).

1645 Mineral abbreviations are after Whitney and Evans (2010)

1646

1647 Figure 14: Large-scale structure and available time-constraints

1648 (a) Three-dimensions large-scale sketch of Ikaria depicting the relationships between the
1649 structural dome, the synkinematic intrusions and the major shear zones. (b) Time-chart
1650 compiling all geochronological constraints available for Ikaria. Note that both the
1651 metamorphic dome and the late intrusions share a common fast cooling from 11-10 Ma
1652 onward. U/Pb ages on zircon (Zrn) are from Bolhar et al. (2010); K/Ar ages on muscovite
1653 (Ms) and biotite (Bt), Rb/Sr ages on muscovite and FT ages on apatite (Ap) are from Altherr
1654 et al. (1982) and Kumerics et al. (2005). (U-Th)/He ages on apatite and FT ages on zircon are
1655 from Kumerics et al. (2005); K/Ar ages on hornblende (Hbl) are from Altherr et al. (1982).
1656 Mean closure temperature are from Harrison (1981), Steck and Hunziker (1994), Grove and
1657 Harrison (1996), Brandon et al. (1998), Ketcham et al. (1999), Farley (2000), Cherniak and

1658 Watson (2001) and Harrison et al. (2009). K/Ar, Ar/Ar and Rb/Sr ages on metamorphic rocks
1659 could reflect deformation-assisted crystallization (Kumerics et al., 2005) or cooling (Altherr et
1660 al., 1982). Same ages on granites are interpreted as cooling ages (Altherr et al., 1982;
1661 Kumerics et al., 2005). The new age of partial melting from this study is also showed.
1662 Question mark is associated with ages which are discussed in the text.

1663

1664 Figure 15: Large-scale implications

1665 (a) Tectonic map of synorogenic and postorogenic structures in the Aegean domain showing
1666 i) how far the back-arc postorogenic extension remains highly asymmetric in the center of the
1667 domain and ii) the footprint of the central Aegean *HT* zone where MCCs are described. Red
1668 and green arrows indicate the ductile sense of shear associated to the Oligo-Miocene
1669 extensional episode associated with the NCDS, the WCDS and the NDP (Lister et al., 1984;
1670 Faure et al., 1991; Gautier and Brun, 1994; Vandenberg and Lister, 1996; Jolivet and Patriat,
1671 1999; Mehl et al., 2005; 2007; Huet et al., 2009; Grasemann et al., 2012; Augier et al, 2015).
1672 On Syros and Sifnos islands, the synorogenic top-to-the-ENE sense of shear is associated to
1673 the synorogenic Vari detachment of Syros (Trotet et al., 2001a; 2001b). (b) Stretching-parallel
1674 cross-section through the central Aegean *HT* zone where intensity of stretching and
1675 asymmetry of the deformation are maximum.

Figure1
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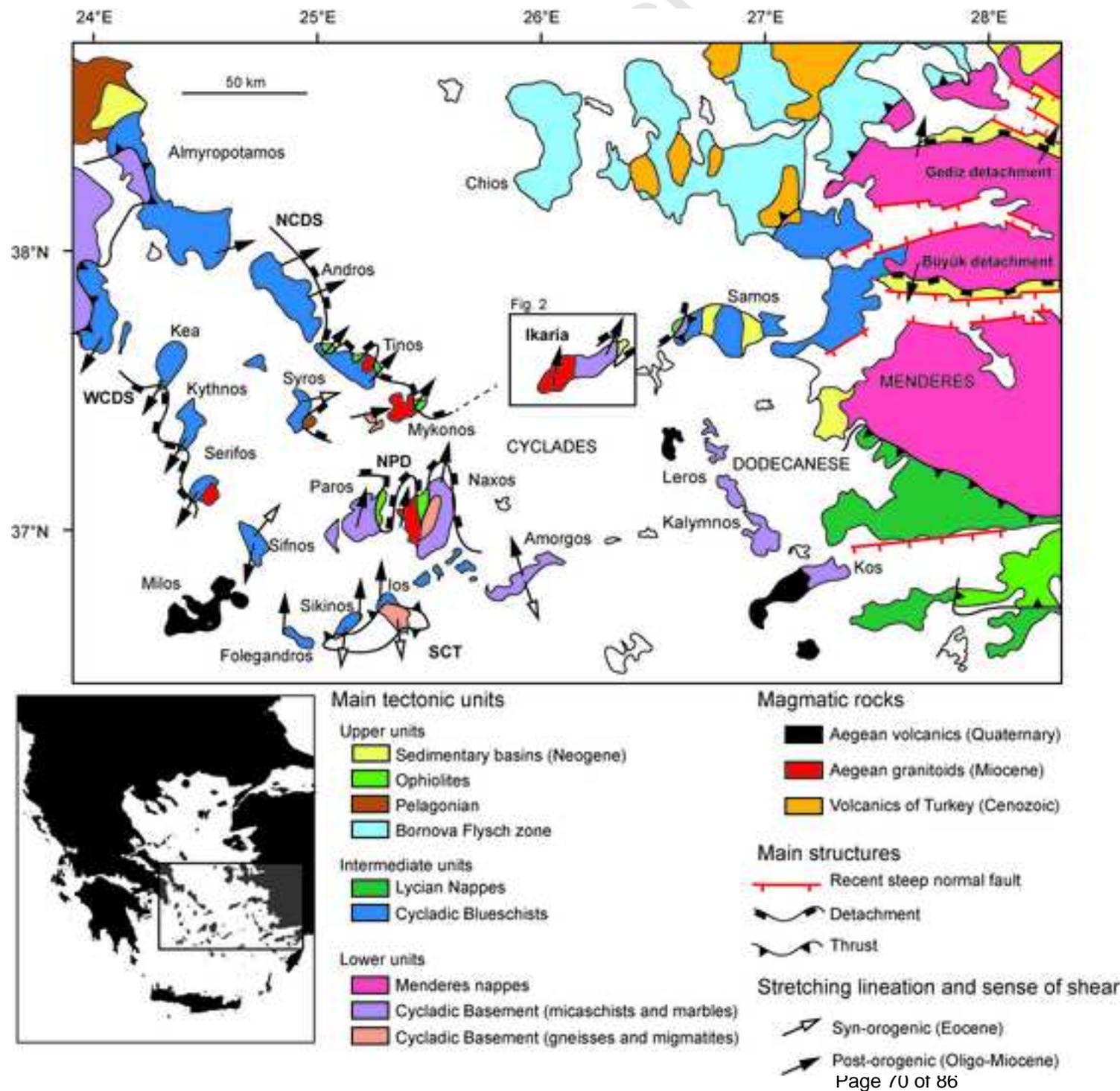
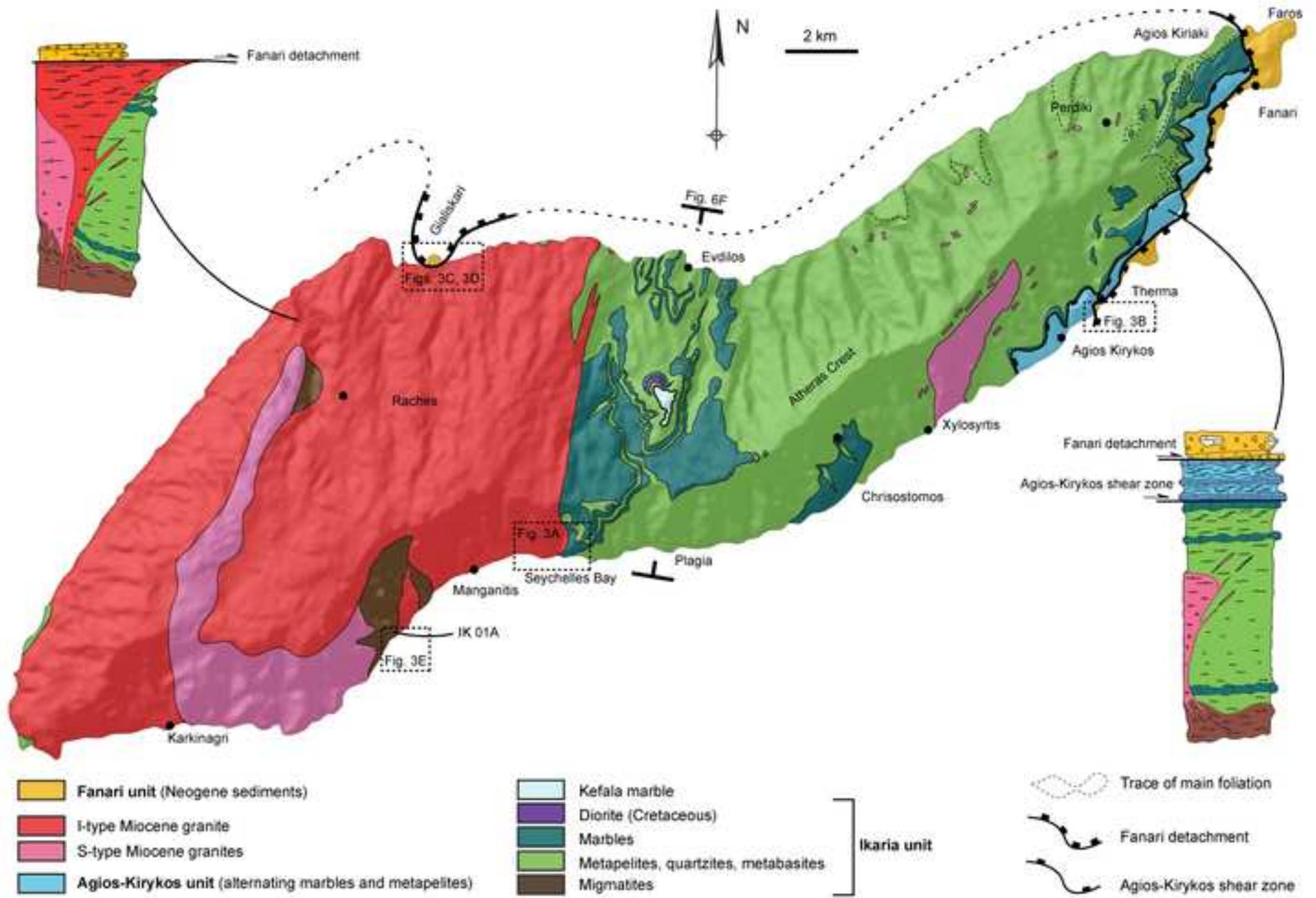


Figure2
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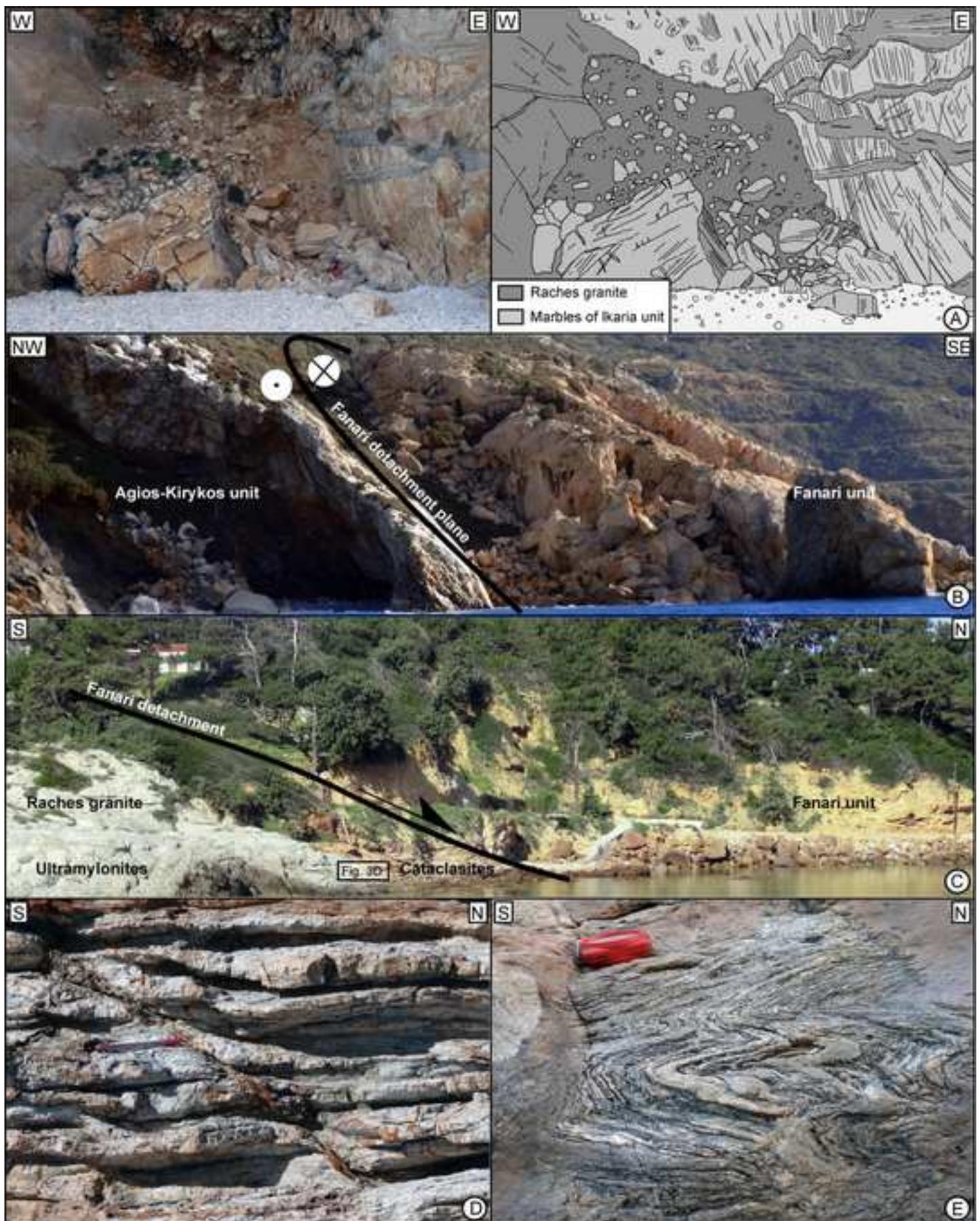


Figure4
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Main planar fabrics

Foliation

- ↖ This study
- ↗ Compilation from Laurent et al. (2015) and Photiadis (2002b)

Bedding in Fanari unit

- ↖ This study

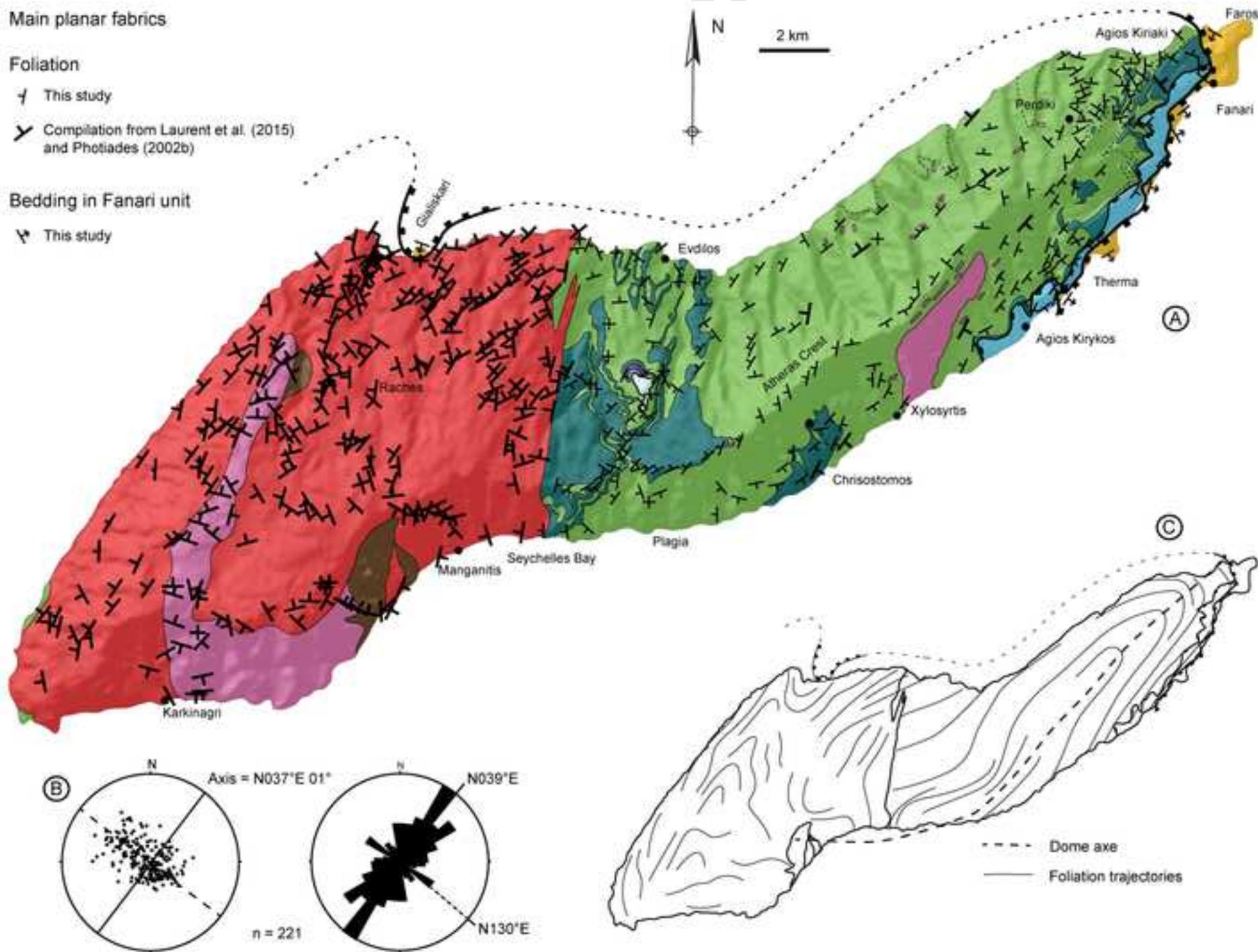
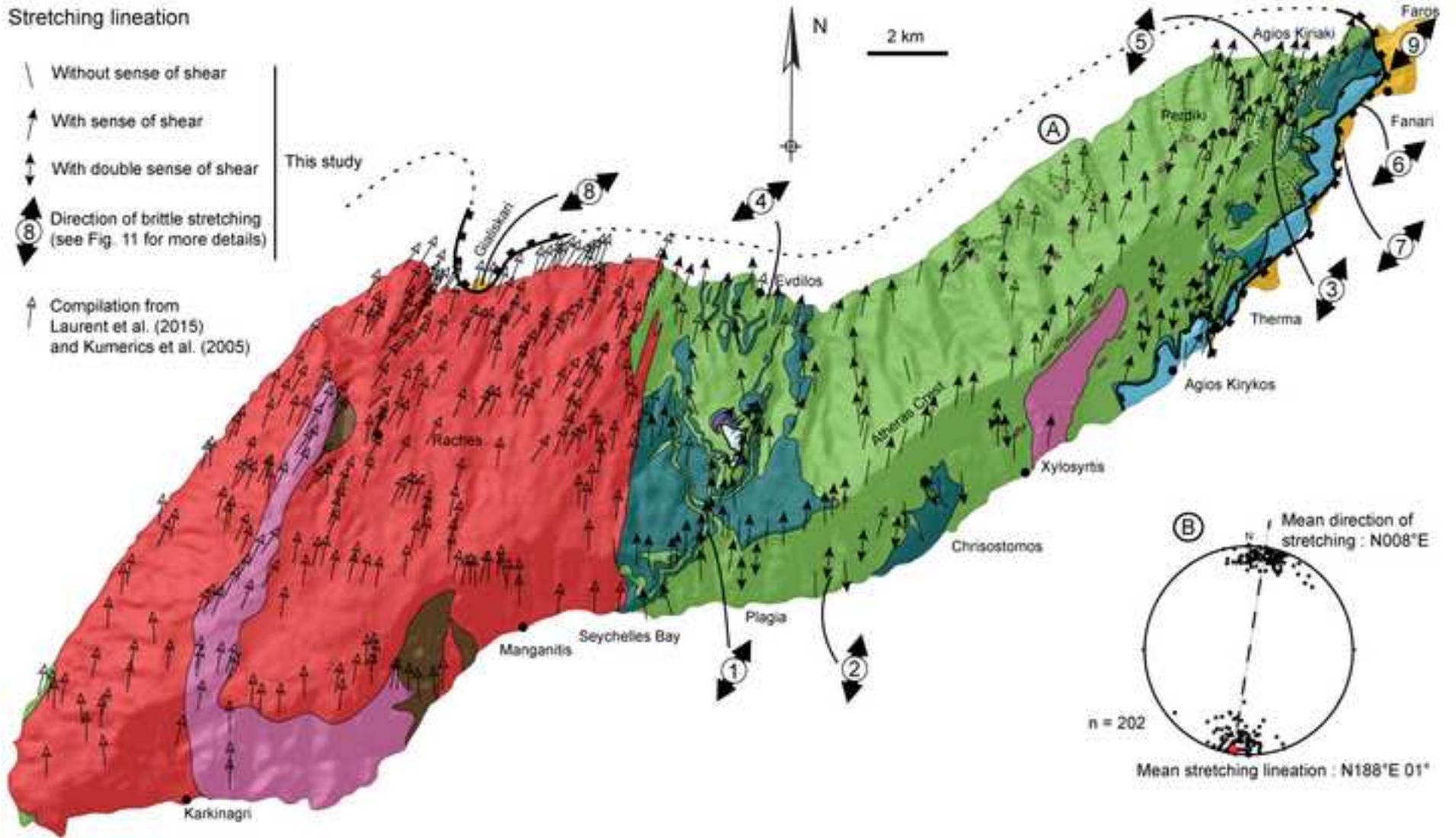


Figure5

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Stretching lineation

-  Without sense of shear
-  With sense of shear
-  With double sense of shear
-  Direction of brittle stretching (see Fig. 11 for more details)
-  Compilation from Laurent et al. (2015) and Kumerics et al. (2005)



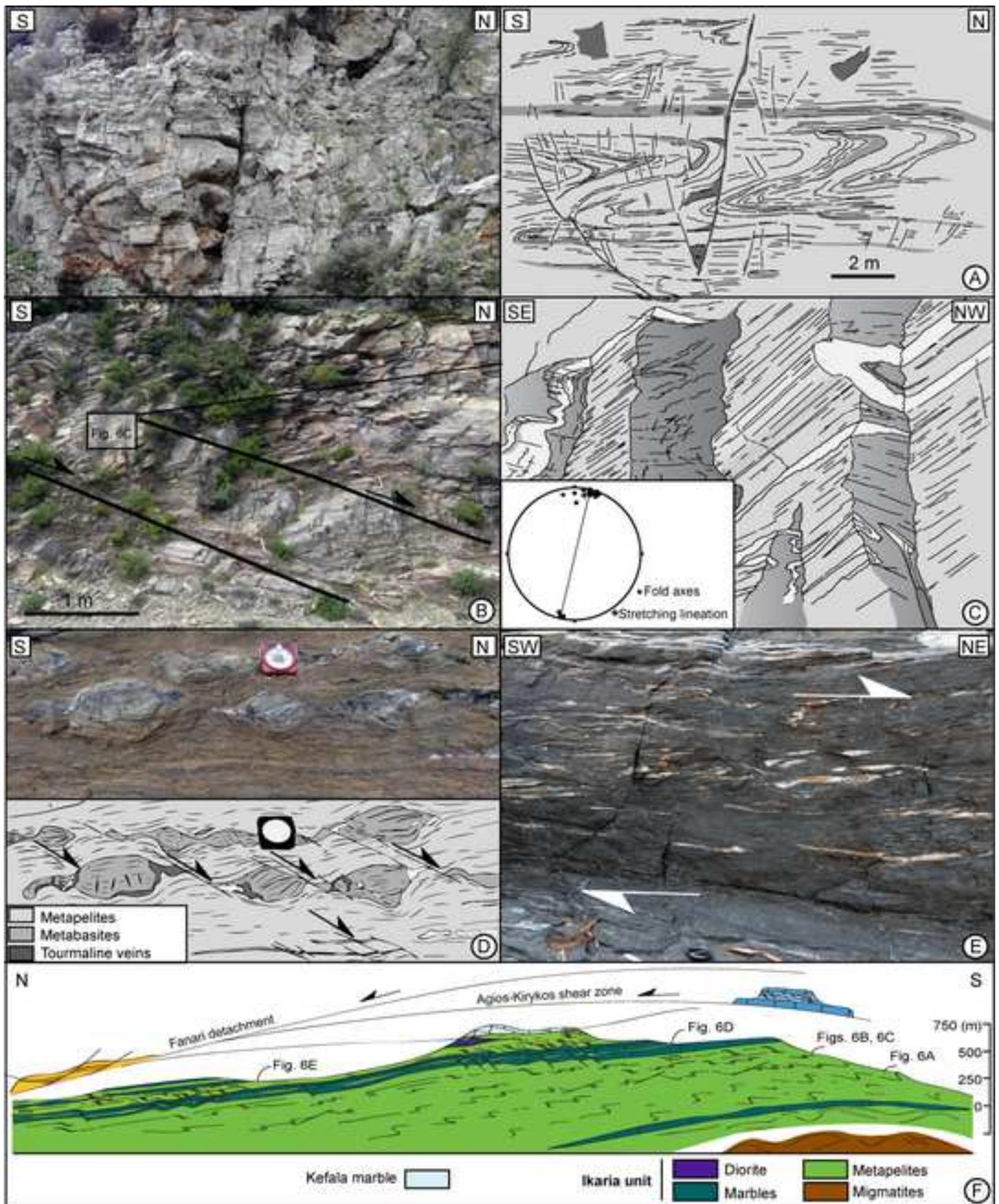
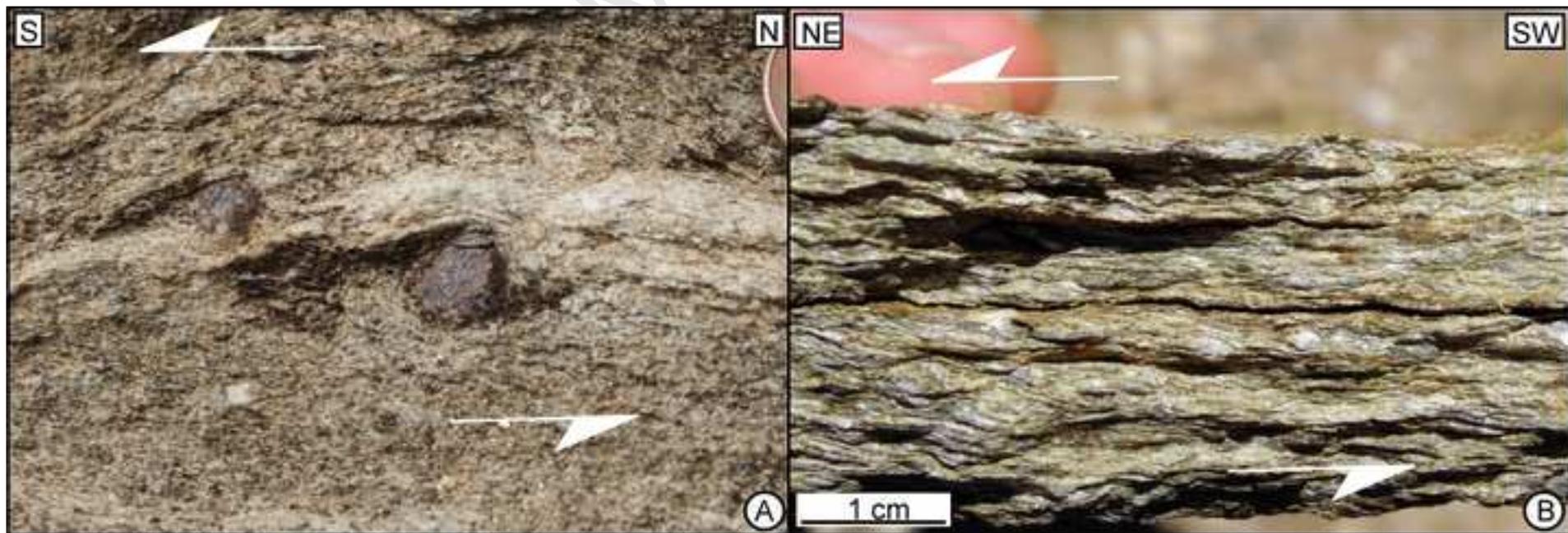


Figure 7
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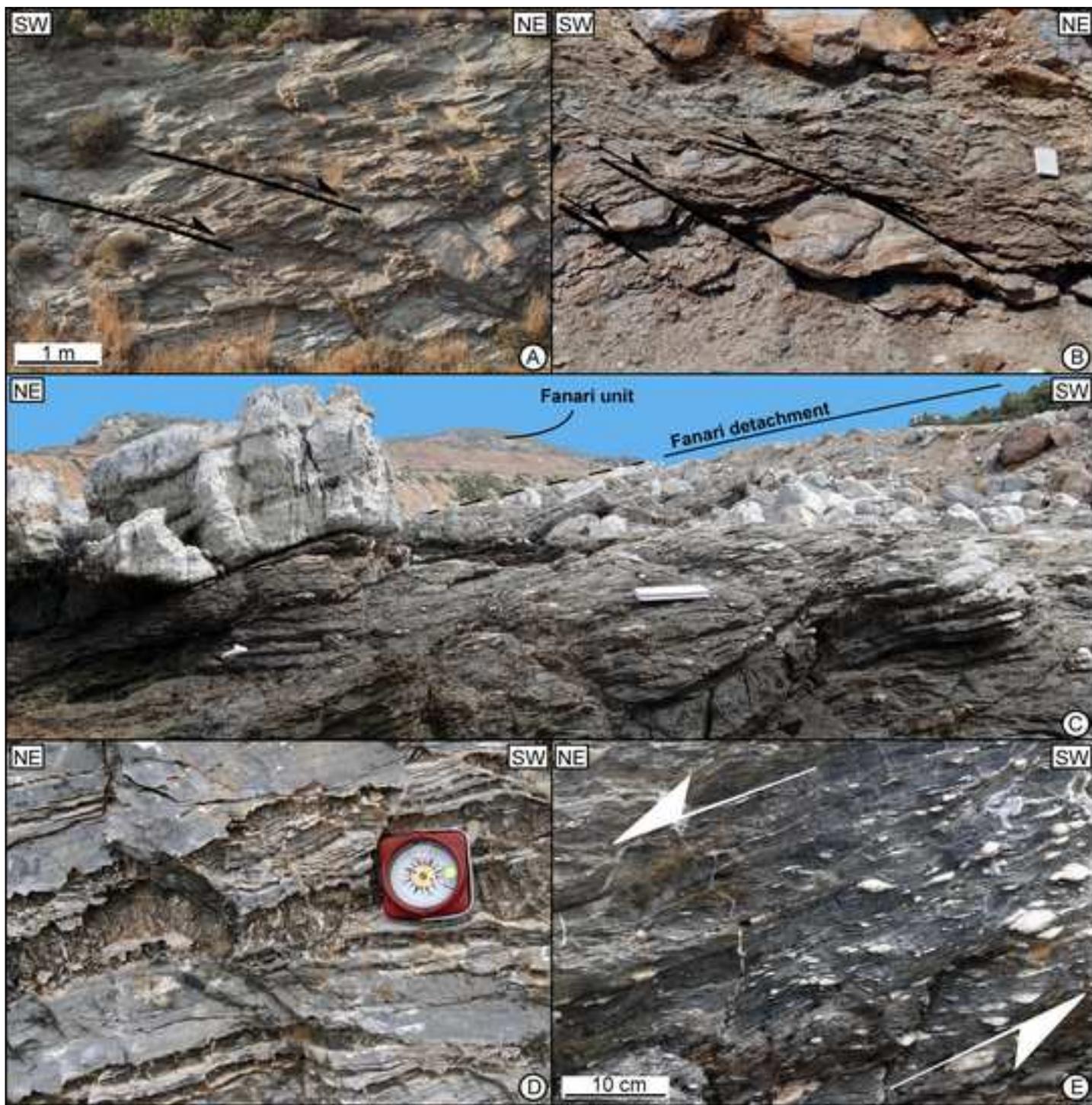


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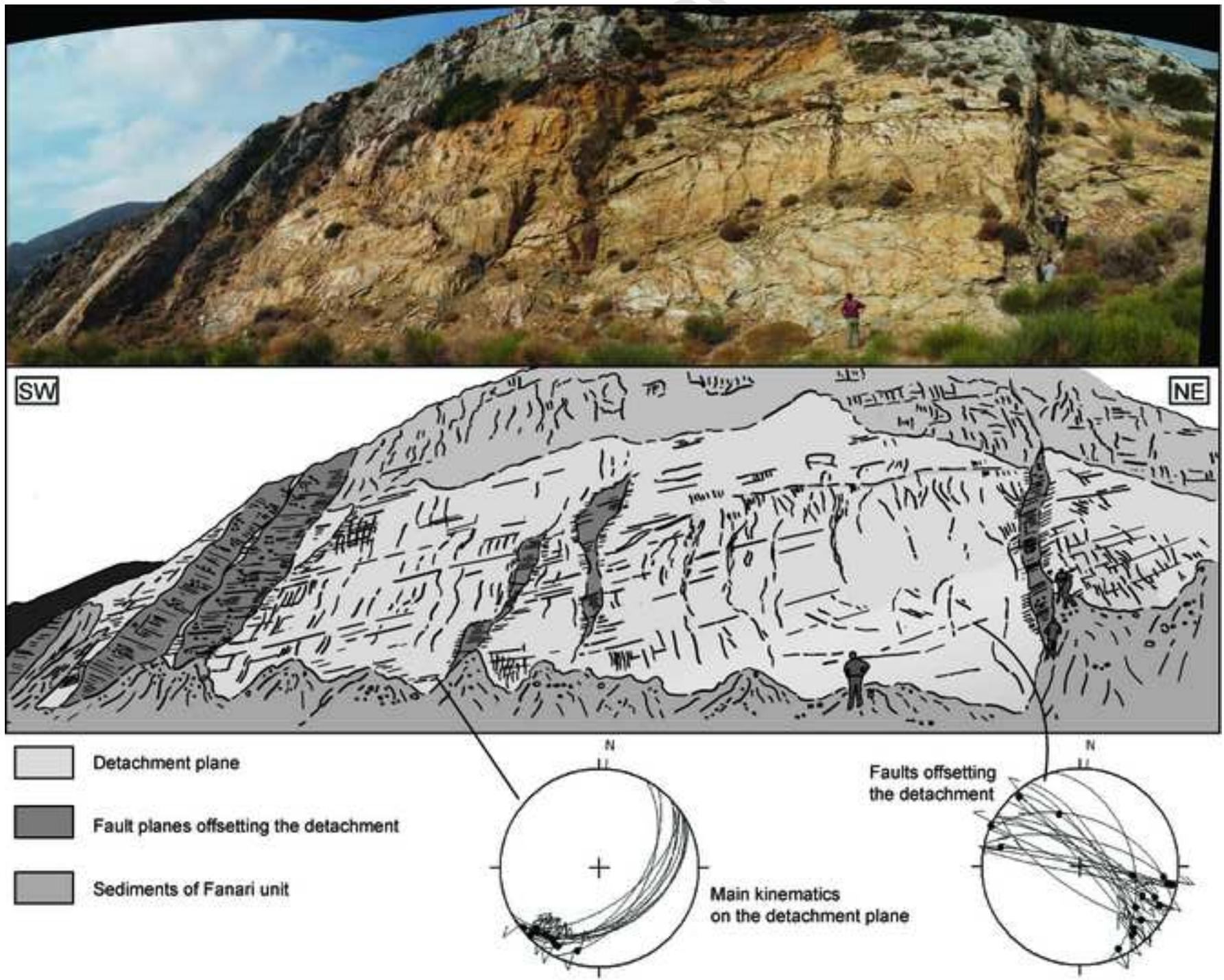
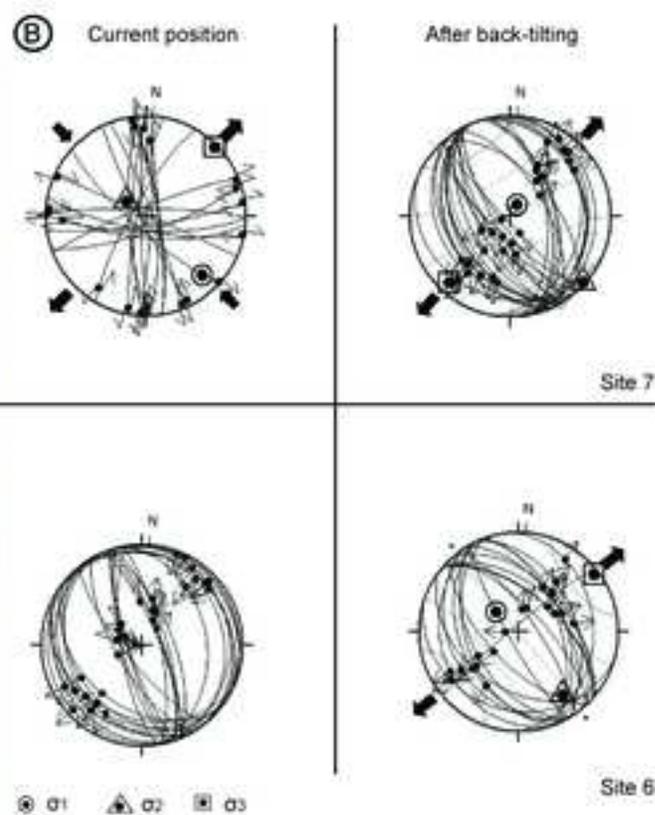
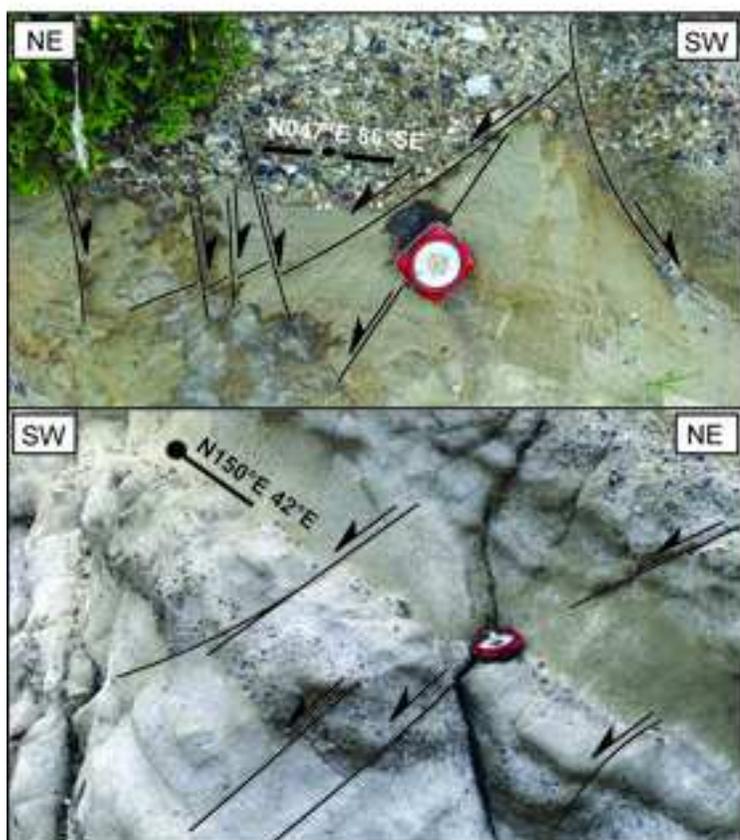
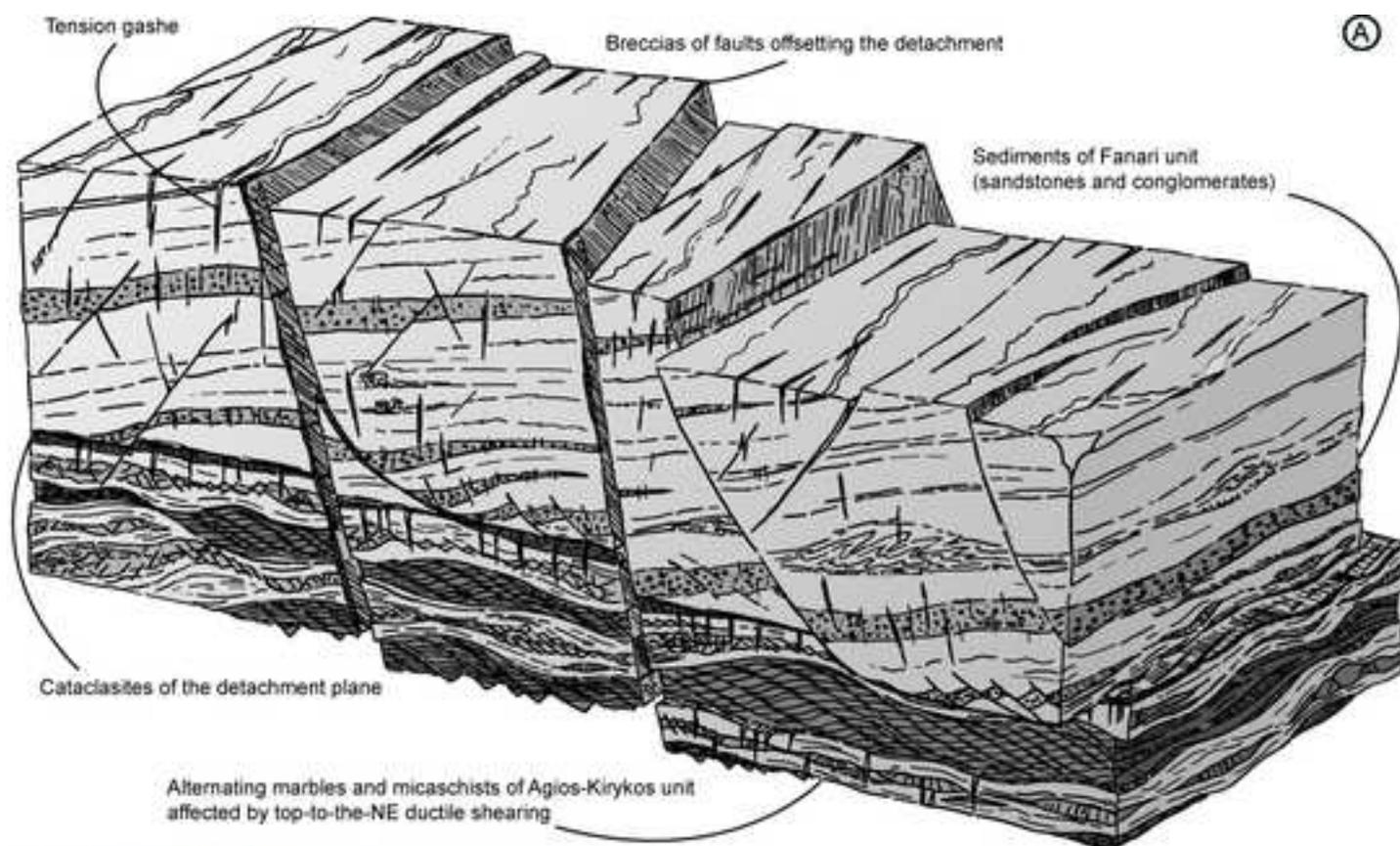
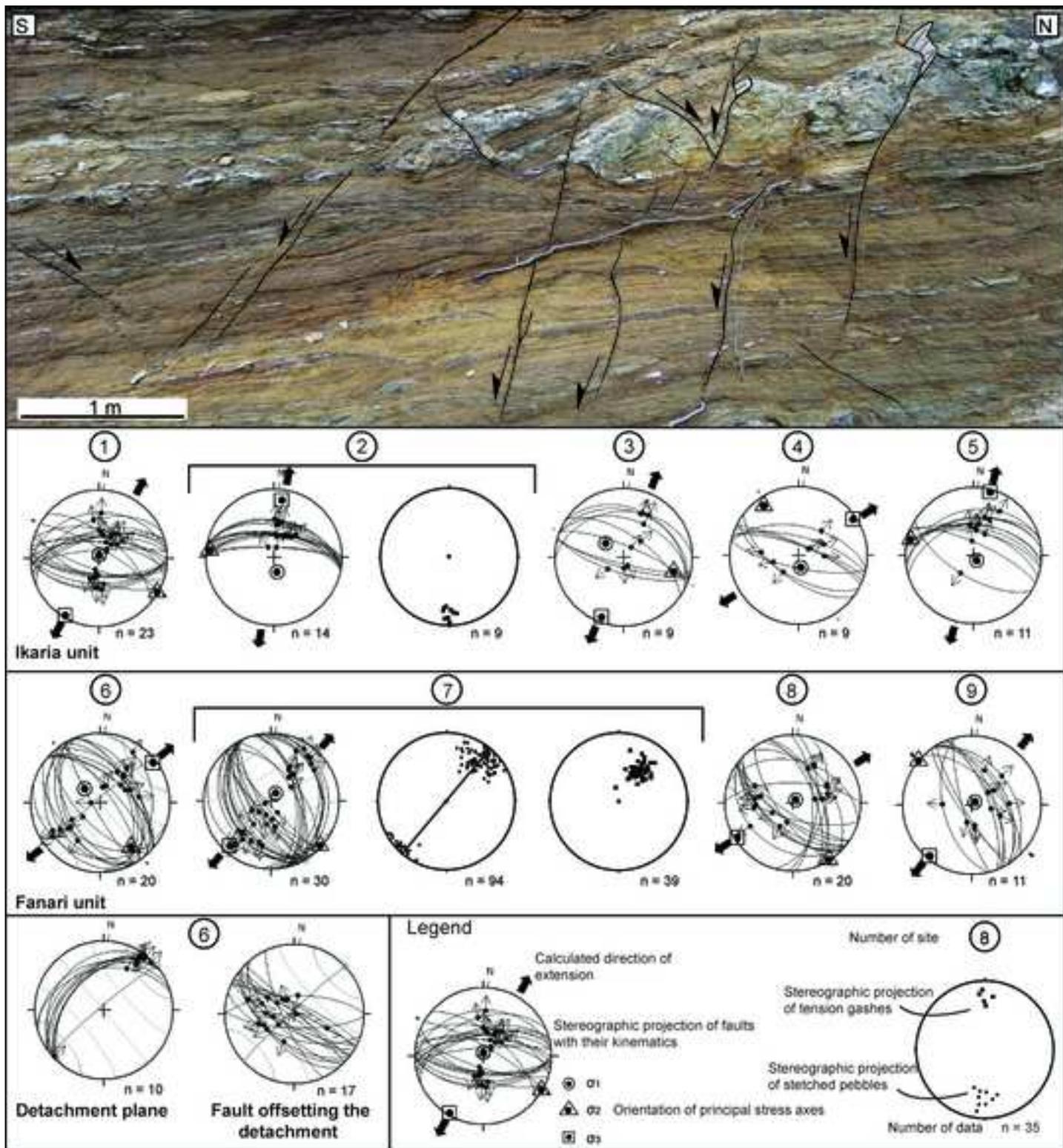


Figure 10

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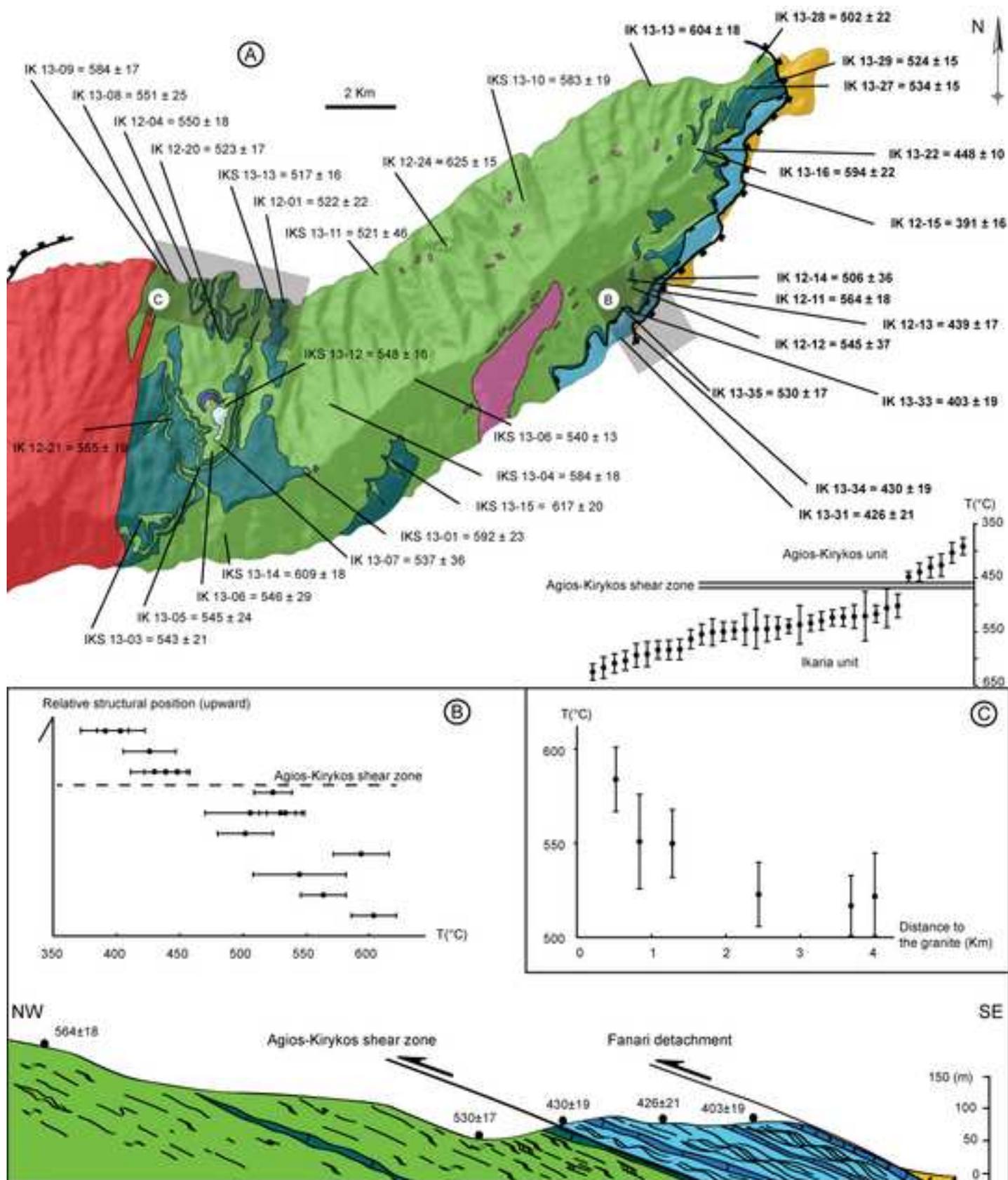
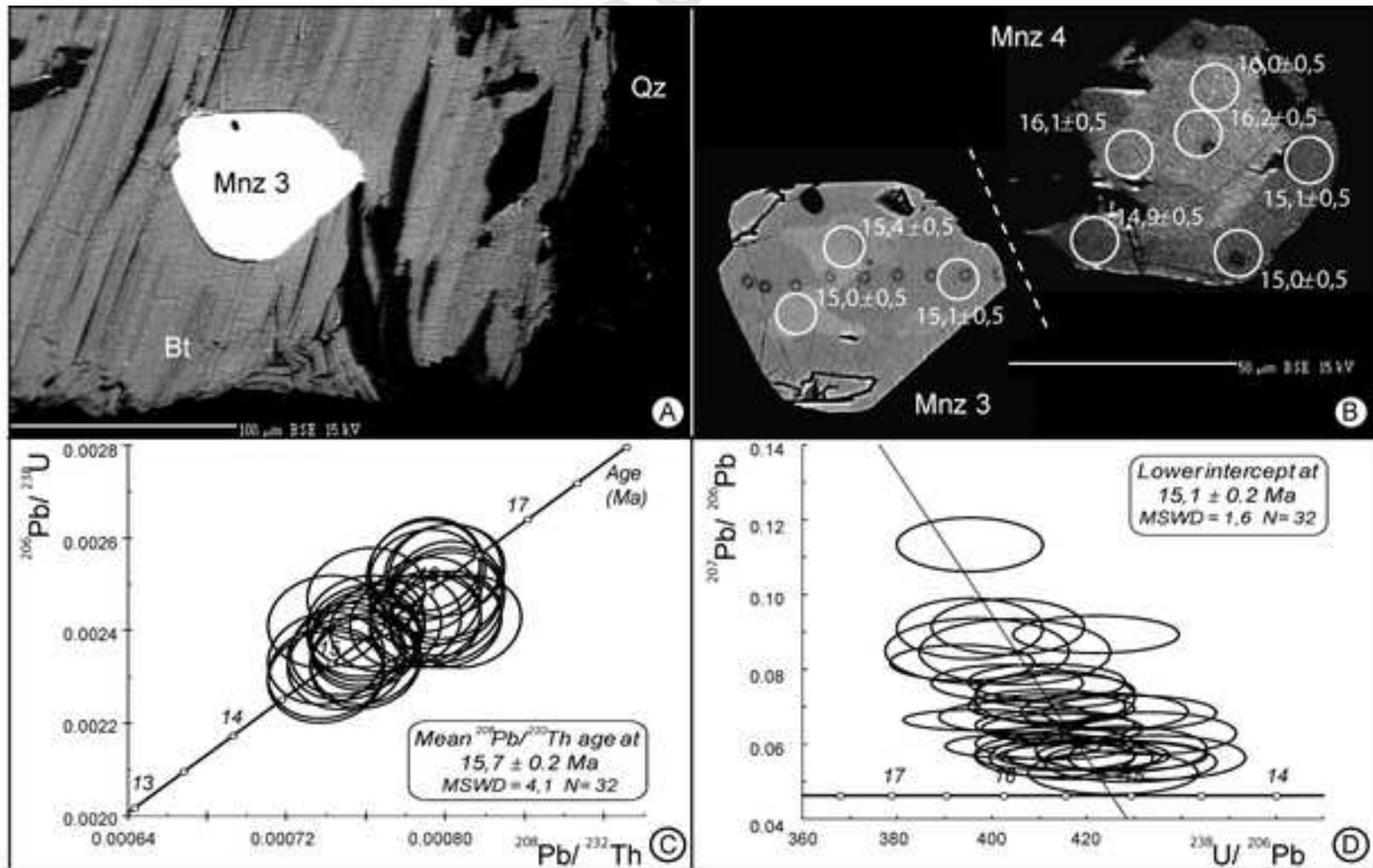


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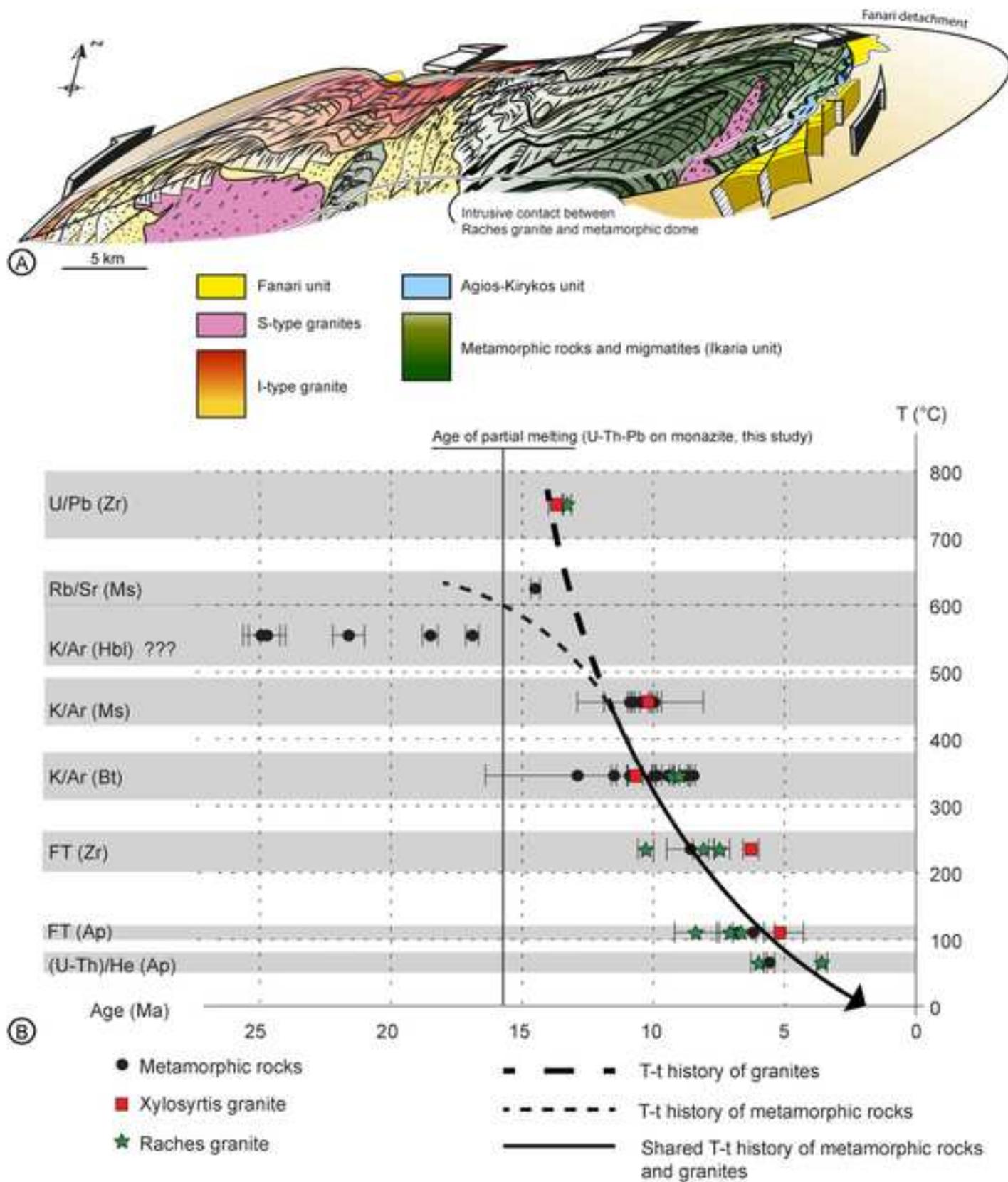
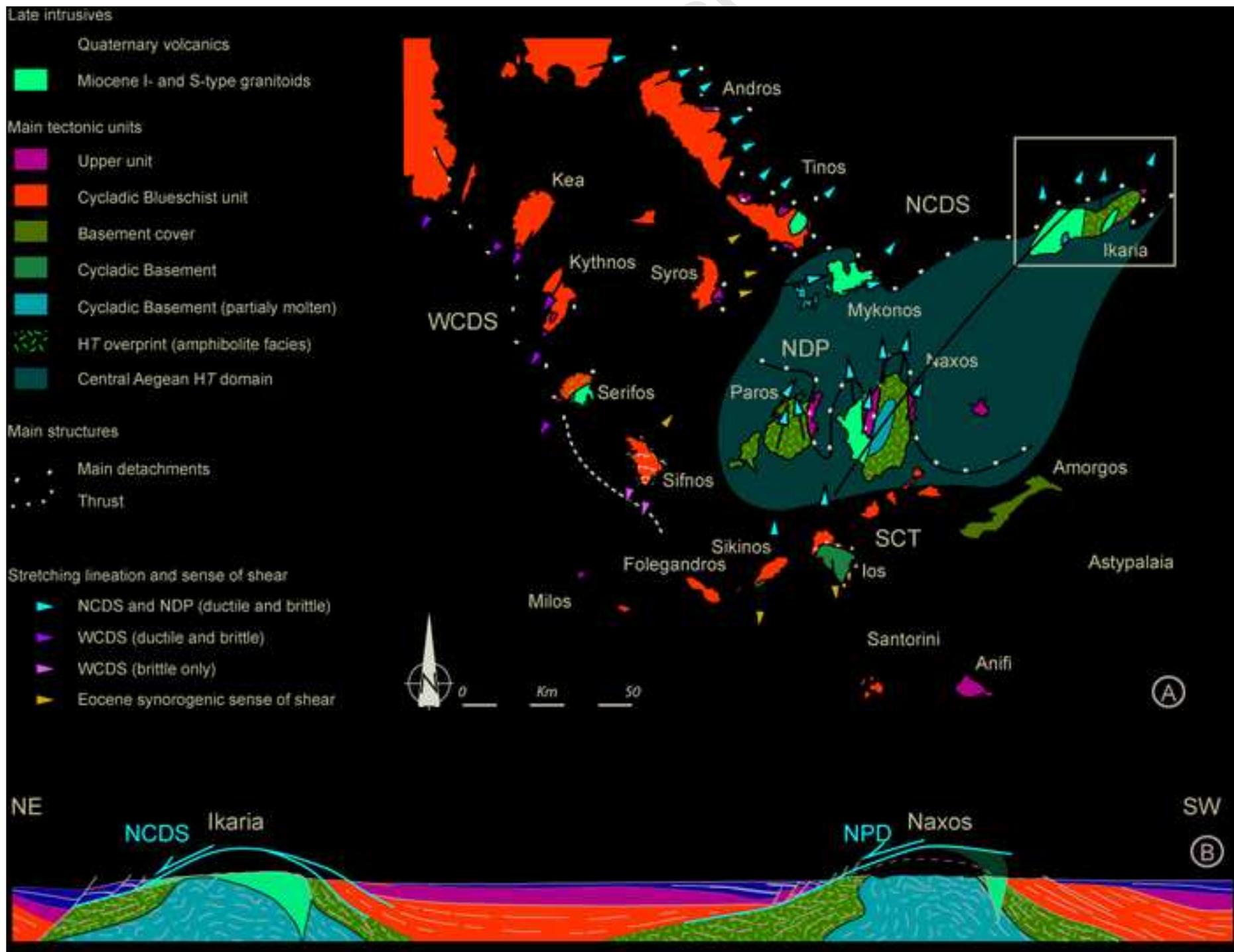


Figure 15
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Sample	Coordinates (UTM 35 N)		n	R2		T (°C)	
				Mean	SD	Mean	SD
Agios-Kirykos unit							
IK 12-15	441548	4168007	21	0,562	0,036	391	16
IK 13-33	438478	4163867	19	0,535	0,043	403	19
IK 13-31	438178	4163625	20	0,482	0,049	426	21
IK 13-34	438453	4164338	19	0,473	0,044	430	19
IK 12-13	439086	4164923	19	0,453	0,037	439	17
IK 13-22	440733	4168835	14	0,433	0,023	448	10
Ikaria unit							
IK 13-28	441898	4171296	19	0,312	0,049	502	22
IK 12-14	439237	4165283	20	0,303	0,082	506	36
IKS 13-13	428570	4164441	17	0,278	0,035	517	16
IKS 13-11	431485	4165492	21	0,269	0,104	521	46
IK 12-01	428758	4164611	23	0,266	0,051	522	22
IK 12-20	427111	4163553	18	0,264	0,038	523	17
IK 13-29	442324	4170727	13	0,262	0,034	524	15
IK 13-35	438248	4164495	15	0,251	0,039	530	17
IK 13-27	441743	4170619	15	0,241	0,033	534	15
IK 13-07	426986	4160696	21	0,233	0,081	537	36
IKS 13-06	432513	4162583	16	0,228	0,029	540	13
IKS 13-03	424903	4158649	16	0,219	0,047	543	21
IK 12-12	438723	4164980	20	0,216	0,082	545	37
IK 13-05	426436	4160041	15	0,216	0,055	545	24
IK 13-06	426795	4160494	16	0,213	0,065	546	29
IKS 13-12	427145	4161730	16	0,209	0,035	548	16
IK 12-04	426177	4165286	16	0,204	0,040	550	18
IK 13-08	425761	4165301	18	0,202	0,057	551	25
IK 12-21	425535	4161591	13	0,194	0,044	555	19
IK 12-11	438363	4165292	19	0,174	0,041	564	18
IKS 13-10	435557	4167534	13	0,130	0,042	583	19
IK 13-09	425573	4165499	17	0,129	0,045	584	20
IKS 13-04	430168	4161819	11	0,129	0,041	584	18
IKS 13-01	429434	4159919	17	0,110	0,051	592	23
IK 13-16	440281	4168893	18	0,107	0,051	594	22
IK 13-13	439031	4170692	15	0,083	0,040	604	18
IKS 13-14	427196	4158256	13	0,072	0,041	609	18
IKS 13-15	431859	4160264	14	0,054	0,045	617	20
IK 12-24	433322	4166192	18	0,037	0,034	625	15

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