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New insights on the fossil arc of the Tyrrhenian Back-Arc Basin (Mediterranean Sea)

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1 **New insights on the fossil arc of the Tyrrhenian Back-Arc Basin**
2 **(Mediterranean Sea)**

3

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18

19 **Abstract**

20 Geology, geophysics and geodynamics of the Tyrrhenian Back-Arc Basin (BAB; central
21 Mediterranean Sea) have been studied extensively during the last 50 years. However, some

22 topics are still open: for example, the possible migration of the volcanic arc during the

23 Ionian subduction of the past few Ma. We improved our knowledge of the geodynamics of

24 the Tyrrhenian BAB in the area South of the Vavilov Volcano by analyzing multibeam

25 bathymetry and unpublished single-channel reflection seismic and magnetic data.

26 Furthermore, we studied the petrology of igneous rocks as well as facies and microfaunas of

27 carbonates dredged from the Aurelia and Augusto seamounts. The Aurelia basement is made
28 of basalts with calc-alkaline affinity. Carbonates from the Aurelia and Augusto seamounts
29 consist of cemented Mg-calcite biomicrite crusts rich in planktonic foraminifera not older
30 than Early Pleistocene. Based on our results, we interpret the Augusto and Aurelia
31 seamounts as part of the active volcanic arc seaward of the Tyrrhenian BAB in Late
32 Pliocene–Early Pleistocene.

33

34 Keywords:

35 Back-Arc Basins Tyrrhenian Sea Volcanic Arcs Geodynamics Transfer Zones

36

37 **Introduction**

38 Back-Arc Basins (BABs) and volcanic arcs are two main features characterizing the upper
39 plates along convergent plate boundaries (Uyeda and Kanamori, 1979; Leat and Larter,
40 2003). The relative kinematics, composition and thermal state of the upper and lower plates,
41 together with the age of the subducting lithosphere and the morphotectonic inheritance of the
42 upper plate, rule the extensional tectonics along the BABs and their progressive evolution
43 from a younger rifting stage to a mature spreading stage (e.g., Parson and Wright, 1996;
44 Fujiwara et al., 2001; Martinez and Taylor, 2002; Sdrolias and Muller, 2006; Weins et al.,
45 2006; Schellart et al., 2007). BABs in a rifting stage do not show a BAB magmatism, as for
46 example the Havre Trough in the Southern Pacific, characterized by an oblique southward
47 propagating extension and by the absence of a clear spreading ridge (Caratori Tontini et al.,
48 2019); mature BABs show a Back-Arc Spreading Center (BASC), i.e. the Mariana
49 Spreading Center in the Pacific (Hynes and Mott, 1985), or the East Scotia Ridge in the
50 Southern Atlantic (Livermore et al., 1997; Fretzdorff et al., 2002).

51 Here we focus on the Tyrrhenian Sea (located in the central Mediterranean Sea, between the
52 Italian Peninsula and the Sardinia; Fig. 1), a peculiar case of BAB associated to

53 compressional tectonics where the lower oceanic plate is subducting under continental
54 lithosphere. The Tyrrhenian is a BAB formed by extensional tectonics due to the progressive
55 eastward/south-estward retreat of the Ionian subduction (e.g., Malinverno and Ryan, 1986;
56 Doglioni, 1991; Doglioni et al., 1999, 2004; Faccenna et al., 1997, 2001; Carminati et al.,
57 1998; Sartori, 2003; Rosenbaum et al., 2008; Conti et al., 2017; Loreto et al., 2020). First
58 studies on the regional geology and geodynamics of the Tyrrhenian were published during
59 1970s and 1980s (e.g., Barberi et al., 1973, 1978; Selli et al., 1977; Hsü et al., 1978; Wezel,
60 1982; Della Vedova et al., 1984; Sartori, 1986; Rehault et al., 1987; Trincardi and Zitellini,
61 1987; Savelli, 1988). An important contribution on the knowledge of the Tyrrhenian basin
62 occurred with the Deep Sea Drilling Project (DSDP) Leg 42, the Ocean Drilling Program
63 (ODP) Leg 107 (Kastens et al., 1988; Kastens and Mascle, 1990). A collection of
64 multidisciplinary papers on the geology and geodynamics of the Tyrrhenian Sea was shown
65 by Marani et al. (2004), after the deep seismic exploration of the central Mediterranean and
66 Italy (CROsta Profonda project; Scrocca et al., 2003; Finetti, 2005) during the 1990s.
67 Recently, seismic refraction data have been acquired during the MEDOC Cruises in the 2010
68 (Ranero et al., 2012) in order to display the velocity structure of the Tyrrhenian crust and
69 uppermost mantle together with the Moho reflector geometry (Prada et al., 2014, 2016,
70 2018).

71 The Tyrrhenian Abyssal Plain (TAP), marked by the isobaths of the 3000 m in Fig. 1a, is
72 floored by basaltic and ultramafic rocks covered by Pliocene-Quaternary sediments (Hsü et
73 al., 1978; Kastens and Mascle, 1990). The TAP shows three huge fissural volcanoes (the
74 Magnaghi, the Vavilov and the Marsili) located in the center of three different basins (Fig.
75 1a). The Marsili and the Aeolian Islands represent respectively the current magmatism in the
76 back-arc basin and in the arc front (Fig. 1; Kastens et al., 1988; Kastens and Mascle, 1990;
77 Faggion et al., 1999; Marani and Trua, 2002; Trua et al., 2002, 2018; Marani et al., 2004;
78 Nicolosi et al., 2006; Rosenbaum et al., 2008; Cocchi et al., 2009; Ventura et al., 2013). The

79 Magnaghi and the Vavilov can be considered segments of extinct back-arc spreading centres
80 evolved naturally in a basin characterized by frequent spreading jumps (Magni et al., 2021;
81 Schliffke et al., 2022). In this paper we focus on the region East of the Magnaghi and South
82 of the Vavilov volcanoes (Fig. 1a). This area shows an alternation of deep basins and high
83 seamounts from which we have little understanding of their age and composition, given the
84 lack of data. Some information can be extracted from studies of regional geology (Marani et
85 al., 2004; Marani and Gamberi, 2004; Rovere and Wurtz, 2015; Palmiotto and Loreto, 2019;
86 Pensa et al., 2019), the lithological and stratigraphic map of the Italian Seas (Colantoni et al.,
87 1981), multichannel seismic reflection profiles (Finetti and Del Ben, 1986; Corradino et al.,
88 2022) and the distribution of the regional magnetic data (Cella et al., 1998; Florio et al.,
89 2022). In particular, we study two different seamounts (Aurelia and Augusto) in order to
90 investigate their origin and to improve the geology and geodynamics of the central
91 Tyrrhenian BAB.

92 We carried out a geophysical study based on: 1) multibeam bathymetry data downloaded
93 from the European Marine Observation and Data Network (EMODnet;
94 <http://doi.org/10.12770/c7b53704-999d-4721-b1a3-4ec60c87238>); 2) single-channel
95 reflection seismics collected by the Institute of Marine Sciences (ISMAR) of the National
96 Research Council (CNR) of Bologna in the 1970s (Fabbri et al., 1981; see Fig. 1a and
97 methods); 3) magnetic data collected by the Institute of Marine Sciences (ISMAR) of the
98 National Research Council (CNR) of Bologna in the 1990s (Bortoluzzi et al., 1999; see Fig.
99 1b and methods). We created regional bathymetry maps and an updated map of the reduced
100 to pole magnetic anomalies of the central / Southern Tyrrhenian using magnetic data.

101 Furthermore, we analyzed for the first time the petrology of igneous rocks and re-analyzed
102 facies and microfauna of carbonates dredged at two seamounts South of the Vavilov
103 Volcano. Results reveal new insights on the geodynamics of the Tyrrhenian BAB during the
104 Late Pliocene-Early Pleistocene.

105

106 **Material and methods**

107 *Bathymetry*

108 Middle resolution bathymetric data (200 m-cell grid size) used in this paper have been
109 downloaded from the European Marine Observation and Data Network (EMODnet;
110 <http://doi.org/10.12770/c7b53704-999d-4721-b1a3-4ec60c87238>). Spatial analysis and
111 mapping of ASCII data used the open source software GMT (Wessel and Smith, 1998) with
112 the nearest neighbour algorithm. Datum and projection used are, respectively, WGS84 and
113 World Mercator. The Global Mapper Software has been used to create 2D digital elevation
114 images.

115

116 *Magnetic data*

117 Magnetic data were collected by the CNR-ISMAR during the TIR-96 cruise onboard the
118 R/V Gelendzhik in the 1996 and the TIR-99 cruise onboard the R/V A.N. Strakhov in the
119 1999 (Bortoluzzi et al., 1999). In total, >25,000 magnetic measurements along 1400 km of
120 lines NNESSW oriented were used to create a new regional map (Fig. 1b). Raw data were
121 corrected for spikes and diurnal variations using the reference station of L'Aquila (Central
122 Italy). Magnetic anomalies were calculated by subtracting the IGRF (International
123 Geomagnetic Reference field) model and then reducing the data to the North Pole by phase
124 shifting them using the regional inclination and declination values of the IGRF.

125

126 *Seismics*

127 Seismic profiles used in this paper are part of an old large (about 46,000 km) dataset,
128 available as profiles printed on paper, collected during several cruises carried out by CNR-
129 ISMAR of Bologna between the 1970s and the 1980s (Fabbri et al., 1981). Seismic data
130 have been shot using a Sparker 30 KJ and recorded with a trace length of 8 s (TWT). A new

131 digital seismic database is under construction at the CNR-ISMAR of Bologna in order to
132 preserve these data, inspired by FAIR (findable, accessible, interoperable and reusable)
133 principles (Wilkinson et al., 2016). Seismic lines were scanned from paper to high resolution
134 raster image (TIFF); they will then be converted into georeferenced SEG-Y format using the
135 free Matlab program IMAGE2SEGY (Farran, 2008) distributed by the Department of
136 Marine Geosciences of the Spanish National Research Council
137 (<http://www.icm.csic.es/gma/en/content/image2segy/>).

138 Here we present parts of five profiles acquired during oceanographic cruises T71, T73 and
139 T75 (yellow, green and pink lines respectively and shown in Fig. 1a) onboard of the R/V
140 Bannock carried out by the “Bacini Sedimentari” Group on behalf of the “Progetto
141 Finalizzato Oceanografia e Fondi Marini” of the Italian CNR in the Tyrrhenian Basin
142 (Fabbri et al., 1981).

143

144 *Samples*

145 Analyzed rocks were dredged by the CNR-ISMAR of Bologna during cruise T75 in the
146 Tyrrhenian Sea. Samples labels refer to cruise, dredge station and rock samples numbers; for
147 example, 75–30-3 refers to rock sample number 3, dredge station 30, carried out during the
148 oceanographic expedition T75.

149 Samples dredged from site 75–30 (orange dot in Fig. 2a) are carbonates and volcanic rocks:
150 volcanic samples have been re-analyzed; carbonates have been analyzed for the first time.

151 Samples dredged from sites 75–35 (yellow dot in Fig. 2a) and 75–36 (green dot in Fig. 2a),
152 composed only of limestones, have been re-analyzed in order to determine the
153 micropaleontological identifications and carbonate facies. Results of past samples analysis
154 are shown in the lithological and stratigraphic map of the Italian Seas (Colantoni et al.,
155 1981).

156 Rock samples underwent macroscopic and thin section examination under the petrographic

157 and binocular microscopes. Bulk-rock abundances of major and minor elements were
158 determined by X-ray fluorescence (XRF). Micropaleontological identifications were based
159 on a bio and chronostratigraphical scheme for the Mediterranean of Iaccarino et al. (2007).
160 Digital photographs and Energy Dispersive Spectroscopy (EDS) for elemental composition
161 were obtained with a scanning electron microscope (FEI QUANTA 400) at the SEM
162 Laboratory of Earth Sciences Department – Sapienza University of Rome (Italy). For the
163 identification of crystalline material X-ray diffraction on sample powder were also carried
164 out by a Phillips PANalytical X'Pert PRO diffractometer using CuK α radiation ($n = 1.5418$
165 \AA), operating at 40 kV and 40 mA at a step size of 0.0260° at the Department of Earth
166 Science, Sapienza University of Rome (Italy). The program used for qualitative analyses is
167 WinPLOTR Programme (CDIFX UMR6226 Rennes/ILL Grenoble).

168

169 **Results**

170 *Geophysics*

171 Bathymetry and the reduced to the pole magnetic anomalies of the Southern Vavilov region
172 maps are shown in Fig. 2. Part of sparker profiles here interpreted, with their values of
173 magnetic anomalies associated, are shown in Fig. 3. Based on Loreto et al. (2020), three
174 main seismic units have been identified: 1) a well-stratified unit (green color in Figs. 3 and
175 6), interpreted as Pliocene-Quaternary (PQ) deposits, based on lithostratigraphic information
176 (Kastens et al., 1988); 2) a poorly-stratified and transparent unit (violet in color in Figs. 3
177 and 6), interpreted as coexisting sediments and volcanic layers; 3) a more chaotic and less
178 reflective unit, interpreted as the basement (brown color in Figs. 3 and 6). Because of the
179 low resolution of the Sparker profiles and their smaller size in Fig. 3, we created
180 supplementary figures in order to zoom them and increase their resolution.

181 The Seamounts D'Ancona I and II, Plinia, Vavilov and Tibullo are located in the Northern
182 part of the region, from West to East (Fig. 2). The D'Ancona is formed by two different

183 seamounts located in the Eastern Magnaghi Abyssal Plain (EMAP; Fig. 2a). The D'Ancona
184 I, a NNW-SSE oriented ridge, is located between 3485 m and 2696 m of depth; it is 19 km
185 long and 12 km wide, with a steep western flank (20°) and an eastern flank 10° steep (Fig.
186 2b). The D'Ancona II is a 13 km long and 11 km wide seamount, E-W oriented, with a depth
187 ranges from 3476 m to 2900 m (Fig. 2a). The northern flank is 12° steep; the southern is 7°
188 (Fig. 2b). From a magnetic viewpoint, the D'Ancona I shows a negative magnetic anomaly
189 (from 0 to -50 nT), whereas the D'Ancona II a positive magnetic anomaly (> 100 nT). East
190 of the D'Ancona II, there is a NNE-SSW oriented topographic ridge 17 km long and 6.5 km
191 wide; it is unnamed in literature and here we called it "Plinia". This seamount ranges
192 between 3150 m and 2647 m of depth (Fig. 2a), and it is characterized by a western flank
193 steeper than the eastern side (20° and 15° respectively; Fig. 2b); the magnetic anomaly is
194 negative and it ranges from 0 to -150 nT. The Southern part of the Plinia can be shown in the
195 in the PM3E sparker profile of Fig. 3 (Box A and A') and Supp. Fig. 1, between the Fix 26
196 and 28, covered by a thin PQ unit. Plinia is located near Vavilov: they show a similar trend
197 and value of magnetic anomaly (Fig. 2).

198 The Vavilov Seamount is located east of the Plinia Seamount. This huge submarine volcano
199 rises from the abyssal plain at a depth of 3500 and arrives at only 793 m below sea level.

200 The Vavilov shows an asymmetric perpendicular profile, with the western flank steeper than
201 the eastern side (23° and 14° respectively; see the slope shader map of Fig. 2b). The depth of
202 the volcano ranges from 3600 m to 823 m. The Vavilov shows a very strong negative
203 magnetic anomaly (from 0 to -681 nT), although a small portion of its eastern flank shows a
204 positive anomaly (from 0 to 150 nT). To the east of Vavilov, there is Tibullo Seamount, an
205 elongated narrow NNE-SSW ridge, only 400 m high, characterized by a symmetric profile
206 (flanks $\sim 13^\circ$ steep); the magnetic anomaly is positive (between 0 and 50 nT) on the northern
207 part, and negative on the southern part (between 0 and 50 nT).

208 The Southern Vavilov Abyssal Plain (SVAP; mean depth ~ 3600 m) is filled by >500 m of

209 sediments (PQ unit along the PM3E profile; Fig. 3 and Supp. Fig. 1). East of the SVAP,
210 there is a seamount unnamed in literature, here called “Aurelia” (Fig. 2). The Aurelia is a
211 WNW/ESE oriented seamount between 3500 m and 2750 m deep, characterized by a strongly
212 asymmetrical perpendicular profile, with the northern flank steeper than the southern flank
213 (25° and 12° respectively; Fig. 2b). The magnetic anomaly ranges from 0 to 50 nT in the
214 western part and from 0 to -50 nT in the eastern part of this seamount. The profile PM4 (Fig.
215 3, Box B and B' and Supp. Fig. 2) crosses perpendicularly the Aurelia, showing between Fix
216 75 and 76 a sub-vertical discontinuity that connects laterally the basement with the PQ unit.
217 The Aurelia is crossed also by Profile PM11 (Fig. 3, Box C and C' and Supp. Fig. 3): also
218 here the asymmetry of the seamount, with a sub-vertical fault affecting the northern flank, is
219 clearly visible. South of the Aurelia, the NNE-SSW oriented ridge Virgilio is formed by two
220 different highs: the northern high (~ 2800 m of depth) with a shallow negative magnetic
221 anomaly (from 0 to -50 nT); in contrast, the southern high (~ 2700 m of depth) with very
222 strong positive values (> 100 nT).

223 The southern part of this region is characterized by two arc-shaped seamounts. One of those,
224 the Augusto, a ~ 55 km-long seamount ranging from 3100 m to 1950 m of depth (Fig. 2a), is
225 characterized by six peaks located between 2400 and 1950 m of depth (Fig. 2b). It shows a
226 very strong positive magnetic anomaly, particularly in its central part (> 200 nT). The
227 Augusto is separated from an unnamed seamount, here we call “Emilia” (minor depth 2200
228 m) from a NNW-SSE oriented basin. West of the Emilia, we have a deep (> -3500 m) and
229 almost circular basin.

230 The NNE-SSW segment of the PM11 profile (Fig. 3, Box C and C' and Supp. Fig. 3) crosses
231 the Aurelia, a small basin filled by PQ sediments and Augusto. PQ sediments in the small
232 basins are well-stratified and undeformed in the shallow part, while gently deformed in the
233 deeper part. This intra-Pliocene unconformity corresponding probably to the “X
234 Unconformity” (Zitellini et al., 1986) is marked with a violet horizon. Several NE and SW-

235 dipping normal faults control the formation of small basins (see at Fix 13 and 2 in the box C'
236 of Fig. 3).
237 The SW-NE oriented segment of the profile PM10E (Fig. 3, Box D and D' and Supp. Fig. 4)
238 crosses the Emilia, Augusto and Virgilio Seamounts. Emilia and Virgilio are bounded by
239 several SW and NE-dipping normal faults, as suggested by the abrupt lateral interruption and
240 dislocation of well-stratified PQ unit (marked in green). These normal faults dislocate also
241 the basement and an intermediate unit (VIOLET) with a transitional seismo-stratigraphic
242 character. The PQ unit covers part of these two seamounts, part of the intermediate unit and
243 forms thick and narrow basins (see at Fix 17 of box D' in Fig. 3). The Augusto seamount,
244 with a chaotic seismo-stratigraphic character, does not show any sedimentary cover nor
245 faults dislocations.

246

247 *Facies and microfaunal associations*

248 The sample dredged from site 75–30-3 (Aurelia Seamount; Fig. 2a) shows the direct contact
249 between a few centimeters thick completely lithified limestone, and a greenish volcanic rock
250 (Fig. 4a). The upper surface of the volcanic rock appears to be irregular and affected by
251 frequent fractures filled by carbonate mud. A thin non-continuous brownish-to-black film
252 marks the contact magmatic rock/limestone and locally also the neptunian-dykes walls. The
253 carbonate crust overlying the volcanic rock appears to be completely lithified; its upper
254 surface, exposed to seawater, is coated by a thin black film and is strongly colonized by
255 serpulids that are, in turn, covered by a black film. Frequent microhollows representing the
256 moulds of eroded and/or dissolved tests are recognizable.

257 Samples dredged from sites 75–35-(5–7-9) and 75–36-7 (Augusto Seamount; Fig. 2a)
258 consist of 2–3 cm of light brownish consolidated limestones. No direct contact with volcanic
259 rocks has been observed here. The limestone upper surface is coated by a black film strongly
260 colonized by serpulids, which are in turn mineralized. In site 75–35 the carbonate crust is

261 totally colonized by corals and serpulids. Most of the recovered corals are solitary cold-
262 water species as *Desmophyllum* with characteristic cup-shaped morphology and marked
263 septa (Fig. 4b). Their size varies from a few up to 6–7 cm; the younger individuals, growing
264 on top of the older ones, simulate pseudo-colonies. The corals represent fossil occurrences;
265 no living corals have been recovered. The corals surface appears to be almost completely
266 covered by a very thin film, black in color.

267 All the carbonates consist of crusts cemented throughout their total thickness without
268 textural evolution from chalk to limestones. Just in one sample (75–36-7a) a small cavity is
269 filled by a not completely lithified planktonic-rich mud.

270 Microfaunal analysis of the carbonate crusts reveals coral fragments, gasteropods, pteropods,
271 sponge spicules and foraminifera. The foraminiferal content is represented by planktonic
272 taxa widely distributed in the hazel-brown micritic carbonate with mudstone texture or
273 gathered inside bioturbation pockets, bioerosive structures and neptunian dykes within the
274 volcanic substratum. Mn-Fe-oxides films (or permeation or crusts) always outline the
275 bioerosion structures. The most frequent taxa are: *Pulleniatina obliquiloculata*, *Globorotalia*
276 *inflata*, *Globorotalia scitula*, *Orbulina universa* and globigerinids (e.g. *G. bulloides*). Genus
277 *Globigerinoides* is less abundant and it is mainly represented by *G. trilobus*. *Globorotalia*
278 *truncatulinoides* (in the 75–30-3) and of *Pulleniatina obliquiloculata* were also observed
279 (Fig. 4d).

280 X-ray diffraction on the carbonate crusts showed Mg-calcite and calcite as the most
281 abundant phases with a subordinate silicate fraction with mica, chlorite, quartz, and probably
282 montmorillonite. Also SEM analysis, coupled with EDS, showed predominance of carbonate
283 composition with a very subordinate clay-sized silicate fraction (Fig. 4a). Isolated volcanic
284 minerals (quartz and feldspar) and volcanic fragments have been found within the carbonate
285 crust from site 75–36-7 (Fig. 4c).

286 The thin brownish-to-black film marking the contact magmatic rock/ limestone and the

287 neptunian-dykes walls is of Mn–Fe oxides (Fig. 4a)
288 and shows a thin layered structure. All limestone surfaces, borings (Fig. 4e) and/or skeletons
289 exposed to sea-water are coated by a Mn and Fe-oxy-hydroxides thin black-brownish
290 botryoidal film with a laminated texture. The major mineral phase is todorokite, with
291 subordinate amounts of phosphates and montmorillonite (X-ray diffraction data).
292 Based on thin section and SEM-EDS observations, the limestones exhibit a complex
293 diagenetic history. Sometimes the carbonate crusts are characterized by a first thin portion of
294 an early consolidated limestone with bioclastic fragments and diffuse serpulids separated, by
295 means of a thin film of Mn-Fe-oxides, from a portion richer in planktonic foraminifera. The
296 lower portion may display borings and/or fractures, coated by a Mn–Fe oxides film, filled by
297 carbonate pelagic sediments that, in turn underwent a new early lithification and
298 mineralization. All these features indicate at least two phases of early diagenesis
299 accompanied by Mn–Fe mineralization. Both the mineralogical composition and the
300 planktonic associations of the two portions do not show significant differences.

301

302 *Petrography*

303 Samples were chosen for petrographic analyses to verify or discard their igneous nature. In
304 particular, we selected samples with phyrlic-like texture, i.e. large minerals in an aphanitic
305 ground mass. Sample T75–303 (Aurelia Seamount; Fig. 2a) shows white to light-brown
306 grains up to two mm across set in a gray-greenish matrix (Fig. 5a1). A vein of 10 mm
307 maximum apparent thickness cuts the sample. The vein is filled by very fine-grained
308 material and the contact with the host rock is sharp. Sample T75–30-5 displays phyrlic
309 texture made of vitreous grains (average 1 mm across) in a relatively soft light-gray matrix
310 (Fig. 5a2). Inspection under a polarizing optical microscope (Fig. 5a3,4) reveals a texture
311 with euhedral to anhedral grains set in a chlorite and opaque-rich matrix. Matrix minerals are
312 locally aligned to form a fluidal texture. Two types of grains are recognized: (1) euhedral to

313 subhedral grains with pseudo-hexagonal and prismatic shapes (Fig. 5a4,5) and (2) rounded
314 fractured grains with clear appearance under plane polarized light (Fig. 5a3). The euhedral to
315 subhedral grains may be former phenocrysts now completely replaced by secondary chlorite
316 and carbonates, mainly dolomite and minor Mg-calcite, as determined by energy-dispersive
317 spectrometry in a SEM. These phenocrysts occur as single grains or as clusters that locally
318 gives the rock a glomeroporphyritic texture. The rounded fractured grains are made of quartz
319 (sample T75–30-3). The matrix is composed of euhedral to subhedral microlites, now
320 completely replaced by secondary low-Fe chlorite, and rounded Ti-rich minerals. High-Fe
321 chlorite fills the interstices among microlites. Sample T75–30-5 is cut by at least two sets of
322 veins. Earlier veins, maximum thickness 0.25 mm, are filled by coarse grained Fe-bearing
323 dolomite with irregular shapes. Late veins are wider, and are filled by euhedral dolomite.
324 Finally, sample T75–30-5 contains amygdale showing a zonation, from rim to core of the
325 amygdala, chlorite, opaque material and dolomite (Fig. 5a6).

326

327 **Discussion**

328 *Environmental significance of the Aurelia and Augusto rocks and carbonate facies*

329 Samples of rock analyzed from the Aurelia seamount preserve a porphyritic texture with
330 different types of phenocrysts set in a matrix with microlites that locally define a flow
331 texture. These features are compatible with an igneous nature of the samples. Their primary
332 mineralogy is now replaced by secondary phases such as low-Fe chlorite and dolomite.
333 Similarly, the matrix contains microlites now replaced mainly by low-Fe chlorite, and Ti-
334 rich phases that possibly represent remnants of primary Fe–Ti oxides that underwent iron
335 loss during low-temperature rock-water interaction. Plotting our data of Ti and Zr, two
336 elements usually interpreted as relatively immobile during low-T alteration (Pearce, 2014),
337 the rocks of the Aurelia Seamount fall in the calc-alkaline field (Fig. 5b).

338 From a carbonate/biostratigraphic viewpoint, samples from the Aurelia and Augusto

339 Seamounts consist of crusts cemented throughout their total thickness without textural
340 evolution from chalk to limestones. The crusts, with thickness between 15 and 30 mm, are
341 made of Mgcalcite biomicrite rich in planktonic foraminifera with a very subordinate silicate
342 component. The occurrence of *Globorotalia truncatulinoides* on the Aurelia Seamount
343 suggests for this sample an age not older than Early Pleistocene (Iaccarino et al., 2007). This
344 is also confirmed by the presence of *Pulleniatina obliquiloculata*, a warm tropical-subtropical
345 species, that although it occurred in the Atlantic basin in Early Pliocene (Zankl, 1969; Bolli
346 and Saunders, 1985; Iaccarino et al., 2007), it is never recorded in Pliocene Mediterranean
347 deposits. Moreover, according to Bolli and Saunders (1985), *P. obliquiloculata* shows major
348 frequency peaks during the Pleistocene–Holocene time interval. This taxon would have
349 entered the Mediterranean during the Pleistocene warmer climatic stages (Conti et al., 2013)
350 probably from the Atlantic occurring exclusively in the western sector of the Mediterranean
351 basin. An interesting feature is the occurrence within the carbonate crust from the Augusto
352 Seamount of isolated volcanic minerals (quartz and feldspar) and of volcanic fragments,
353 probably indicating a coeval magmatic activity or alternatively a supply of volcanic material
354 eroded from a nearby seamount.

355 The occurrence in many areas of the Mediterranean of Quaternary deep-water cemented
356 limestones with different grades of consolidation, from brittle to consolidated chalk to
357 cemented limestones, has been reported in literature (Allouc, 1990). The study of the
358 carbonate crusts and their relationships with the volcanic substrate, are important for
359 understanding the rapid formation of hardgrounds in ancient sedimentary sequences.

360 A number of studies have focused on the driving mechanisms for the early lithification of the
361 Quaternary deep-sea crusts (Emelyanov and Shimkus, 1986; McKenzie and Bernoulli, 1982;
362 Allouc, 1987, 1990; Remia et al., 2004). Early lithification, occurring at or below the
363 seafloor, takes place under varying conditions: it may occur in sedimentstarved
364 environments, it may depend on ascending interstitial carbonate-rich waters, or on

365 microbiological precipitation or it may be physiochemically controlled and related to inward
366 diffusion of seawater solutions. Concerning the degree of saturation relative to calcite,
367 Mediterranean waters remained saturated at all depths also during the Pleistocene cold
368 phases (Alloué, 1987). Another significant factor controlling carbonate precipitation or
369 dissolution is the concentration of dissolved phosphates and organic matter (Morse, 1986).
370 The carbonate crusts taken into consideration in this study are predominantly made of Mg-
371 calcite with a very subordinate silicate fraction, mostly represented by mica, quartz chlorite
372 and montmorillonite, probably deriving from the weathered volcanic substrate. One of the
373 factors controlling early lithification is the purity of lime mud; <2% of insoluble residue
374 (especially clay minerals) favors cementation and recrystallization (Zankl, 1969).
375 Nevertheless an excess of hydrothermal metals (i.e. Mn-Fe-oxyhydroxides) may “fertilize”
376 areas of normal biological productivity, resulting in massive phytoplankton blooms (Coale et
377 al., 1996; Larson and Erba, 1999; Corda and Palmiotto, 2015). The microorganisms activity
378 may have a significant influence in precipitating hydrothermal Mn-Fe-oxyhydroxide (Dekov
379 and Savelli, 2004); in addition, Mutti and Bernoulli (2003) stress the relationship between
380 phosphate mineralization, trophic resources and microbial micrite precipitation. Based on
381 our observations, we assume a significant role of the microbiological precipitation of calcite
382 in facilitating the early lithification of planktonic ooze.
383 Sources of magnesium are usually seawater and sometimes, fresh waters but could also
384 derive from weathered magmatic Fe–Mg rich rocks. When magnesium is delivered to
385 seawater, Mg-calcite can precipitate (Mackenzie and Andersson, 2013; Morse and
386 McKenzie, 1990).
387 The limited thickness of the carbonate crusts points to very slow rates of sedimentation
388 and/or accumulation probably related to high hydrodynamics. The very thin brownish film of
389 Mn–Fe oxides, even if noncontinuous, covering the volcanic substrate hints at a period of
390 waterrock exposure before the carbonate planktonic-rich mud began to deposit. During this

391 period the volcanic substrate underwent a phase of extensional tectonics promoting intense
392 fracturing, as testified by neptunian dykes partially mineralized and successively filled by
393 pelagic sediments. This substrate should be firstly colonized by small rapidlygrowing
394 opportunistic organisms such as serpulids, favored by strong storm activity, and covered by
395 few millimeters of carbonate mud cemented early. Above this first phase of sedimentation,
396 accompanied by early diagenesis, bioturbation processes and Mn–Fe mineralization, a new
397 phase of sedimentation of carbonate planktonic-rich ooze began, that was rapidly lithified
398 and mineralized.

399 Ultimately our findings suggest that the pelagic sediments settled on the magmatic substrates
400 underwent an early lithification process induced by precipitation of Mg-calcite within the
401 pelagic carbonate matrix in areas of very slow sedimentation rates. The volcanic substrates
402 underwent a tectonic instability, testified by fractures and neptunian dykes infilled by
403 planktonic mudstones. Their morphostructural configuration as isolated highs, suffering high
404 hydrodynamic conditions, is here interpreted as responsible for low
405 sedimentation/accumulation rates, clustering of planktonic foraminifera and skeletal
406 fragmentation. The slow rates of deposition favored prolonged conditions of exposure at the
407 seawater-sediment interface of the pelagic and skeletal carbonates, enhancing the diffusion
408 of seawater-ions throughout the sediments and promoting chemical precipitation and the
409 consequent development of hardgrounds prone to be colonized by corals and serpulids. In
410 addition, the presence of phosphorous as evidenced by the SEM-EDS analysis, suggests that
411 the pelagic carbonates underwent early diagenetic lithification phases under high-fertility
412 conditions which, favoring an increase of microbiota communities, promoted an increase of
413 microbiological micrite precipitation. The widespread Mn–Fe mineralizations covering both
414 the carbonate deposits and the encrusting/colonizing biota on the base of crust morphologies,
415 textural evidence and tectonic setting, can be related to hydrothermal processes.

416

417 *Geodynamic significance of the Aurelia and Augusto seamounts*

418 We discuss here the significance of the Aurelia and Augusto seamounts in the geodynamics
419 of the central Tyrrhenian BAB, based on our results, on data from literature, on the regional
420 bathymetry map and the map of the distribution of the reduced to the pole magnetic
421 anomalies of the Southern Tyrrhenian (Fig. 6a and b, respectively).

422 The Magnaghi Basin (Fig. 6a) starts opening during the Tortonian/ Messinian (Loreto et al.,
423 2020). Volcanism of the Magnaghi Volcano is characterized by basalts with an Na-alkaline
424 affinity, dated from 3.1 to 2.7 Ma (Late Pliocene; Serri et al., 2001), in line with the subchron
425 C2An.1n (3.040 to 2.581 Ma; Cande and Kent, 1995) shown in Fig. 6b. The Vavilov Basin
426 (Fig. 6a) opened between the Late Miocene and the Early Pliocene (the older basalts from
427 the ODP 373 are 7.5 Ma old; Hsü et al., 1978) when their composition is similar to mid-
428 ocean ridges (MORB), as shown by the basalts sampled from the ODP 655 on the Gortani
429 Ridge (Fig. 6a), dated ~4 Ma (Kastens et al., 1988). The beginning of the Vavilov Volcano
430 activity is placed at 3 Ma, at the same time of the spreading of the basin (Robin et al., 1987).
431 The Vavilov lavas, very similar to those of the Magnaghi Volcano, are formed by basalts
432 ranging from tholeiitic to Na-alkaline (Peccerillo, 2017). The end of extension in the Vavilov
433 Basin allows the growth of the volcano, which shows in Fig. 6b a negative magnetic
434 anomaly (C2r.2r; 2.581 to 2.150 Ma; Savelli and Ligi, 2017). According to Parson et al.
435 (1990), the change in chemical composition from MORB to calc-alkaline marks the
436 evolution of a back-arc basin from an early stage, where a pure extensional tectonics
437 produced an oceanic crust with MORB affinity, to a mature stage of back arc spreading,
438 where the oceanic crust has a calc-alkaline affinity. Based on the age of the boundary
439 between the older sediments and calc-alkaline rocks sampled in the ODP 651 (Northern
440 Vavilov Basin; Kastens et al., 1988; Bonatti et al., 1990), we assume that the extensional
441 tectonics of the Vavilov Plain continued until 2.6 Ma.

442 The Marsili Basin (Fig. 6a) starts opening in the Early Pleistocene (basalts dated ~2 Ma from

443 the OPD 650; Kastens et al., 1988); it shows a negative/inverse magnetic anomaly falling in
444 the chron C1r.2r (1.770 to 1.070 Ma; Fig. 6b). The Marsili Volcano is located in the center
445 of the basin, formed by calc-alkaline rocks (Trua et al., 2002) not older than 1.07 Ma
446 (Cocchi et al., 2009), corresponding with the positive/normal magnetic anomaly C1n (0.780
447 to 0.00 Ma; Cocchi et al., 2009; see Fig. 6b). The Marsili is surrounded by several volcanic
448 islands and seamounts (Fig. 6a): the Palinuro Volcanic Complex, a E-W oriented volcanic
449 structure formed by basaltic-andesite compositions lavas dated 0.8–0.3 Ma (e.g., Colantoni
450 et al., 1981; Cocchi et al., 2017); the Alcione, and Lametini 1 and 2 seamounts, related to the
451 geodynamic environment to the Aeolian arc and younger than 1 Ma (Barberi et al., 1973;
452 Beccaluva et al., 1985; Lupton et al., 2011); the Aeolian Islands, consisting of calc-alkaline
453 to shoshonitic lavas and pyro-clastics, with minor potassic alkaline rocks (Peccerillo, 2005),
454 originated by volcanism due to the “wet” melting of a suprasubduction mantle wedge
455 (Lupton et al., 2011); the Enarete, Eolo, Sisifo and Tiro seamounts, the oldest volcanoes of
456 the Aeolian Islands Arc (Beccaluva et al., 1982, 1985). All these volcanoes and seamounts
457 show a normal/positive magnetic anomaly (i. e. Bortoluzzi et al., 2010; Cocchi et al., 2017)
458 that, as in Marsili, can be attributed to the C1n (0.780 to 0.00 Ma), although some of them
459 show a volcanism >0.780 Ma, before the formation of the Marsili Volcano. For example, the
460 Tiro and Sisifo seamounts are formed by calc-alkaline and high K₂O calcalkaline rocks
461 dated ~ 1.5 Ma and from 1.3 Ma to 0.9, respectively (Beccaluva et al., 1985); Enarete and
462 Eolo, formed by basalts, dacites and rhyolites, have been dated between 0.85 and 0.77 Ma
463 (Beccaluva et al., 1982, 1985; Trua et al., 2004).

464 Based on the geodynamic models of Carminati et al. (2010), we reconstructed a cartoon
465 showing the migration of the volcanic arc associated to the Ionian slab during the last 3 Ma
466 (Fig. 7). In this cartoon, at 3 Ma, when the Magnaghi and Vavilov volcanoes were active, we
467 assume that Aurelia, Virgilio and the western part of Augusto were part of the active
468 volcanic arc. We based on: 1) the calc-alkaline basement rocks of the Aurelia (Ti/Z diagram

469 data of Fig. 5b), indicating a subduction-related volcanism; 2) the calc-alkaline basement
470 rocks of the western part of the Augusto (Colantoni et al., 1981), testified also by the
471 occurrence of volcanic fragments in our carbonate samples, and on the strong positive
472 magnetic anomaly (Fig. 6b and Cella et al., 1998); 3) the volcanic nature of the Virgilio,
473 considered a NNE-SSW oriented composite volcanic structure made by a coalescence of a
474 series of centres (Finetti and Del Ben, 1986); 4) the recognized planktonic foraminifera
475 assemblages of the Aurelia and Augusto carbonates, cannot be older than Early Pleistocene,
476 indicating a volcanic activity before 2.58 Ma.

477 Based on the bathymetry maps (Figs. 2 and 6a), the Augusto seamount is composed by
478 several peaks forming a curved arc, located between 2400 and 1900 m below sea level. This
479 curved morphology resembles more the fossil Aeolian volcanic arc, South of the Marsili,
480 formed by the seamounts Sisifo and Tiro, active between 1.5 and 1 Ma (Fig. 7 1 Ma;
481 Beccaluva et al., 1985), than the morphology of the modern Aeolian volcanic arc (Fig. 7–0
482 Ma). Considering the location of the Augusto, between the Aurelia and the Augusto, and the
483 Sisifo and Tiro, and based on the age of their volcanism, we assume the Augusto as part of
484 the Early/Middle Pleistocene Tyrrhenian volcanic arc (Fig. 7 2 Ma). The different trend of
485 the Tyrrhenian volcanic arc between the Late Pliocene (Fig. 7 3 Ma) and the Early
486 Pleistocene (Fig. 7 – from 2 to 0 Ma) could testify the change from eastward to
487 southeastward of the Ionian slab retreat, due to collision with the Apulia platform to the
488 north and the Hyblean platform to the south, during the Pliocene (Van Van Dijk et al., 2000).

489 The geodynamic model by Corradino et al. (2022) considers the area between Vavilov and
490 Marsili as the present place of back-arc extension, and the Marsili as part of a Pliocene
491 volcanic arc. In contrast, our geological, geochronological and petrological results show that
492 the seamounts Aurelia and Augusto were part of the active Tyrrhenian volcanic arc during
493 the Late Pliocene-Early Pleistocene, and that the Marsili is the present back-arc spreading
494 center (positive/normal magnetic anomaly C1n 0.780 to 0.00 Ma; Fig. 6b).

495

496 *The Vavilov-Marsili transfer zone*

497 According to Macdonald et al. (1988) and Pagli et al. (2019), along the BABs, seafloor
498 spreading is offset by transfer zones where strike-slip tectonics transfers displacement
499 between two similar non-coplanar structures. Transfer zones are striking parallel to the
500 regional direction of extension. They are local and passive fracture formed in response to
501 active faulting on faults which link with the transfer; an example is transfer zone is the
502 Central Lau Spreading Centre (CLSC) and the Eastern Lau Spreading Centre (ELSC) in the
503 Lau Back-arc Basin (SW Pacific Ocean; Parson and Wright, 1996). Also, transfer zones
504 could also be inherited from the overriding plate, as in the Northern Lau Basin, where the
505 upper plate is affected by strike-slip tectonics along older tectonic lineaments (Palmiotto et
506 al., 2022).

507 The central region of the Tyrrhenian BAB is characterized by a WNWESE oriented basin,
508 located between the Vavilov and the Marsili basins, perpendicular to the NNE-SSW trend of
509 the Vavilov and Marsili volcanoes (Fig. 6a). This basin shows low values of magnetic
510 anomalies (Fig. 6b) and, based on seismic refraction data (Recq et al., 1984), it is
511 characterized by a minimum value of crustal thickness (the velocity is 6 km/s at only 5 km of
512 depth). We show here part of a sparker profile (PM12; Fig. 6c and Supp. Fig. 5), acquired
513 during the oceanographic expedition T71 (see the Fig. 1a and the methods). The profile
514 PM12 is NE-SW oriented and crosses perpendicular the basin between Vavilov and Marsili.
515 An interpretation of the profile shows that the basin is covered by PQ sediments and it shows
516 in its center, from fix 32 to 37 (Fig. 6c and Supp. Fig. 5), a feature we interpreted as a
517 positive flower structure. This feature can be attributed to a transcurrent fault that in the
518 bathymetry map (Fig. 6a) can be followed from the Southern Vavilov Basin, where affects
519 the Aurelia basement (see the sub-vertical fault in the profiles PM4 and PM11 of Fig. 3 and
520 Supp. Figs.3,4), to the Western Marsili Basin. Furthermore, based on the maps of the

521 distribution of the reduced to the pole magnetic anomalies (Figs. 2c,d and 6b), the upper part
522 of the Vavilov Volcano and a small part of its eastern flank show a positive magnetic
523 anomaly which could be attributed to a recent volcanic event, considering that the summit
524 lavas have been dated between 0.37 and 0.09 Ma (C1n; Robin et al., 1987; Savelli and Ligi,
525 2017). Based on those considerations, we interpreted the Tyrrhenian as a BAB where two
526 different segment of spreadings are active at the same time and, the basin between the
527 Vavilov and the Marsili, their “Transfer Zone” (Figs. 6 and 7).

528

529 **Conclusions**

530 We analyzed geophysical and geological data of two seamounts (Augusto and Aurelia)
531 located South of the Vavilov Volcano. The Augusto is characterized by an arc-shaped
532 morphology, with several peaks located between 1950 and 2400 m below sea level; the
533 Aurelia shows an asymmetric perpendicular profile due sub-vertical faults affecting its
534 northern side, visible both from bathymetry and from the sparker profiles. The distribution of
535 the reduced to the pole magnetic anomalies shows positive values on the Augusto, and low
536 negative values on the Aurelia. Samples of rocks dredges from the Aurelia and the Augusto
537 show a magmatic nature of their basement (basalts with a calkalkaline affinity). Carbonate
538 samples consist of thin crusts cemented early made of Mg-calcite biomicrite rich in
539 planktonic foraminifera, dated not older than Early Pleistocene. Based on our results, we
540 interpret the Augusto and Aurelia as part of the volcanic arc of the Tyrrhenian BAB during
541 the Late Pliocene–Early Pleistocene time.

542

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550

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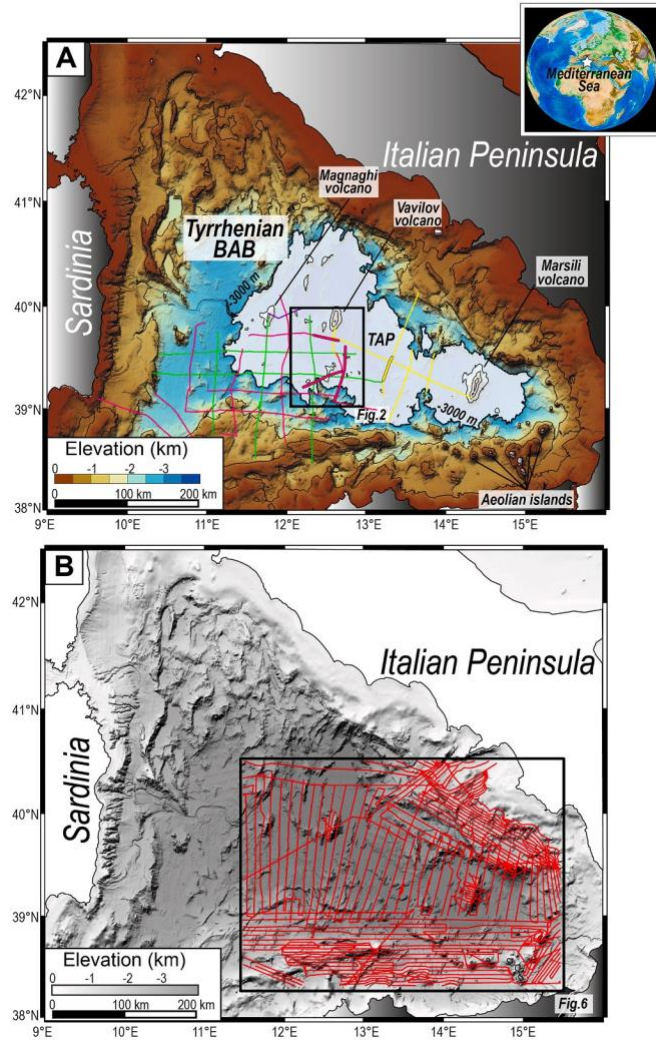
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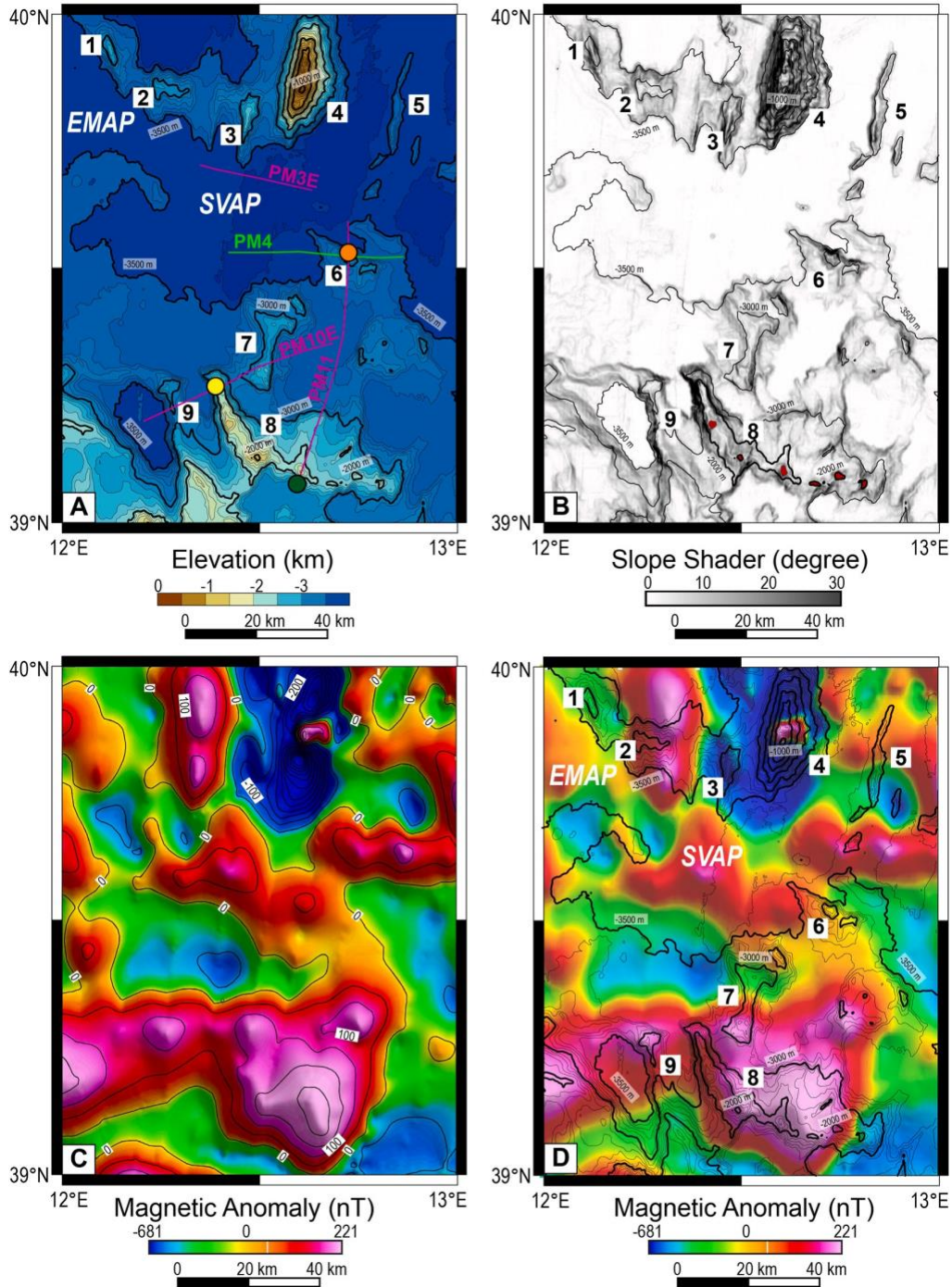
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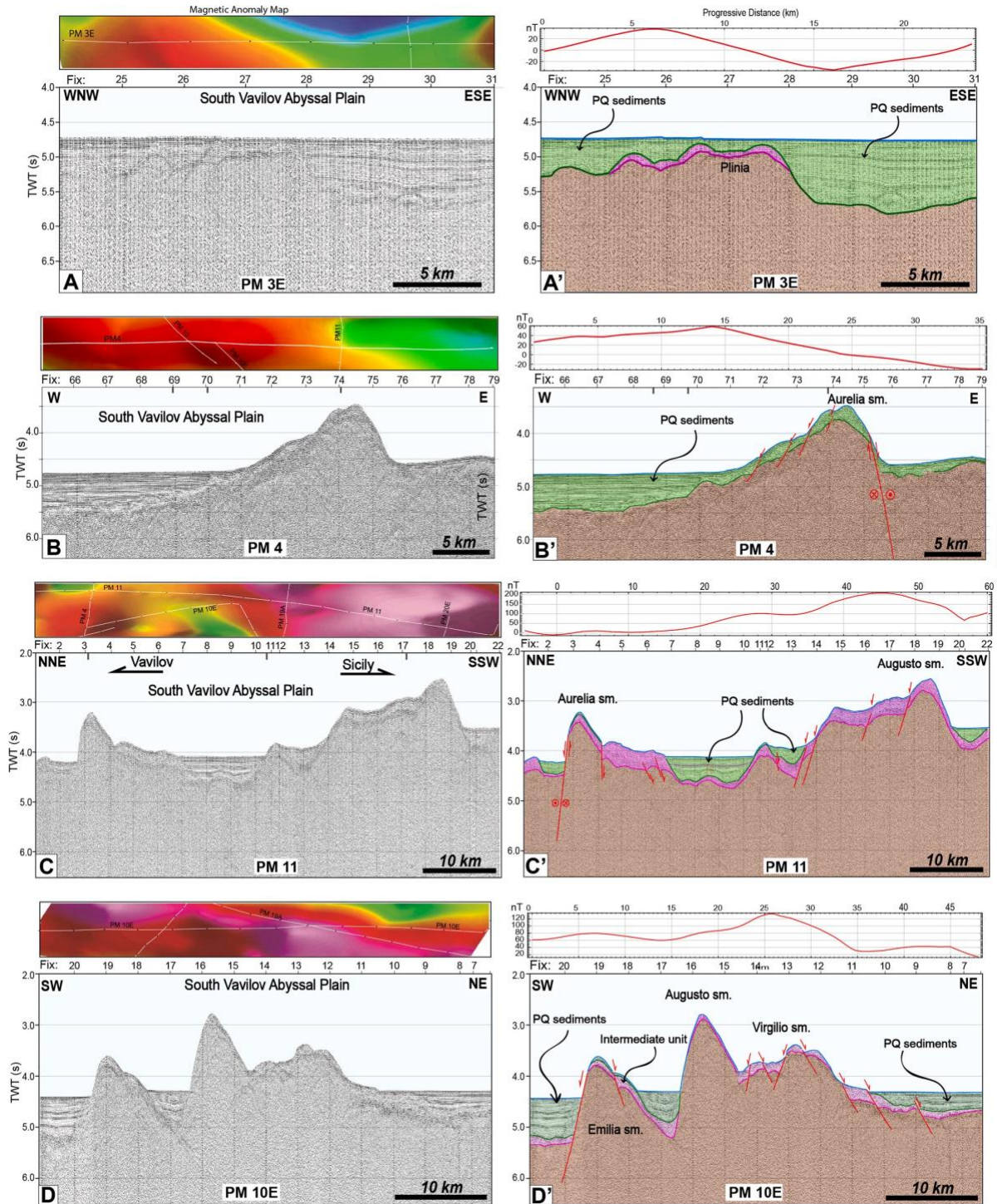
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Fig. 1. Geographic setting of the Mediterranean Sea (white star). A) Base colored bathymetry map of the Tyrrhenian Back-Arc Basin (BAB). Black square is the area shown in Fig. 2; black lines are isobaths (interval of 1000 m); the Tyrrhenian Abyssal Plain (TAP), deeper than 3000 m, has been evidenced with the white area; yellow, green and pink lines are the location of sparker lines acquired in the 1971, 1973 and 1975, respectively. Thickened lines indicate parts of the seismic profiles interpreted in this paper. B) Base shaded bathymetry map of the Tyrrhenian BAB. Black square is the area shown in Fig. 6; black line is the coast line; red lines are the location of the magnetic lines acquired in the 1996 and 1999. Bathymetry has been downloaded EMODnet portal (<http://portal.emodnet-bathymetry.eu/gebco-bathymetry-basemap>) and gridded using GMT open software; bathymetric data have been used to create 2D digital elevation model image using Global Mapper Software.



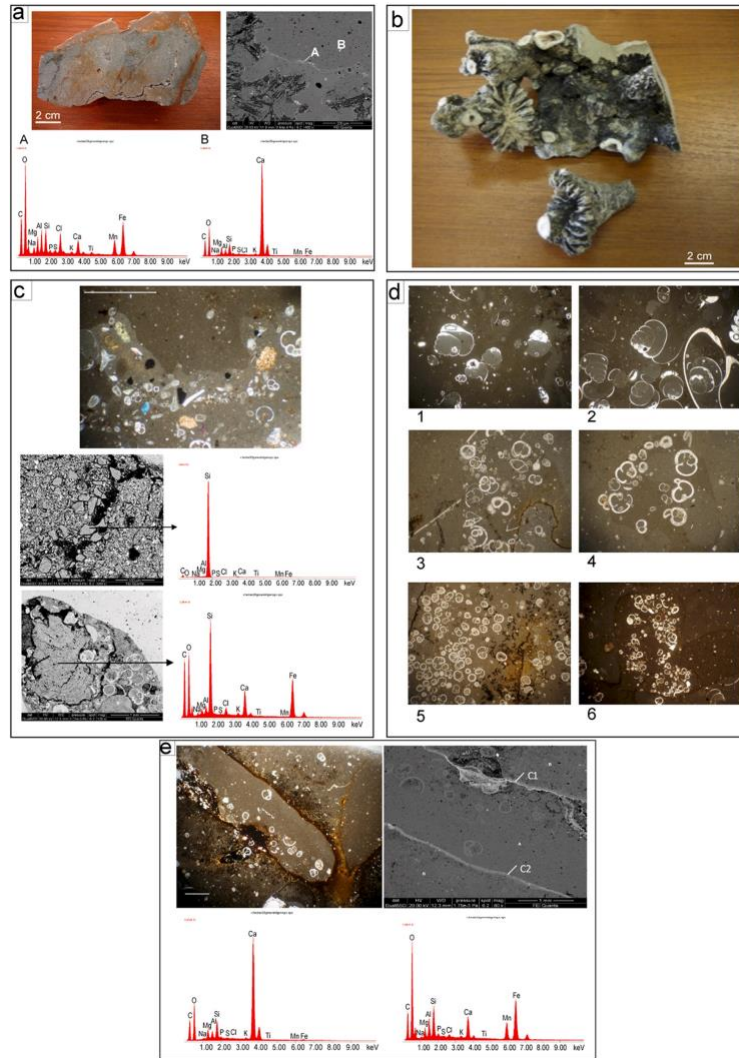
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Fig. 2. Maps of the bathymetry and of the reduced to the pole magnetic anomalies of the Southern Vavilov region. A) Shaded relief image of the bathymetry. Sun angle: 70°; Azimuth: 330°. Vertical Exaggeration: 10. Contour black lines are isobaths every 100 m; green and pink lines indicate the location of the sparker lines shown in Fig. 3; orange, yellow and green dots are the points of the dredges (75–30, 75–35 and 75–36, respectively). B) Slope shader relief map of the bathymetry (isobath interval of 500 m). Red areas show the six peaks along the summit of the Augusto Seamount. Sun angle: 70°; Azimuth: 330°. Vertical Exaggeration: 10. C) Shaded relief image of the distribution of the reduced to the pole magnetic anomalies where contour black lines indicate the lines with the same anomaly value (interval of 50 nT). D) Shaded relief image of the distribution of the reduced to the pole magnetic anomalies where contour black lines are isobaths (interval of 100 m). 1. D’Ancona I Seamount; 2. D’Ancona II Seamount; 3. Plinia Seamount; 4. Vavilov Seamount; 5. Tibullo Seamount; 6. Aurelia Seamount; 7. Virgilio Seamount; 8. Augusto Seamount; 9. Emilia Seamount.



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Fig. 3. Sparker profiles with their relative interpretation, maps of magnetic anomaly and magnetic profiles extracted along the profiles. (A-A') Sparker Line PM3E; (B-B') Sparker Line PM4; (C-C') Sparker Line PM11; (D-D') Sparker Line PM10E



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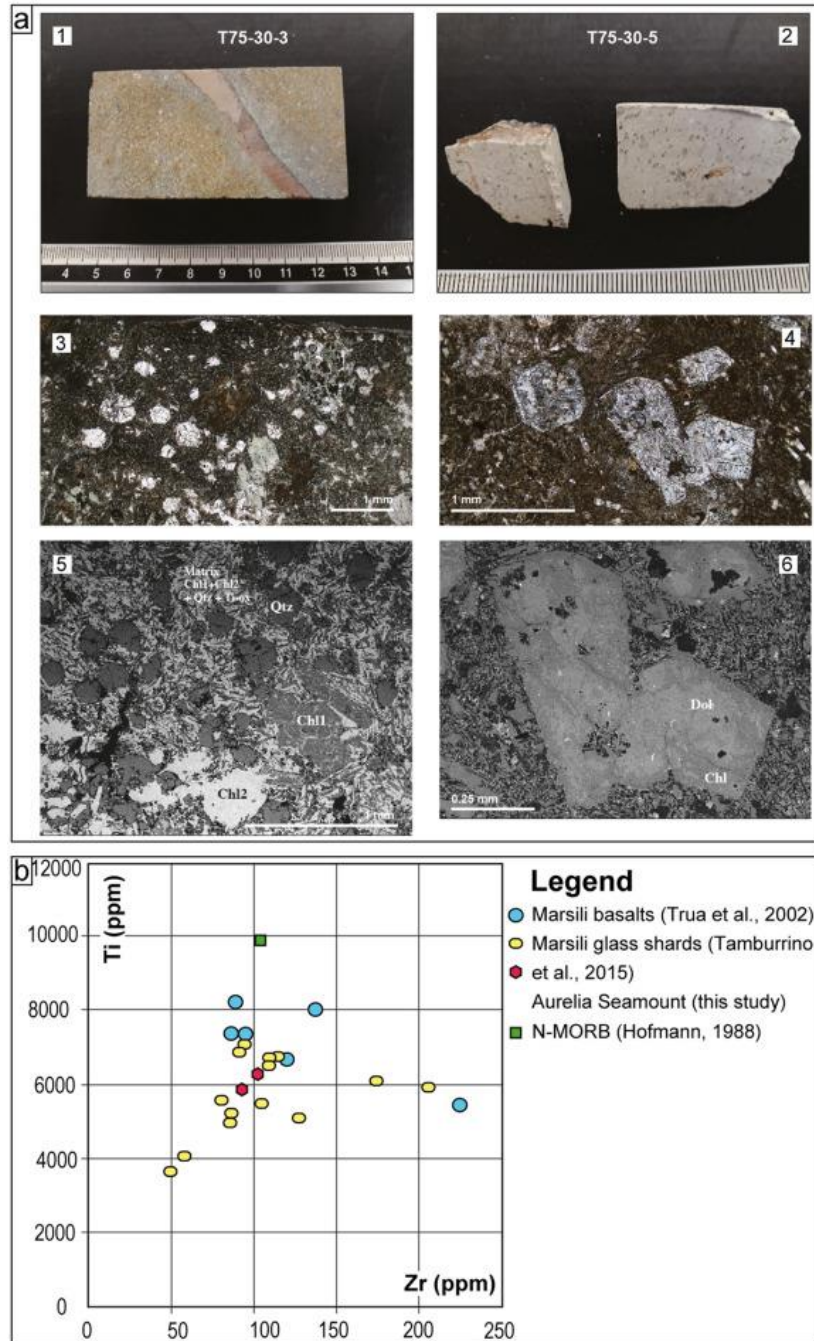
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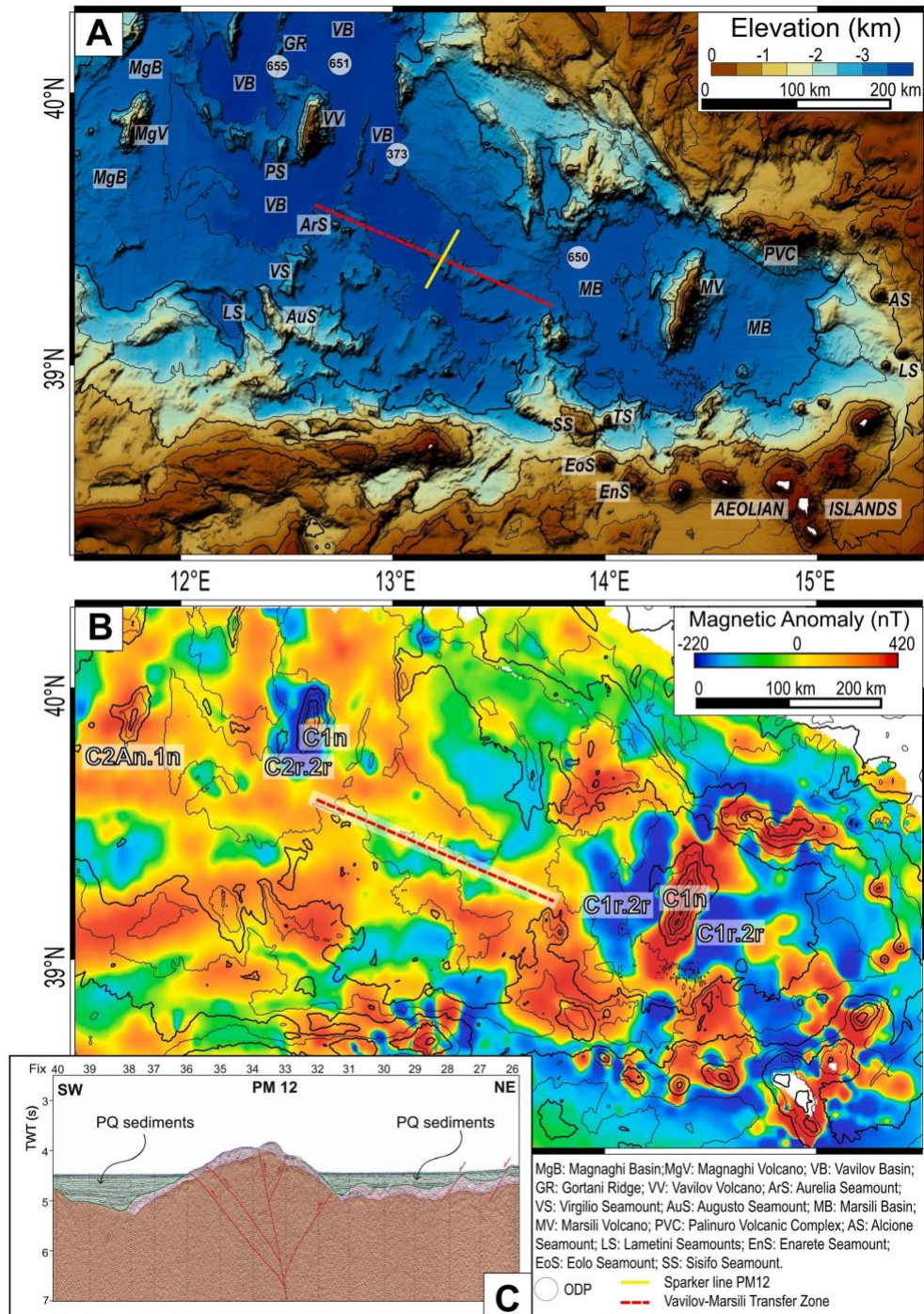
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Fig. 4. (a) Sample from dredge 75-30-3 (Aurelia Seamount). Macroscopic view of the direct contact between the volcanic substrate and the overlying carbonate crust. Neptunian dykes filled by carbonate mud are also visible; SEM photomicrograph of the sample: the related EDAS spectrum B evidenced a predominantly carbonate composition of the crust with a very subordinate clay-sized silicate fraction; the black film marking the contact with the volcanic substrate is represented by Mn-Fe oxides (spectrum A). (b) Black-coated corals (*Desmophyllum*) from dredge 75-35-5 (western Augusto Seamount), scale bar 2 cm. (c) Thin section photomicrographs from dredge 75-36-7A sample (eastern Augusto Seamount) showing isolated volcanic minerals scattered within the planktonic-rich carbonate mud; quartz grains are also evidenced by the SEM photomicrograph of the sample and the related EDAS spectrum. (d) Thin section photomicrographs of the carbonate crusts: 1-2) geopetal structures in pteropod-foraminifer wackestone, pteropod sections are partially filled by foraminifer micrite and microspar at the top; 3) planktonic foraminifera: *Pulleniatina obliquiloculata*, globigerinids, *Globorotalia truncatulinoides* (dredge 75-30-3 sample), scale bar 500 μ m; 4) planktonic foraminifera: *Pulleniatina obliquiloculata*, globigerinids (dredge 75-30-3 sample), scale bar 500 μ m; 5) planktonic foraminifera: *Pulleniatina obliquiloculata*, globigerinids, *Globorotalia inflata*, *Globigerinoides* spp. (dredge 75-36-7A sample), scale bar 1 mm; 6) cloud of planktonic foraminifera, pteropods shell, bioturbation evidences (dredge 75-36-7A sample), scale bar 1 mm. (e) Sample from dredge 75-36-7A. Thin section photomicrograph showing a boring structure filled by planktonic-foraminifer mud; SEM photomicrograph of the sample and the related EDAS spectra evidencing the same composition of the carbonate mud outside and inside the cavity (A and B spectra) and the Mn-Fe oxides-rich film coating the boring.



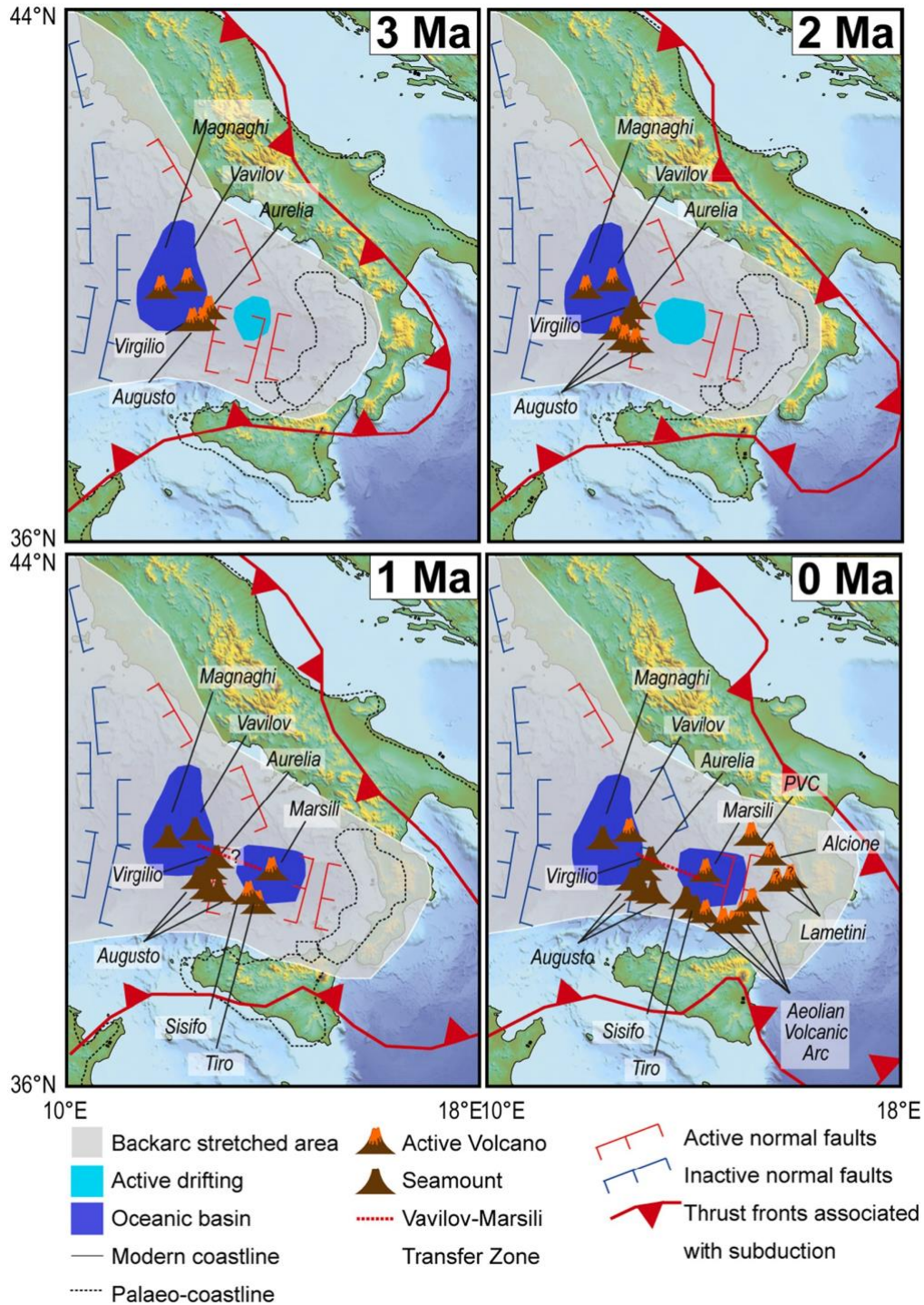
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Fig. 5. (a) 1–2) Low-magnification overview of the chosen samples. (3–4) Photomicrographs (plane polarized light) showing altered phenocryst with euhedral to subhedral shape. (5–6) Backscattered electron images with phase labelling based on EDS microanalysis. Early igneous phenocrysts are now replaced by secondary phases, mostly chlorite and dolomite. (b) Ti–Zr discrimination diagram (Cann, 1970; Pearce and Cann, 1973) where the data of this study are compared to available data from the Marsili Volcano (Pearce, 2014).



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Fig. 6. A) Shaded relief bathymetric image of the Southern Tyrrhenian Sea. Bathymetry has been downloaded EMODnet portal (<http://portal.emodnet-bathymetry.eu/gebco-bathymetry-basemap>) and gridded using GMT open software; bathymetric data have been used to create 2D digital elevation model image using Global Mapper Software. Sun angle: 70°; Azimuth: 330°. Vertical Exaggeration: 10. B) Map of the distribution of the reduced to the pole magnetic anomalies in the Southern Tyrrhenian Sea. Black contour lines are isobaths every 500 m. C) Sparker Profile PM12 with interpretation.



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Fig. 7. Cartoon showing the formation and evolution of the volcanic arc related to the Ionian subduction during the last 3 Ma. Geodynamic reconstruction is based on the model published by Carminati et al. (2010). 3 Ma) Active volcanism of the Magnaghi and Vavilov volcanoes; the active arc is formed by the Aurelia, the Virgilio and the Western part of the Augusto Seamount. 2 Ma) Active volcanism of the Magnaghi and Vavilov volcanoes; the active arc is formed by the Augusto Seamount. 1 Ma) Active volcanism of the Marsili volcano; the active arc is formed by the Sisifo and Tiro Seamounts. 0 Ma) Active volcanism of the Vavilov and Marsili volcanoes; the active arc is formed by the Aeolian Islands and Seamounts.