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Formation and Persistence of Extensional Internally Drained Basins: The Case of the Fucino Basin (Central Apennines, Italy)

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1 **Formation and persistence of extensional internally-drained basins: the case of the**
2 **Fucino basin (Central Apennines, Italy)**

3

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18 **Keywords**

19 Fucino Basin, Slip-direction, Central Apennines, sedimentary load.

20

21 **Highlights:**

22

- 23 • Role of tectonics and drainage systems in the evolution of extensional basins
- 24 • Extensional fault network and sedimentary loading keep the Fucino Basin internally
- 25 drained during its evolution.
- 26 • Sediment load contributes to incrementing of max 30% the faults total geological throw
- 27 and locally reorganizes the stress field.

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32 **Abstract**

33 The interaction between sedimentation/erosion and faulting represents one of the most intriguing
34 topics in landscape and tectonics evolution. Only few studies have been able to document the
35 feedback between faulting and sedimentary loading from field observations. Here, we focus on
36 how sediment loading/unloading influences the dynamics of fault systems in the Fucino Basin, in
37 the Central Apennines (Italy). The Fucino Basin represents a remarkable case study with respect to
38 the other main extensional basins in the Apennines, because of its large dimension, square shape,
39 significant sediment thickness and, its endorheic nature throughout its evolution.

40 We present a detailed structural and geomorphologic analysis of the Fucino Basin and its
41 surroundings, investigating the kinematic and geometry of each main fault strand. The slickenlines
42 analysis reveals multiple families of slip-vectors and timing of activity, suggesting a change in
43 extension slip-direction from N240° to N200° during middle Pleistocene. Using a local isostatic
44 model, we estimate that up to the 30% of the vertical geological displacement of the faults, which
45 overall ranges from 0.5 to 2.5 km, is related to the sediment loading/unloading. We demonstrate a
46 positive feedback between sedimentation and faulting which may also lead to a re-organization in
47 fault kinematics related to a significant increase in vertical stress. We propose a conceptual model
48 for the permanent endorheic configuration of the Fucino Basin, which includes the effect of
49 sediment loading.

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64 **1. Introduction**

65 Extension of the crust generates normal faults and basins where sediments are trapped.
66 Morphologically, extensional basins are generally classified as internally (endorheic) or externally
67 (exorheic) drained. In the case of an internally drained basin, sediments are transported and
68 stored within it. If the basin is externally drained, sediments are transported out of the basin and
69 deposited in distal regions. The internally drained stage of extensional basins is generally a
70 transient feature as these basins often evolve into an externally drained basin. Endorheic-exorheic
71 and exorheic-endorheic transitions have been documented in several rifts basins, in compressional
72 and extensional settings, to unravel the feedback between sedimentation-erosion-climate and
73 tectonics (e.g. Berry et al., 2019; Bialas and Buck, 2009; Bucher et al., 2021; Garcia-Castellanos,
74 2007; Garcia-Castellanos et al., 2003; Geurts et al., 2018; Nichols, 2004; Sobel et al., 2003; Strecker
75 et al., 2007). Numerical models of landscape evolution describe the role of strain rate in
76 extensional basin systems, such as Rio Grande and East African rifts (Berry et al., 2019).

77 Drainage systems and sedimentary budgets of basins impact their fault systems. In fact, the
78 sedimentary budget may influence the fault patterns (e.g. Ballato et al., 2019; Maniatis et al.,
79 2009; Olive et al., 2014; Steer et al., 2014; Turpeinen et al., 2008; Vernant et al., 2013; Whipple,
80 2009). Local unloading may influence fault slip, and even trigger seismicity (e.g. Calais et al., 2010;
81 Hampel et al., 2009, 2007; Hampel and Hetzel, 2006; Hetzel and Hampel, 2005; Mey et al., 2016).
82 The feedback between faulting and sedimentary loading may thus be particularly relevant for
83 internally drained basins, where the sediments are expected to be more abundant.

84 In this paper, we investigate the relation between faulting and sedimentation, focusing on the
85 Fucino Basin in the Central Apennines (Italy, Fig. 1). Here, normal faulting and erosion processes
86 shaped the landscape, producing a basin-and-range-type mountain belt forming both endorheic
87 and exorheic basins. The main tectonic basins are Rieti, Leonessa, L'Aquila, Sulmona, Fucino and
88 Subequana Valley (Fig. 1a). Some of these basins are externally drained by the Aterno-Pescara, the
89 Sangro, the Salto-Tevere and the Liri rivers. Other basins are internally drained such as the Fucino,
90 the Campo Felice, the Piani di Pezza and the Altopiano delle Rocche basins (Fig. 1b). Despite the
91 fact that all these basins evolved at the same time and under the same tectonic regime (e.g.
92 Cosentino et al., 2017), they show remarkable differences in terms of dimensions and sediments
93 thickness. We focus in this study on the Fucino Basin, because of: (1) its large dimensions (~200
94 km²); (2) an almost square shape due to the presence of two orthogonal faults (Fig.1) rather than
95 an elongated shape (e.g. Sulmona or L'Aquila Basin; Fig.1); (3) sediment thickness which reaches

96 almost 1 km (Cavinato et al., 2002) compared to 350 m (e.g. Sumona Basin, Subequana Valley,
97 L'Aquila; Gori et al., 2017; Nocentini et al., 2018, 2017), and 600 m (eastern L'Aquila basin;
98 Nocentini et al., 2018). The Fucino Basin has remained endorheic whereas the other intermontane
99 basins underwent through endorheic-exorheic transitions and vice versa (e.g. D'Agostino et al.,
100 2001b; Geurts et al., 2019, 2018).

101 The aim is to understand the conditions that allowed the Fucino Basin to preserve its internal
102 drainage during its entire evolution. We perform a detailed structural analysis to define the
103 geometry and the kinematic evolution of the main faults around the Fucino Basin. We also provide
104 a geomorphological analysis including swath profiles, steepness index, local relief and slope maps,
105 and drainage divide stability analysis (Forte and Whipple, 2018) to investigate the linkage between
106 erosion and faulting. Our results reveal that sedimentary processes increase the fault offsets
107 around the Fucino area up to 30%, mainly due to both hanging wall loading and footwall
108 unloading. We interpret the observed change in strain direction as an increase of the vertical
109 stress due to the sedimentary loading. We speculate that the coexistence of two orthogonal faults
110 (which is the major difference between the Fucino Basin and any other intermontane depression
111 in the Central Apennines) combined with the sediment (un)loading is responsible for enhancing
112 the subsidence and the stress reorganization. This prevents the Fucino Basin to be captured by
113 nearby rivers, favoring the long-lasting endorheic state of the basin.

114

115 **2. Geological Setting**

116 **2.1. Structural setting, kinematic and throw distribution along the Central Apennines**

117 The Apennine mountain belt is an asymmetric Neogene fold-and-thrust accretionary wedge
118 produced by the subduction of the Adria Plate (Cipollari and Cosentino, 1996; Cosentino et al.,
119 2010; Merlini and Mostardini, 1986; Patacca et al., 1992). Crustal accretion is progressively
120 younger eastward toward the foreland and is followed by extensional deformation. Extensional
121 deformation produced a set of normal faults striking sub-parallel to the older thrusts and trending
122 mainly NW-SE. The most active normal fault system is located on the top of the wedge and it is
123 responsible for the formation of intermontane basins. The only exception to this pattern is the
124 Avezzano-Buzzi fault system which strikes instead NE-SW from the Fucino to the Sulmona basin
125 (Fig. 1a; Gori et al., 2017).

126 Several studies documented and explored the geometry and kinematics of the normal faults,
127 which are particularly well exposed on the carbonate deposits of the Central Apennines (Cowie et

128 al., 2017; Cowie and Roberts, 2001; Faure Walker et al., 2012, 2010, 2009; Galadini, 1999; Galadini
129 et al., 1997; Galadini and Galli, 2000; Gori et al., 2017; Papanikolaou et al., 2005; Piccardi et al.,
130 1999; Pizzi and Pugliese, 2004; Roberts et al., 2004; Roberts and Michetti, 2004; Salvi et al., 2003;
131 Smeraglia et al., 2016; Wilkinson et al., 2015); investigating also the role of tectonics in landscape
132 evolution of the Central Apennines (e.g. Bubeck et al., 2015; Geurts et al., 2019, 2018; Whittaker
133 et al., 2007). Structural analysis of the faults shows a prevalent NNE-SSW to ENE-WSW direction of
134 extension (Faure Walker et al., 2010), with a mean direction oriented NE-SW. This is consistent
135 with focal mechanisms, borehole breakout data, inversion of micro-earthquakes and geodetic
136 strain (Fig. 1a; D'Agostino et al., 2001a; Montone et al., 2012, 1999). In several locations
137 surrounding the Fucino basin, multiple direction of slickenlines have been measured and: (1)
138 averaged and interpreted as due to fault growth and linkage (e.g. Roberts and Michetti, 2004), (2)
139 associated to block rotation (Piccardi et al., 1999) or in general to a previous tectonic deformation
140 (Galadini, 1999). The geological throw of those normal faults ranges from 500 to 2500 m and is
141 larger on the Sulmona, L'Aquila and Rieti faults (Morewood and Roberts, 2000; Roberts and
142 Michetti, 2004). A maximum cumulative throw of 6600 ± 800 m has been estimated along a
143 section crossing the Fucino and the Sulmona basins (Roberts and Michetti, 2004).

144

145 **2.2. Structural setting of the Fucino Basin**

146 The Fucino Basin is one of the largest extensional basins in the Central Apennines and is at an
147 elevation of 600 m a.s.l. It is bordered to the east by the San Benedetto, Pescara and Collarmele
148 faults, which extend southward and northward to become the Gioia de Marsi and the Ovindoli-
149 Celano faults, respectively; by the Vallelonga fault on the western side (Fig. 2) and bordered to the
150 north by the Tremonti and Velino-Magnola faults. These two sets of orthogonal normal faults give
151 the Fucino Basin an almost square shape. Other extensional basins are more elongated and
152 controlled by only one faults strand (Fig.1a). The existence of the two sets of orthogonal faults as
153 well for the age of the Avezzano-Buzzi faults is unclear (Tremonti and Celano-Aielli; Fig. 1). It has
154 been suggested that the Avezzano-Buzzi fault was active during the Apennines compressional
155 phase as a lateral ramp and then has been reactivated during the Plio-Quaternary extension (e.g.
156 Gori et al., 2017). Further north, two other main faults are present: the Piani di Pezza and Campo
157 Felice (Figs. 2 and 3).

158 The Fucino Basin is surrounded by four major exorheic catchments: the Salto-Tevere and Liri
159 basins to the west and the Sangro and the Aterno-Pescara basins to the east (Fig. 1b). On its

160 northern side, the Fucino Basin is delimited by small endorheic catchments, such as the Campo
161 Felice, the Piani di Pezza and the Altopiano delle Rocche basins (Fig. 1b).
162 Paleoseismological trenching across these structures show evidence of several large earthquakes
163 in the last 7 ky affecting the Piani di Pezza-Ovindoli, Gioia de Marsi and Vallelonga faults with a slip
164 from submeter to meters (D'Addezio et al., 1996; Galadini and Galli, 1999; Michetti et al., 1996;
165 Pantosti et al., 1996; Salvi et al., 2003). Slip rate deduced by ^{36}Cl cosmogenic dating along the
166 faults escarpments ranges from 0.2 to 1.2 mm/yr (Benedetti et al., 2013; Palumbo et al., 2004;
167 Schlagenhauf et al., 2011). Cosmogenic analysis and paleoseismological trenching attest an
168 increase of Holocene slip around the Fucino Basin (Benedetti et al., 2013).
169 Interpretation of seismic lines and borehole stratigraphy data provide depth, geometry, and
170 maximum sediment thickness of the Fucino Basin (Cavinato et al., 2002). The sedimentary infill is
171 up to 950 meters thick, and consists of lacustrine, alluvial and fluvial sediments deposited from
172 late Pliocene onward (Cavinato et al., 2002; Mondati et al., 2021). Two sedimentary sequences
173 (sequence 3 and 4, Fig. 2; Cavinato et al., 2002), separated by an unconformity, indicate the
174 occurrence of main extensional or climate events during the late Pliocene and since middle
175 Pleistocene (Cavinato et al., 2002). The Piani di Pezza and Campo Felice faults are younger than
176 the Fucino faults, and have been active since the middle Pleistocene (Giraudi, 2012; Giraudi et al.,
177 2011; Villani et al., 2015).

178

179 **3. Methods**

180 **3.1. Structural analysis**

181 We carried out a structural analysis around the Fucino Basin and measured the strike and dip of
182 the main faults and the pitch angle or the azimuth of the slickenlines. The sense of shear of these
183 structures has been deduced by kinematic indicators such as Riedel shear and lunate structures. In
184 case of multiple slickenlines, cross cutting relationships provide the relative chronology. We
185 plotted the analyze data using the '*Daisy*' software (Salvini, 2004). Figure 3 shows the transport
186 direction of the hanging wall with respect to the footwall of each measurement site and the
187 cumulative along each fault (stereoplot; Fig. 3). Field observations and frequency analysis of the
188 azimuth of the slip-directions were performed to distinguish between different slickenside
189 populations. In supporting information (Figs. S3, S4 and S5), we show data and gaussian
190 distribution of the slip directions for each fault strand. This analysis, together with the cross-

191 cutting observations, is essential to distinguish between a dispersion of data collection along a
192 mean direction and/or two distinct clusters.

193 Serial cross sections provide the throws along faults using geological markers and corrected for the
194 apparent angle if sections were not orthogonal to the fault strands (Figs. S3, S4 and S5 supporting
195 information). Since multiple fault slip-directions are present, the fault throw cannot be estimated
196 easily, thus we estimate the vertical component of the fault throw.

197

198 **3.2. Morphometric quantitative analysis**

199 We perform a morphometric quantitative analysis, including swath profiles, maps of local relief
200 and slope, river network analysis (k_{sn}), and drainage divide stability analysis to investigate the
201 landscape and the drainage area of the Fucino Basin. All these analyses have been realized by
202 using SRTM DEM raw topographic data (~90 m in horizontal resolution)
203 (<http://srtm.csi.cgiar.org/srtmdata/>) in GIS environments and MatLab. Two swath profiles,
204 respectively NW-SE and SW-NE trending, are extracted from SRTM DEM data by using
205 TopoToolBox (Schwanghart and Kuhn, 2010). The elevation along each profile is sampled each 100
206 m into an observation window 5 km wide. The final plot represents the patterns of maximum,
207 minimum and mean topography (Isacks, 1992).

208 We perform a divide stability analysis around the Fucino Basin following Forte and Whipple (2018).
209 This analysis provides a snapshot of a possible watershed migration by a comparison of four
210 metrics across divide: gradient, local relief, χ (a proxy of steady-state channel elevation; see Willet
211 et al., 2014), and elevation. We do not discuss the χ outcome since it is strongly influenced by the
212 choice of the outlet elevation (Whipple et al., 2017) and in this respect, the Fucino base level
213 elevation is too different from any other reasonably chosen outlet for the external rivers (e.g. sea
214 level, change from bedrock to alluvial channel). In the analysis we use 10^4 m^2 as reference area
215 and 1000 m as radius of the local relief. We chose 10 segments along the main drainage divide
216 from different location in the Fucino Basin. Using a free MatLab tool
217 (<http://github.com/amforte/DivideTools>; Forte and Whipple, 2018), we plotted the relative results
218 in a single graph (Fig. 8b) where the error bar is the standard deviation derived from the
219 histograms included in the supporting information (Fig. S17). This plot provides informations of
220 the stability or of the direction of migration of each divide segment (Fig. 8b).

221 The slope and the local relief maps are created in a GIS environment. The latter represents the
222 result of the subtraction of maximum and minimum topography elevation within a circular moving

223 window of 5 km in radius. This proxy allows in the first order to identify (assuming the same
224 lithology, as for the case for this region) incised sectors, and in comparison with faults, their
225 bearing with tectonic.

226 The map of the normalized along-channel variation in steepness index (k_{sn}) has been elaborated by
227 TopotoolBox (Schwanghart and Kuhn, 2010). K_s is a morphometric parameter which, together with
228 the concavity index (ϑ), describes the shape of a river longitudinal profiles providing information
229 about the state of equilibrium of the drainage systems (Duvall et al., 2003; Kirby et al., 2003; Kirby
230 and Whipple, 2012; Snyder et al., 2000; Zaprowski et al., 2005). Usually, to compare river profiles
231 irrespective of basin drainage area, the k_s is normalized (k_{sn}) by using a reference concavity index
232 value in a range from 0.3 to 0.8. The map of the along-channel variation in steepness index shows
233 (Fig. 8e) the changes in channel slope allowing an easy identification of knickpoints and their
234 possible correspondence with tectonic structures or rock type changes.

235 Although these analyses are strongly affected by the erodibility of the rock types, the study area
236 (Fig. 2) is composed mainly by one lithology: the Meso-Cenozoic limestone (Fig. 2). Thus, in the
237 study area, any variation in incision rates can be associated to climate or tectonic events rather
238 than lithological variations.

239

240 **4. Results**

241 **4.1. Structural data**

242 The Fucino Basin has a square shape, controlled mainly by Pescara-San Benedetto-Collarmele set
243 of faults trending NW-SE to the east, and the Tremonti and Velino-Magnola faults trending WSW-
244 ENE and NW-SE, to the north (Figs. 2 and 3). We integrated our structural data and geological
245 displacements (long-term displacement; see Puliti et al. (2020) for details) for each fault strand are
246 compiled with data from previous studies concerning the escarpments height (coseismic rupture
247 providing the short-term displacement; see Puliti et al. (2020) for details), the morphologic
248 displacements (morphologic evidence of faulting such as triangular facets providing the mid-term
249 displacement; see Puliti et al. (2020) for details), the age of the sediments and the slip-rate from
250 cosmogenic dating and paleoseismological trenching.

251 All the structural data collected in the field are listed in the 'Supporting information' (Table S1 and
252 Fig. S1).

253 • The Campo Felice fault (CF) borders the endorheic basin to the NE. The fault strikes for ~12 km
254 with an average direction of N125° (Figs. 2 and 3). We measured two senses of motions: one

255 oriented N202° and one, measured only at the eastern tip of the fault, oriented N263° (Table S1
256 and Fig. S3; supporting information). Using the Cretaceous bauxite levels as marker we estimated
257 a maximum vertical geological offset of ~1500 m (Fig. S3 supporting information). The maximum
258 morphologic displacement along the CF fault is 230 m (Benedetti et al., 2013), the height of the
259 most recent escarpment is ~10 m (Wilkinson et al., 2015) and the Holocene slip rate is ~1 mm/y
260 (Benedetti et al., 2013). Although the onset of the CF is unconstrained, the oldest deposits within
261 the basin are middle Pleistocene in age (Giraudi, 2012; Giraudi et al., 2011).

262 • The Ovindoli-Celano fault (O) strikes for ~12 km with a N-S orientation controlled at depth by the
263 Ovindoli-Monti D'Ocre thrust front (De Vittorio and Faccenna, 1987). The structural analysis shows
264 two senses of motion oriented N230° and N183° but no crosscutting relationship have been
265 observed (Fig. 3 and Fig. S4 within supporting information).

266 • The Piani di Pezza fault (PP) represents the northern prosecution of the Ovindoli-Celano fault
267 and borders the endorheic basin to the NE (Figs. 2 and 3). The fault strikes for ~5 km and it is
268 oriented from WNW-ESE in the western region, to NW-SE in the eastern region. Two sets of
269 striations (PP2 in Figs. 2 and 4C) with clear cross cutting relationship indicate two directions of slip
270 oriented N224° (older event) and oriented N167° (younger event) (Table S1 and Fig. S3; Supporting
271 information). We estimated a maximum vertical geological offset of ~1800 m (Fig. S3 Supporting
272 information). The height of the most recent escarpment is ~5.7 m (Villani et al., 2015). Trench
273 exploration gives a Holocene slip rate ranging between 0.6 to 2.3 mm/y (Pantosti et al., 1996).
274 Base on the oldest deposit within the basin the onset of the PP fault is middle-late Pleistocene
275 (Cinti et al., 1992; Giraudi and Frezzotti, 1997; Pantosti et al., 1996; Villani et al., 2015).

276 • The Velino fault (V) borders the SW flank of the Velino Mountain (Fig. 2). It strikes for ~14 km
277 and it is oriented N150° (Fig. 3). The Velino fault structural analysis shows one set of striations
278 oriented N200° (Fig. S2; Supporting information). We estimated a vertical geological offset of
279 ~1000 m (Fig. S3; Supporting information). The maximum morphological displacement along the
280 Velino faults is ~750 m (Schlagenhauf et al., 2011) while the average scarp height is ~9.5 m
281 (Benedetti et al., 2013). Even though the onset of extension for this structure is not well
282 constrained, the Velino fault is considered active since the middle-late Pliocene (Cavinato et al.,
283 2002).

284 • The Magnola fault is the eastern portion of the Velino fault and borders the southern flank of the
285 Magnola Mountain striking WNW-ESE (Figs. 2 and 3). Slip-directions cluster around N199° (Table
286 S1 and Fig. S4; supporting information). At the measurement site M2 (Figs. 3 and 5), two families

287 of striations and cross cutting relationships indicate two directions: one recent oriented N195° and
288 one older oriented N235° (Fig. 5, M2). Using the stratigraphic contact between Miocene and
289 Cretaceous rocks we estimated a vertical geological offset of ~2500 m (Fig. S4; supporting
290 information). The morphologic displacement along the Magnola faults is up to 800 m
291 (Schlagenhauf et al., 2011) while the escarpment height is up to 20 m with a slip rate of ~1.3 mm/y
292 (Benedetti et al., 2013; Galli et al., 2012). The Magnola and Velino faults are considered active since
293 the middle-late Pliocene (Cavinato et al., 2002), but the exact onset of extension is not
294 constrained.

295 • The Tremonti fault borders the homonymous mountain, and it represents the direct north-west
296 rim of the Fucino Basin (Figs. 2 and 3). It regionally strikes WSW-ENE. The slickenlines analysis
297 shows multiple striations oriented N210°, N126° and N266° (Table S1 and Fig. S4; supporting
298 information). At site TRM4, we observed striations oriented N183° crosscutting the ones oriented
299 N261° (Fig. 6). The vertical geological displacement is ~1000 m (after Cavinato et al., 2002),
300 morphologic displacement is ~450 m and the escarpment height is ~3.5 m with a slip rate of ~0.2
301 mm/y (Benedetti et al., 2013). The onset of faulting is debated due to its possible activation during
302 the Apennines orogenesis as a lateral ramp (Gori et al., 2017). Cavinato et al. (2002) proposed a
303 late Pliocene onset of activation.

304 • The Celano-Aielli fault represents the eastern portion of the Tremonti fault, located east of the
305 San Benedetto fault (Fig. 2). It strikes NE-SW, and we measured three main senses of motion: one
306 N267°, one N209° and one N98° (Fig. 3 and S5 supporting information). The Celano-Aielli fault is
307 considered to be active from middle-late Pliocene (Cavinato et al., 2002; Gori et al., 2017).

308 • The Aschi Alto (or San Sebastiano) fault represents the southern portion of the Pescara Fault (Fig.
309 2). It strikes NW-SE and show two sets of striations; N214° and N162° (Table S1 and Fig. S5;
310 supporting information). The vertical displacement is ~500 m (Praturlon, 1968) (Fig. 2), the
311 escarpment height is ~4 m, and the slip rate is ~0.5 mm/y (Benedetti et al., 2013). The Aschi Alto
312 fault is active from middle-late Pleistocene (Cavinato et al., 2002; Gori et al., 2017).

313 • The Gioia Dei Marsi fault represents the southern portion of the San Benedetto fault (Figs. 2 and
314 3) and strikes N120°. We measured two senses of motion with cross cutting relationships (Fig. 3
315 and 7). The older striation is oriented N182° while the younger striation is oriented N225° (Fig. 7,
316 Table S1 and Fig. S5; supporting information). The vertical geological displacement is of ~400 m
317 (Praturlon, 1968) (Fig. 2). The Gioia Dei Marsi fault is considered active since middle-late Pliocene
318 (Cavinato et al., 2002; Gori et al., 2017).

319 • The Vallelonga (or Trasacco) fault borders the valley located in the southern rim of the Fucino
320 Basin (Figs. 2 and 3). It strikes ~20 km with a NW-SE orientation. We measured two slip-directions
321 and a cross cutting relationship: one younger oriented N253° and one older oriented N171° (Table
322 S1 and Fig. S5; supporting information). The vertical geological displacement is ~300 m (Praturlon,
323 1968), the maximum morphologic displacement is ~250 m and the maximum escarpment height is
324 ~4.7 m with a slip rate of ~0.2 mm/y (Benedetti et al., 2013).

325

326 **4.2. Geomorphological and morphometric analysis**

327 The swath profile A-B (Fig. 8a) shows a marked difference in the general topographic setting
328 between the northwestern (at the border between the Fucino and the Salto R. basin) and the
329 southeastern side (at the border between the Fucino and the Sangro R. basins). The northwestern
330 side shows a maximum topography lower (< 1000 m) than the southeastern side (~2000 m) and a
331 minimum elevation only ~50 m lower than the Fucino Basin one. To the SE, the minimum
332 topography is above 1000 m of elevation, with a peak, representative of the drainage divide,
333 around 1300 m.

334 The swath profile C-D runs orthogonal to the main tectonic structures of the area, highlighting the
335 influence of tectonics on landscape (Fig. 8a). In general, the pattern of minimum, mean, and
336 maximum elevation is similar. To the SW, the Liri R. valley interrupts the topography decreasing
337 towards the Fucino Basin plain. Eastward, beyond the plain, the topography increases regularly up
338 to ~1500 m and then decreases towards the Aterno R. valley.

339 Our analysis of the drainage divide stability (Fig. 8b; and Fig. S10 Supporting information) suggest
340 overall stability in the Fucino Basin. The maps of local relief, slope and k_{sn} (Figs. 8 c, d, e) show a
341 similar pattern with the highest values (local relief of 1200-2100 m, k_{sn} of 80-500 and slope >30°)
342 corresponding with the footwalls of normal faults and hanging walls of thrusts. The remainder of
343 the area is characterized by medium-low values with the lowest ones (local relief and k_{sn} ranges of
344 0-300 m and 0-20, respectively) associated with the bottom of main fluvial trunks (Figs. 8 c, d, s).

345

346 **5. Discussion**

347 **5.1. Synthesis of the structural data and comparison with previous studies**

348 Previous studies analyzed the structural setting of the region (Faure Walker et al., 2012, 2010,
349 2009; Galadini, 1999; Galadini et al., 1997; Galadini and Galli, 2000, 1999; Morewood and Roberts,
350 2000; Papanikolaou et al., 2005; Piccardi et al., 1999; Pizzi and Pugliese, 2004; Roberts and

351 Michetti, 2004; Salvi et al., 2003; Smeraglia et al., 2016; Wilkinson et al., 2015). Our field
352 measurements, and the differences with previous studies are illustrated in the supporting
353 information. Nine figures and one table, synthesized field data measurements (Figs. S7 to S15 and
354 Table S2; Supporting information; Morewood and Roberts, 2000; Piccardi et al., 1999; Roberts and
355 Michetti, 2004; Smeraglia et al., 2016; Faure Walker et al., 2009; Wilkinson et al., 2015).
356 Measurements from this and previous studies are similar, in some case identical (Fig. S7 to S15;
357 Supporting information). Stereoplots from all previous works show the presence of multiple slip-
358 directions that have been interpreted: (1) as due to block rotation (as Piccardi et al., 1999), (2)
359 older tectonic phase (e.g. Galadini, 1999); or (3) have been averaged on a single direction
360 (Morewood and Roberts, 2000; Roberts and Michetti, 2004) to demonstrate variation of slip-
361 direction along-strike (Roberts and Michetti, 2004). In this work, the crosscutting relationship from
362 field observations (Figs. 4C, 5C, 6C and 7B), and frequency distribution of population of slip-
363 directions (rather than average), has been used as criteria to distinguish between one or two sets
364 of striations. At several sites, we define two distinct sets of slickensides (e.g. sites PP2, M2 and
365 TRM4), that we attribute to distinct episodes of transport directions along the faults. This
366 approach and result are in agreement with earlier study along the Tremonti fault (Galadini, 1999
367 and Piccardi et al., 1999; Smeraglia et al., 2016).

368 Our structural analysis reveals that the faults around the Fucino Basin show normal to transtensive
369 kinematics, and are oriented from NW-SE to NNW-SSE, with the only exception of the Tremonti-
370 Celano-Aielli fault (western tip of the Avezzano-Buzzi system; Fig. 1) which strikes NE-SW. Two
371 main slip-directions can be distinguished: N230° to N260° and N200° to N220°. Few faults also
372 exhibit slip-direction of N120° to N170°. In the northern region, slickenlines crosscutting
373 relationships indicate a coherent pattern of slip-direction change from N240° to N200°
374 (constrained with the M2, PP2 and TRM4 measurement sites, Figs. 4, 5, 6 and Fig. S6; supporting
375 information), in good agreement with previous works (e.g. Piccardi et al., 1999; Smeraglia et al.,
376 2016). Conversely, in the southern regions, slip-directions change from N170° to N220°
377 (constrained with the GdM3 measurement site; Fig. 7 and Fig. S6 supporting information).

378 The vertical displacement on the faults increases, as expected, towards to the center of the Fucino
379 Basin. The southern faults, including the Vallelonga, Aschi Alto and Gioia de Marsi, show an offset
380 of ~500 m. The northern faults, including Campo Felice and Piani Di Pezza faults, have larger
381 offsets ranging between 1000 and 1800 m, except for the Magnola fault, which displays an
382 exceptional vertical throw of ~2500 m. Offset has been estimated by direct field sections and

383 updated 1:50,000 geological maps excluding for Vallelonga, Gioia dei Marsi and Aschi Alto where
384 we used the updated “1:100.000 Foglio Geologico Sora N152” (Praturlon, 1968). Some of our
385 estimates are in accord with previous ones (e.g. Magnola fault; Morewood and Roberts, 2000).
386 Others instead differ from previous ones, as for the case of Vallelonga where we estimate ~300 m
387 (Praturlon, 1968) while others, in the same section, estimated up to ~1400 m (Roberts and
388 Michetti, 2004); or the case of Piani di Pezza fault where we estimate ~1800 m of offset, while
389 Roberts and Michetti (2004) estimated only 400 m (See Fig. S16 in the Supporting Information).

390 The onset of extension in the study area is not contemporaneous on each fault strand. The
391 Tremonti, San Benedetto and Magnola-Velino faults are the oldest structures (Cavinato et al.,
392 2002; Gori et al., 2017), formed probably during the late Pliocene, while faults located on the
393 southern and northern side of the basin formed since the early-middle Pleistocene (Cavinato et al.,
394 2002; Cinti et al., 1992; Giraudi et al., 2011; Giraudi and Frezzotti, 1997; Villani et al., 2015).

395

396 **5.2. Morphometric observations and drainage divide position of the Fucino Basin**

397 The landscape of the area is characterized by deep valleys and steep ranges. The highest values of
398 local relief, slope, and k_{sn} correspond to both extensional and compressive structures, suggesting
399 that tectonics influences the landscape. Evidence of this tectonic control is also shown by the
400 current positions of the drainage divides that are mainly located on the topographic culminations
401 related to thrusts anticline and normal faults, especially on the eastern-north/eastern and on the
402 southwest sides (Fig. 8). Moreover, low values of local relief, slope, and k_{sn} at in summit areas of
403 mountains suggest that these sectors have not been yet reached by the erosion wave induced by
404 the regional uplift likely started around 3 Ma (e.g. Cosentino et al., 2017, D’Agostino et al., 2001b;
405 San Jose et al., 2020).

406 In the Central Apennines, D’Agostino et al. (2001b) proposed that although the Fucino Basin has
407 been internally drained throughout its evolution, the original size has been reduced by the Salto
408 River capturing (Figs. 1 and 8a). Here, the divide is located in an almost flat area (see slope and k_{sn}
409 in Fig. 8d and e) filled by late-Pliocene/early-Pleistocene lacustrine sediments (Centamore, 2004)
410 and by the fluvio-glacial Majelama Valley fan (Frezzotti and Giraudi, 1992). The low energy
411 lacustrine environment is inconsistent with the current base level of the Salto R. (Tyrrhenian Sea),
412 suggesting a southeastward drainage divide migration. The Salto-Fucino divide represents the
413 potential outlet for the Fucino Basin, but the ~30 ka Majelama Fan emplacement (Fig. 2; Frezzotti
414 and Giraudi, 1992) might have prevented the Salto river from capturing the Fucino Basin.

415 The drainage divide stability analysis (Forte and Whipple, 2018; see also Fig. S17; supporting
416 information) indicates an overall stability (as N°1-2-5-6-7 and 8, Fig. 8b) that is apparently
417 surprising as the Fucino is a large endorheic basin at 700 m of elevation. This stability could be
418 interpreted as due to the activity of the faults that keeps the divide high in elevation and contrast
419 the integration of the external rivers into the Fucino Basin. This is due to the fact that the faults
420 deformation rates are higher than the incision rates of the external rivers, pinning the location of
421 the drainage divide. The stability of the Fucino-Salto divide (divide 3; Fig.8b) is preserved by the
422 low stream power of the Salto R. since its upstream portion has not been reached by the erosion
423 related to the onset of the Quaternary regional uplift as evidenced by the very low values of
424 channel steepness (Fig. 8e) and of the local relief (Fig. 8c). The Fucino-Sangro stable divide (divide
425 9 and 10; Fig. 8b) may represent a snapshot of a stability referred to previous boundary
426 conditions. Indeed, the upstream Sangro R. valley is wide and hosts Lower Pleistocene fan deposits
427 scarcely incised (Miccadei et al., 2012), suggesting limited erosion.

428 In summary, morphometric observations of the Fucino Basin indicate that tectonic deformation
429 rates are faster than the erosion rates. This can be also inferred by the fact that the inactive (since
430 middle Pliocene) compressional structures are still shaping the landscape and by the overall stable
431 divides. We suggest that the Salto-Fucino divide migration occurred in recent time (likely since
432 late-Pleistocene, Centamore, 2004), but it has been halted by the Valle Majelama fan (Frezzotti
433 and Giraudi, 1992). Since this upstream portion of the Salto River has not been reached by the
434 erosion relative to the regional uplift, the divides tend to have a very low rate of migration.

435

436 **5.3. Schematic evolution of the Fucino Basin and changes in the direction of extension**

437 Slip-directions, vertical offsets, Holocene slip rates and timing indicate an evolving pattern of
438 deformation around the Fucino Basin that we can summarize in two main stages shown in Fig. 9.

439 Stage 1 (Fig. 9a). The San Benedetto-Ovindoli, Tremonti and Velino-Magnola faults are the oldest
440 structures, and likely formed during the late Pliocene (Cavinato et al., 2002), coeval with the onset
441 of the Sulmona, L'Aquila and the Rieti basins (Cosentino et al., 2017). The kinematics of the
442 Tremonti and Velino-Magnola faults were transtensive to dextral strike-slip (see also Cavinato et
443 al., 2002), acting as transfer of the San Benedetto master fault. The coeval activity is well
444 constrained by the oldest sedimentary sequence that is deposited at the junction between the San
445 Benedetto and the Tremonti fault (Cavinato et al., 2002) (Fig. 9a).

446 Stage 2 (Fig. 9b, c and d). The San Benedetto fault laterally propagates (northward and
447 southward), reactivating older thrust faults, as the case of the Ovindoli-Celano structure (De
448 Vittorio and Faccenna, 1987). During this time, the direction of extension rotated from N240° to
449 N200° (Figure 9b). This new direction of extension (N200°) might have favored ruptures along new
450 fault planes oriented even WNW-ESE, such as the Piani di Pezza or Campo Felice, rather than
451 reactivating the deeper structures more oriented NNW-SSE, such as the Ovindoli or San Benedetto
452 ones. Moreover, this new extensional field (N200°) favors reactivation with dip-slip kinematics of
453 the Velino-Magnola and Tremonti structures which strike instead from WNW-ESE to NNW-SSE
454 (Fig. 9b).

455 We suggest that the timing of the rotations in the slip-direction might have occurred around the
456 middle Pleistocene before the formation of western Piani di Pezza (PP1 site of measurement) and
457 Campo Felice faults that seem to present only the recent ~N200° slip-direction (see station CF3
458 and CF1 from this study, and Morewood and Roberts, 2000; Wilkinson et al., 2015; Fig S7
459 supporting Information). This is also in good agreement with the age of a sedimentary hiatus
460 between two sedimentary sequences within the Fucino Basin (Fig. 2), Pliocene-early Pleistocene
461 and since middle-late Pleistocene in age, related to two separated extensional events (Cavinato et
462 al., 2002).

463

464 **5.4. The Fucino Basin, the Central Apennines, and the role of the sediment loading**

465 The faults that border the Fucino Basin behave differently with respect to the Central Apennines
466 geological context. In particular:

467 1) The most recent N200° direction of extension only occurs in the northern sector of the Fucino
468 Basin. In fact, in the Central Apennines, slip-directions collected by previous studies, show a
469 main extension, oriented NE-SW (Faure Walker et al., 2010). Moreover, stress maps derived
470 from earthquakes and borehole breakouts indicate an overall direction for the minimum
471 horizontal sigma ranging between ENE-WSW to NE-SW (Montone et al., 2012, 1999). This also
472 agrees with a second invariant strain rate from geodesy showing a principal direction oriented
473 NE-SW (Fig. 1a; D'Agostino et al., 2011). Similarly, in the southern Apennines, the current
474 direction of extension is oriented NE-SW (Montone et al., 2012, 1999). This NE-SW direction of
475 extension is also documented by previous slickenlines analysis (e.g. Papanikolaou and Roberts,
476 2007). For this region, the extensional direction over-imposed a previous one oriented NNW-
477 SSE developed since late Miocene (Hippolyte et al., 1994).

478 2) The cumulative offsets across the faults located on the northern side of the Fucino Basin are
479 larger than other faults in the Central Apennines. The vertical throw along the Velino-Magnola
480 and Tremonti faults is ~3.5 km which is higher than the vertical offsets of San Benedetto-
481 Pescina/Collarmele faults or L'Aquila, Sulmona and Rieti faults, that all range between 1.0 and
482 2.5 km (Roberts and Michetti, 2004).

483 These differences between the Fucino and the Central Apennines faults in terms of direction of
484 extension and cumulative vertical displacements could suggest that a local process is occurring at
485 the scale of the Fucino Basin. In fact, comparison between independent data such as vertical
486 geological displacement from this study, morphological displacement and the fault escarpments
487 height from literature (Benedetti et al., 2013; Schlagenhauf et al., 2011; Villani et al., 2015;
488 Wilkinson et al., 2015), indicates a remarkable increase of the values from north to south, i.e.
489 towards the Fucino Basin (Fig. 10). The only exception to this pattern is the Velino vertical
490 geological displacement which is lower than the northern faults. In this context, we note that only
491 the Velino fault exhibits a gentle fault dip measured on the surface (<45°; see Table S1 and Fig. S3;
492 supporting information). The overall pattern shown in Fig. 10 strongly accords with Holocene slip-
493 rates that also increases toward the basin, reaching maximum values along the Velino-Magnola
494 fault (Benedetti et al., 2013). Hence, our and previous data indicate that the northern structures of
495 the Fucino Basin are characterized by larger and faster displacement than the southern ones.

496 We propose that sedimentary (un)loading plays an important role in the internal dynamics of the
497 Fucino Basin, influencing the deformation pattern. On a first approximation, the sedimentary
498 (un)loading over the faults may be estimated assuming local Airy isostasy (Fig. S18; Supporting
499 information; England and Molnar, 1990). For the Fucino Basin, assuming a sediment thickness of
500 ~1 km (Cavinato et al., 2002) with density of 1700 kg/m³, we estimate ~400 m of subsidence of the
501 San Benedetto hanging wall (Fig. 11 and Fig. S19; Supporting information). For the Magnola sector,
502 assuming ~1200 meters as a minimum estimation of thickness of the eroded material (Wise,
503 1985), we estimate maximum 1000 m of footwall uplift (Supporting information, Fig. S20).
504 Assuming flexural isostasy (see details in supporting information) and using elastic thicknesses (Te)
505 of 2, 4 and 6 km (cfr. with Te of 3.7 km; D'Agostino and McKenzie, 1999) we generally obtain, as
506 expected, lower values (see Figs. S21 and S22; Supporting information). For the hanging wall
507 subsidence, we estimate ~369 (Te= 2 km), ~223 (Te= 4 km) and ~165 m (Te= 6 km) while for the
508 footwall uplift ~425 (Te= 2 km), ~150 (Te= 4 km) and ~81 m (Te= 6 km). These isostatic vertical

509 movements due to the (un)loading adjustment represent almost the 30% of the total geological
510 vertical slip.

511 The sedimentary loading may also produce a local rotation of slip-direction, from N240° to N200°.
512 According to Anderson regime (Anderson, 1951), the increase in vertical stress due to sediment
513 loading, which for the strike/oblique-slip is the intermediate stress (σ_2 ; Anderson, 1951), may be
514 enough to turn vertical stress into the principal one (Fig. 11b) (σ_1 ; Anderson, 1951). This would
515 result in a re-organization of the stresses. As discussed in section 4.3, we suggest that the change
516 in the slip-direction, from N240° to N200°, and in kinematics has occurred during middle
517 Pleistocene, according to the hiatus related to an extensional phase and to the regional uplift
518 between two sequences within the Fucino Basin (Cavinato et al., 2002). This implies that the older
519 sequence 3 was related to an extensional event oriented N240°, while the younger sequence 4
520 was deposited during an extensional event oriented N200°. Based on the thickness of the two
521 sequences, we estimate that, since the onset of the deformation, the total vertical stress
522 incremented by ~15 MPa. According to this speculation, only ~4 MPa (estimated accounting only
523 the thickness of sequence 4; Cavinato et al., 2002) were needed to turn the kinematic for
524 strike/oblique- to dip-slip, as the case of the Tremonti fault (Fig. 11b).

525

526 **5.5. Feedback between tectonics, sediment loading and erosion**

527 The feedback between tectonics and sediment (un)loading might have a fundamental impact on
528 the evolution of deformation. In extensional settings, hanging wall sedimentary load increases
529 faults slip (Hampel et al., 2009, 2007; Turpeinen et al., 2008) by up to 15% (Maniatis et al., 2009).
530 This base level lowering as a result of fault slip increase, might increase the river slope profile and
531 sediment production, resulting in an increased loading of the hanging wall. Conversely, in
532 compressional settings, the sediment load over the hanging wall may delay thrust slip (Ballato et
533 al., 2019; Hampel et al., 2009). Other studies reported that sediments or ice sheets (un)loading
534 influence or even trigger seismicity (Hampel et al., 2009). Erosional unloading may gently flexure
535 the upper crust, reducing the normal stress sufficiently to reactivate pre-existing structures, as
536 occurred in the New Madrid seismic zone (United States) (Calais et al., 2010; Hampel et al., 2007;
537 Hampel and Hetzel, 2006; Hetzel and Hampel, 2005; Steer et al., 2014; Vernant et al., 2013).
538 In this context, the development of internally drained basins, and therefore the resultant storage
539 of remarkable volumes of sediment, may exert a significant control on the kinematics of
540 deformation, even at the scale of the orogens (e.g. Royden, 1996; Willett and Beaumont, 1994).

541 Generally, endorheic-exorheic or exorheic-endorheic transitions are the result of a competition
542 between faulting, headward erosion and climate (e.g. Berry et al., 2019; D'Agostino et al., 2001b).
543 Although this is theoretically applicable in several geological contexts, it is important to
544 differentiate the formation of significantly large endorheic basins at the scale of mountain wedges
545 and the formation of small intermontane endorheic basins, as for example the case of the
546 Apennines. At the scale of an orogen, where significant crustal shortening is occurring, the
547 development of topographic barrier in the frontal range may limit the precipitation behind,
548 significantly reducing fluvial incision. This, in turn, may develop large internally drained basins,
549 such as on Puna-Altiplano Plateau, the Tibetan Plateau, and the Tarim basin (e.g. Sobel et al.,
550 2003). The accumulation of kilometers of sediments at high-elevation basin forces the
551 deformation to shift toward the margins of the orogeny, significantly enlarging the wedge. In turn,
552 this produces a lowering in slope, relief and therefore fluvial incision, favoring the preservation
553 through time of the endorheic configuration (Royden, 1996; Willett, 1999; Willett and Beaumont,
554 1994).

555 At smaller scales, such as the Apennines, where the shadow effect of precipitation can be
556 neglected, endorheic-exorheic or exorheic-endorheic transitions are a common process mainly
557 due to the competition between faulting and fluvial headward erosion. If fluvial incision is faster
558 than footwall uplift along a fault, the rivers might cut across the fault, capturing a basin drainage
559 system (Fig. 12a). If the slip or footwall uplift rate become faster than the fluvial incision, the basin
560 becomes endorheic (Fig. 12a). The surface processes are affected by climate variations, that locally
561 might be the main controlling factor in decreasing in the fluvial incision. In our study area, this
562 process may have occurred in the Campo Felice or Piani di Pezza basins, where wind gaps
563 (Forcamiccia and Valico Di Pezza; Figs. 12 and 4A) suggest a previous exorheic configuration (Fig.
564 12a; Giraudi and Frezzotti, 1997; Giraudi et al., 2011). In any case, the small internally drainage
565 condition, especially for intermontane extensional basins, is not permanent and it is commonly
566 interrupted by a new capture from a nearby catchment.

567 If at the onset of deformation, the fault slip rate is faster than the erosion rate, the basin will be
568 endorheic (Fig. 12b). If the fluvial incision of adjacent basins becomes faster than the slip-rate, the
569 basin might be captured (Fig. 12b). A second scenario to explain the endorheic-exorheic transition
570 is the one associate to overspill processes (e.g. Geurts et al., 2019, 2018). This last model requires
571 that the amount of water and sediment load accumulated within the internally drained basin rises
572 to overflow the topographic barrier triggering the incision. In any case, in both scenarios

573 (endorheic-exorheic transition due to headward erosion or overspill), this new externally drainage
574 configuration might be permanent since, once captured, the increased discharge makes the fluvial
575 incision rates much faster than vertical movement of the faults (D'Agostino et al., 2001b). In the
576 Central Apennines an example of endorheic-exorheic transition is the Sulmona basin, where the
577 Popoli Gorges cut through the Monte Morrone, linking the Sulmona basin with the Adriatic Sea
578 (D'Agostino et al., 2001b; Geurts et al., 2019; Fig. 1 and 12b).

579 Interaction between erosion and faulting implies that during their evolution, the basins might be
580 characterized by both endorheic and exorheic phases, making the permanent endorheic nature, as
581 the case of the Fucino Basin, non-sustainable.

582

583 **5.6. Why is the endorheic nature of the Fucino Basin preserved?**

584 We propose a conceptual model that considers the original structural setting of the Fucino Basin.
585 We attribute the endorheic nature of the Fucino Basin to the coeval presence of master (San
586 Benedetto) and transfer faults (Tremonti and Velino-Magnola). We propose that the presence of
587 these two orthogonal faults favors the endorheic drainage stage since the onset of extension (Fig.
588 13).

589 In our model for the preservation of the internally drained configuration since the onset of
590 extension and through time the two following conditions are strongly tied to each other: (1) the
591 sediment load and (2) the combination of the two orthogonal sets of faults.

592 1) Our work indicates that sediment-related vertical geological displacements only
593 represent up to 30% of cumulative geologic throw across the San Benedetto-Pescina-
594 Collarmele and the Magnola faults. This non-extensional vertical displacement, in turn,
595 increases the Fucino base level drop-down, enhancing erosion processes at higher
596 elevation surface. At the same time, the enlarged accommodation space developed by
597 faulting, allows to store more sediments overloading the faults hanging wall, inducing a
598 self-sustaining positive feedback process.

599 2) The coexistence of two orthogonal faults (Tremonti-Celano Aielli Vs San Benetto-
600 Collarmele-Pescina) represents the main difference between the Fucino Basin and all the
601 other extensional basins in the Central Apennines (Fig. 1). We suggest that this is the key
602 for the preservation of the endorheic nature of the Fucino Basin through time. Our
603 structural analysis shows a re-orientation along the transfer faults in the stress from N240°
604 to N200°, from strike/oblique- to dip-slip. We proposed that this local change in the

605 direction of extension and kinematic along the transfer faults might be related to the
606 sediment loading. This increment of vertical stress may turn the vertical intermediate
607 stress (σ_2) into the principal stress (σ_1 ; Anderson, 1951; Fig. 11). In turn, this kinematic
608 changing along the original transfer faults favors the vertical displacement and subsidence
609 on both master and transfer structures, self-triggering this process as described in point
610 (1).

611 We suggest that the sediment-related additional vertical movement is one of the controlling
612 factors for the permanent endorheic configuration of the Fucino Basin. It is worth to note that the
613 Fucino Basin is located in a region that experiences both regional extension and uplift (San Jose et
614 al., 2020). The uplift is fundamental to create, with the local footwall uplift, a topographic barrier
615 against the nearby drainage basins.

616 In conclusion, we suggest that the competition between the sediment loading and the two
617 orthogonal sets of faults that self-sustain each other with a positive feedback, could explain why
618 the Fucino Basin remained endorheic through time. This increased subsidence prevents the basin
619 to be captured by nearby catchments while the orthogonal set of structures that boarder the basin
620 keeps the internal rivers to drain outside (Fig. 13).

621

622 **6. Conclusions**

623 We presented a detailed structural analysis of the Fucino Basin, estimating the vertical geological
624 displacement of the main faults surrounding the region and the expected effects on vertical
625 displacement using both local and flexural isostasy. We propose a conceptual model to explain the
626 peculiar and permanent endorheic nature of the Fucino Basin.

627 The vertical geological offset of the main faults ranges from 500 to 2500 m increasing from north
628 to south toward the Fucino Basin. Our estimates indicate that up to 30% of these geological offsets
629 are due to sediment (un)loading. We also suggest that the consequent significant increase in
630 vertical stress may lead to a re-organization of the kinematics, from oblique- to dip-slip, especially
631 along the transfer faults such as Tremonti and Magnola faults. This change in the kinematics
632 produces a local rotation in the main slip-direction from N240° to N200°.

633 These findings allowed us to propose a conceptual model to better explain the endorheic nature
634 of the Fucino Basin through time. We proposed that the combination of two orthogonal families of
635 faults (San Benedetto and Tremonti) and the increasing slip, related to the sediment loading,
636 prevents the Fucino Basin to be captured by nearby catchments. In turn, this process produces a

637 positive feedback increasing the sediment overloading the hanging wall, which in turn increases
638 the slip of the fault, preserving the internally drainage through time.

639

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645

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862 **Figure Captions**

863

864 **Figure 1:** a) Topographic map of the Central Apennines with schematic traces of normal and thrust faults b)
865 Topographic map of the Central Apennines with the river network and the main rivers.

866

867 **Figure 2:** Simplified geological map of the study area (see Fig. 1 for location) with in black the cross-section

868 traces. At the bottom, schematic geological sections modified after Cavinato et al. (2002) and from
869 'Foglio Geologico Sora 152' (Praturlon, 1968).

870

871 **Figure 3:** Simplified structural map of the study area (see Fig. 1 for location) with the location of the
872 measurement sites. The black arrows indicate the direction of movement of the hanging wall with
873 respect to the footwall for each site. A merged stereoplot is shown for each fault with the hanging
874 wall slip-direction (blue arrows).

875

876 **Figure 4:** A) Panoramic view of the Piani Di Pezza Fault. B and C) Detail of two slickenlines directions and
877 cross cutting relationship on PP2 measurement site (Fig. 3). D) Stereoplot of the fault planes and
878 slickenlines contour of PP2 measurement site.

879

880 **Figure 5:** A) Panoramic view of the Magnola Fault. B) Detail of the M2 measurement site (Fig. 3). C) Detail of
881 two slickenlines directions and cross cutting relationship on M2 measurement site (Fig. 3). D)
882 Stereoplot of the fault planes and slickenlines contour of M2 measurement site.

883

884 **Figure 6:** A) Panoramic view of the Tremonti Fault. B) Detail of the TRM4 measurement site (Fig.3). C) Detail
885 of two slickenlines directions and cross cutting relationship on TRM4 measurement site (Fig. 3). D)
886 Stereoplot of the fault planes and slickenlines contour of TRM4 measurement site.

887

888 **Figure 7:** A) Panoramic view of the Gioia dei Marsi Fault. B) Detail of the GdM1 measurement site (Fig. 3). C)
889 Detail of two slickenlines directions and cross cutting relationship on GdM3 measurement site (Fig. 3).
890 D) Stereoplot of the fault planes and slickenlines contour of GdM3 measurement site.

891

892 **Figure 8:** a) Topographic map of the study area (left) and two swath profiles (right). The white boxes
893 indicate the swath profile locations and the numbers the divide sections included in the divide
894 stability analysis. b) Plot showing the results of the drainage divide stability analysis following Forte
895 and Whipple (2018). c) Local relief map with 5 km of average circular window. d) Slope map. e) Map of
896 the along channel variation of k_{sn} .

897

898 **Figure 9:** Schematic evolutionary model from late Pliocene to Holocene, inspired by Cavinato et al. (2002).

899

900 **Figure 10:** Simplified structural map of the study area (see Fig.1 for location) with the fault labeled based on
901 a) escarpment height (short-term displacement; see Puliti et al. (2020) for detail) from previous
902 studies (see main text for references), b) Morphological displacement (mid-term displacement; see

903 Puliti et al. (2020) for detail) from previous studies (see main text for references) and c) geological
904 throw (long-term displacement; see Puliti et al. (2020) for detail) from this work. In the lower plot
905 displacements over faults.

906

907 **Figure 11:** The role of the sediments in the deformation. a) Sediments thickness from Cavinato et al. (2002)
908 on the left and subsidence calculated using local isostasy model from Molnar and England, (1990). In
909 the lower part topographic section: current topography, bedrock depth and expected bedrock depth
910 without sediment loading. b) Kinematic changing and stress re-organization on the transfer fault
911 (Tremonti Fault).

912

913 **Figure 12:** Conceptual model for endorheic-exorheic and exorheic-endorheic transition for small
914 intermontane basins. a) Evolution from exorheic to endorheic with a panoramic view of Campo Felice
915 basin. b) Evolution from endorheic to exorheic with a panoramic view of Sulmona basin.

916

917 **Figure 13:** Schematic and conceptual model for the endorheic preservation through time of the Fucino
918 basin.

919

Figure1.

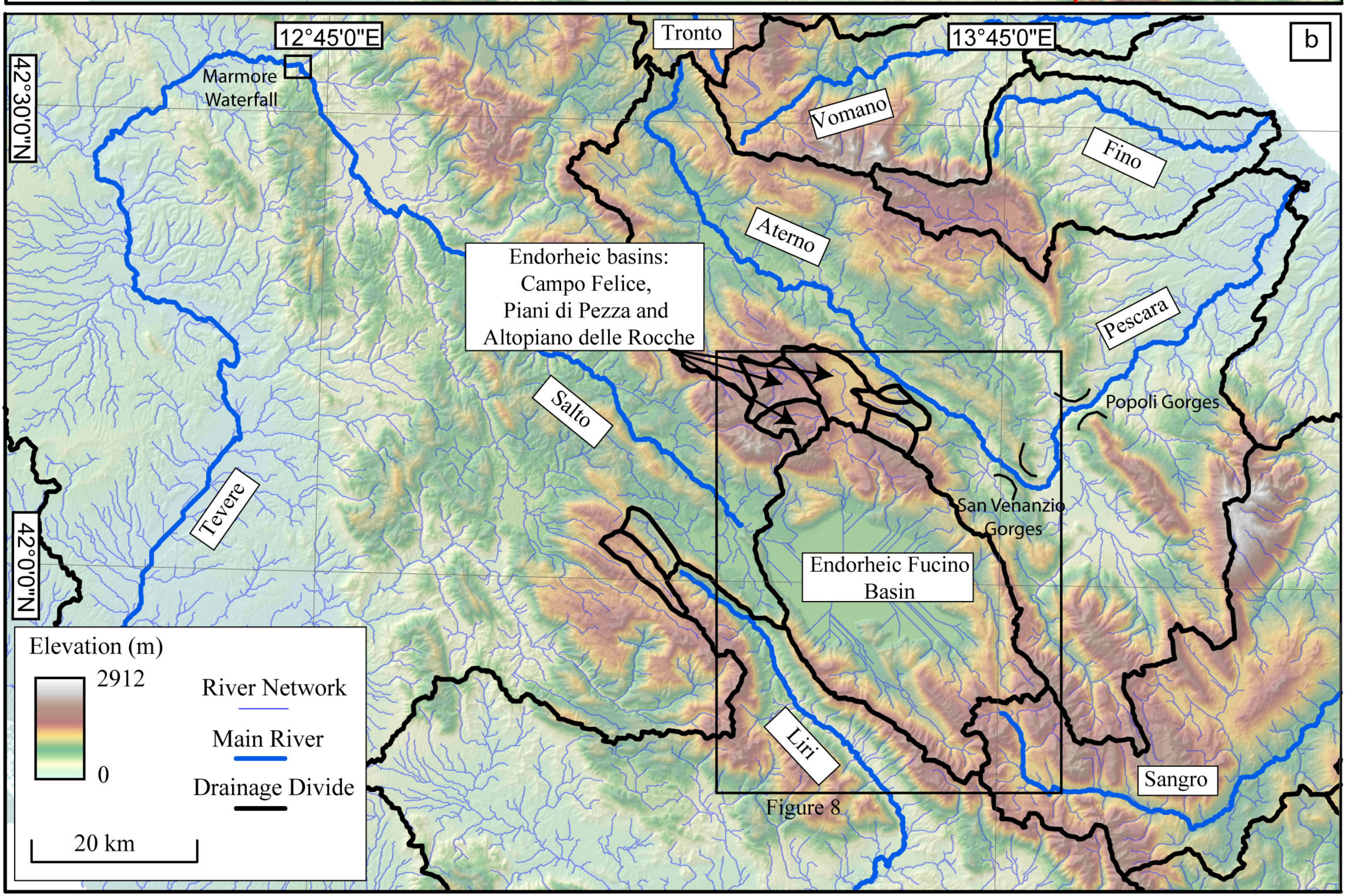
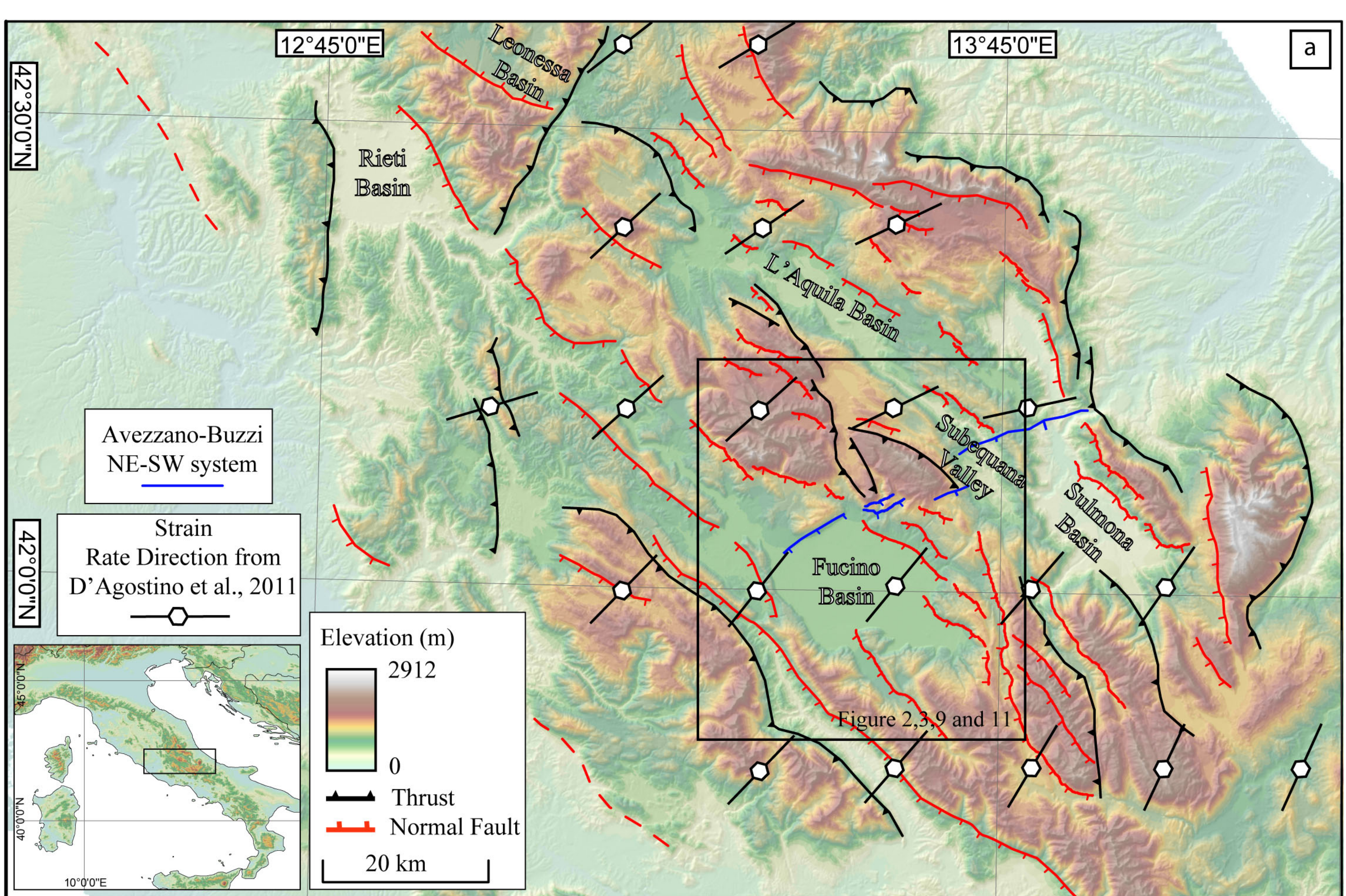


Figure2.

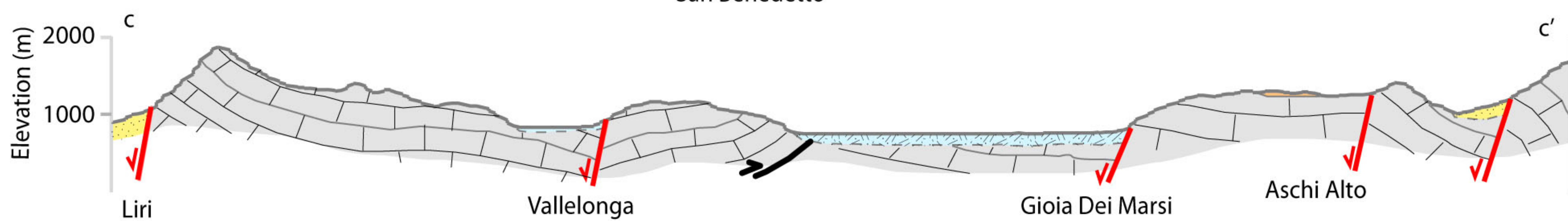
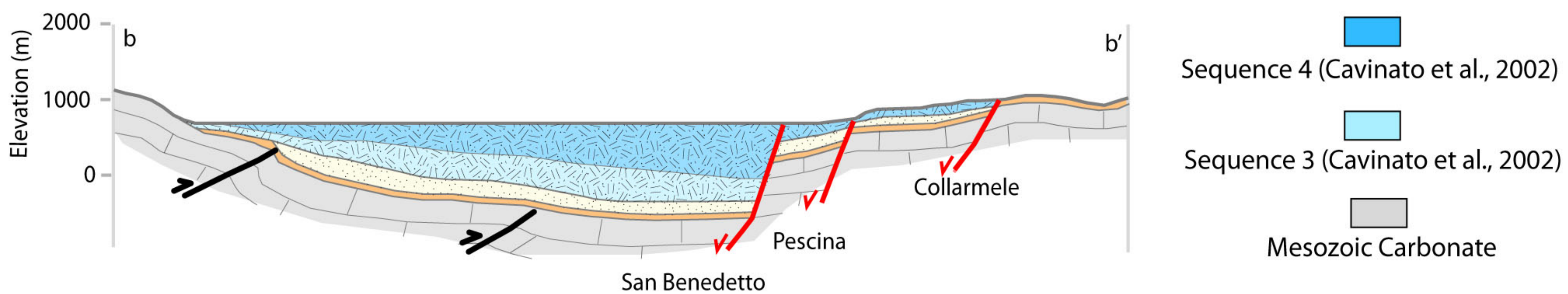
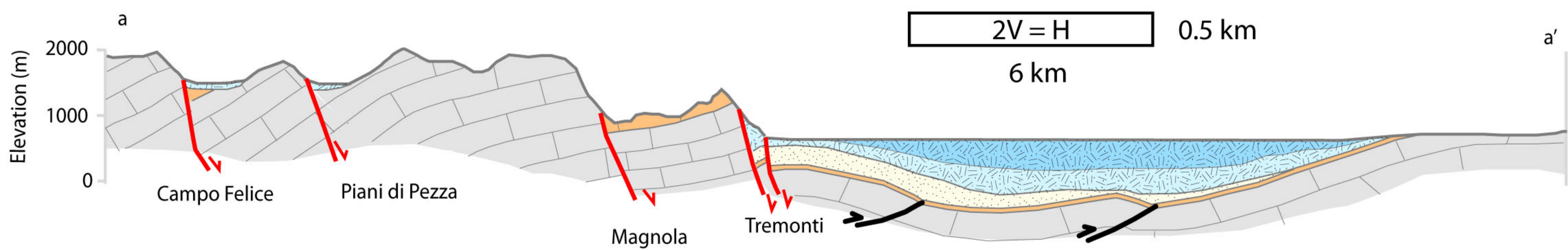
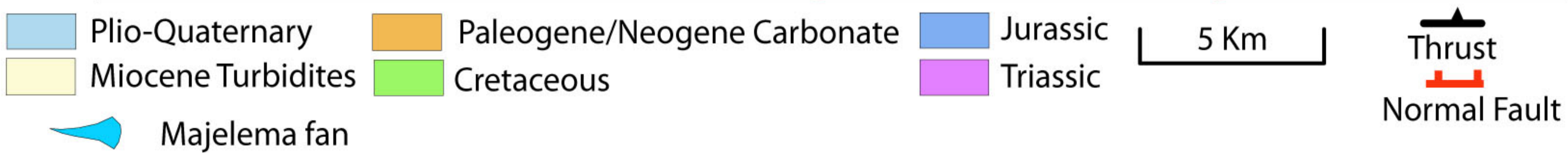
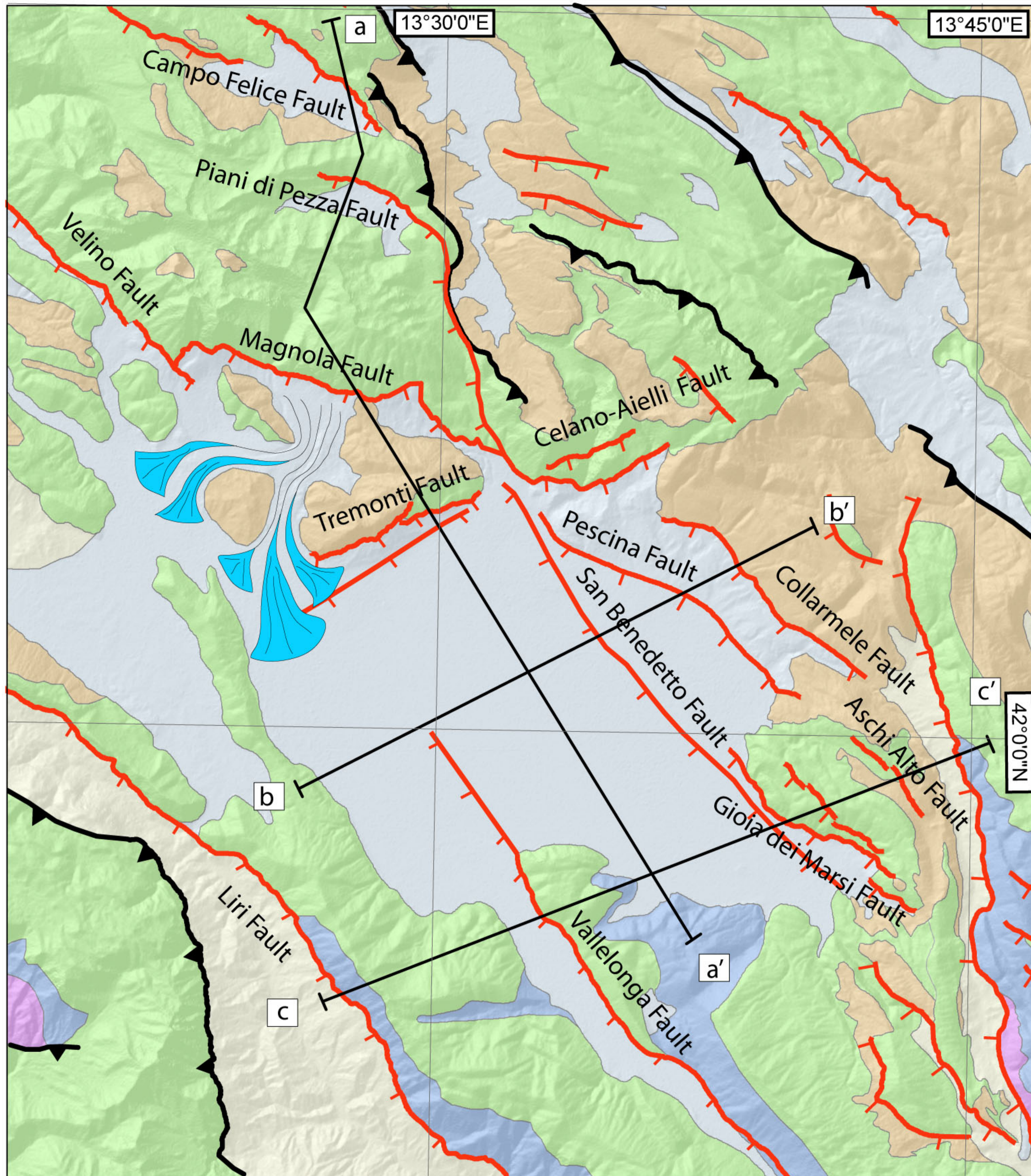
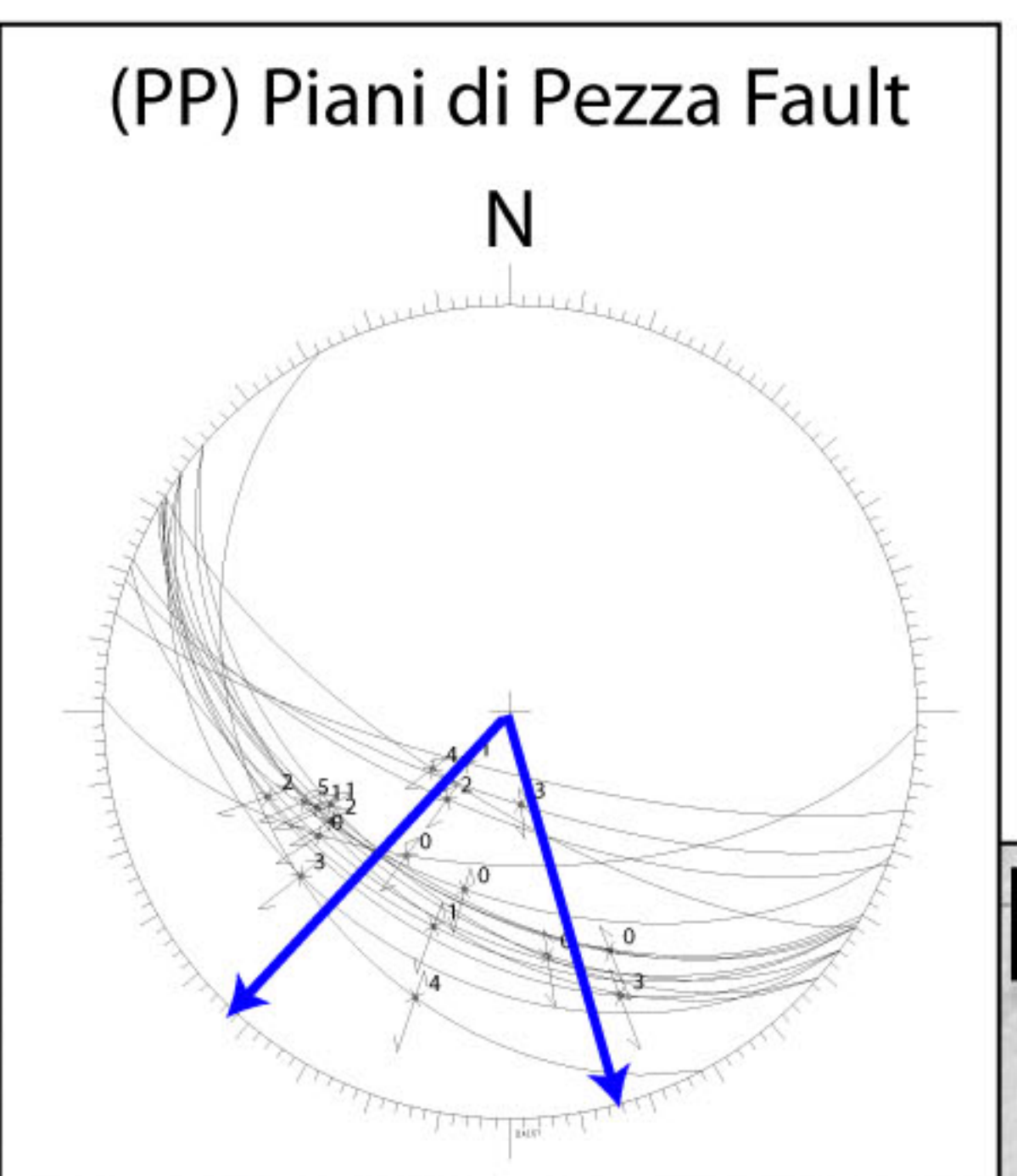
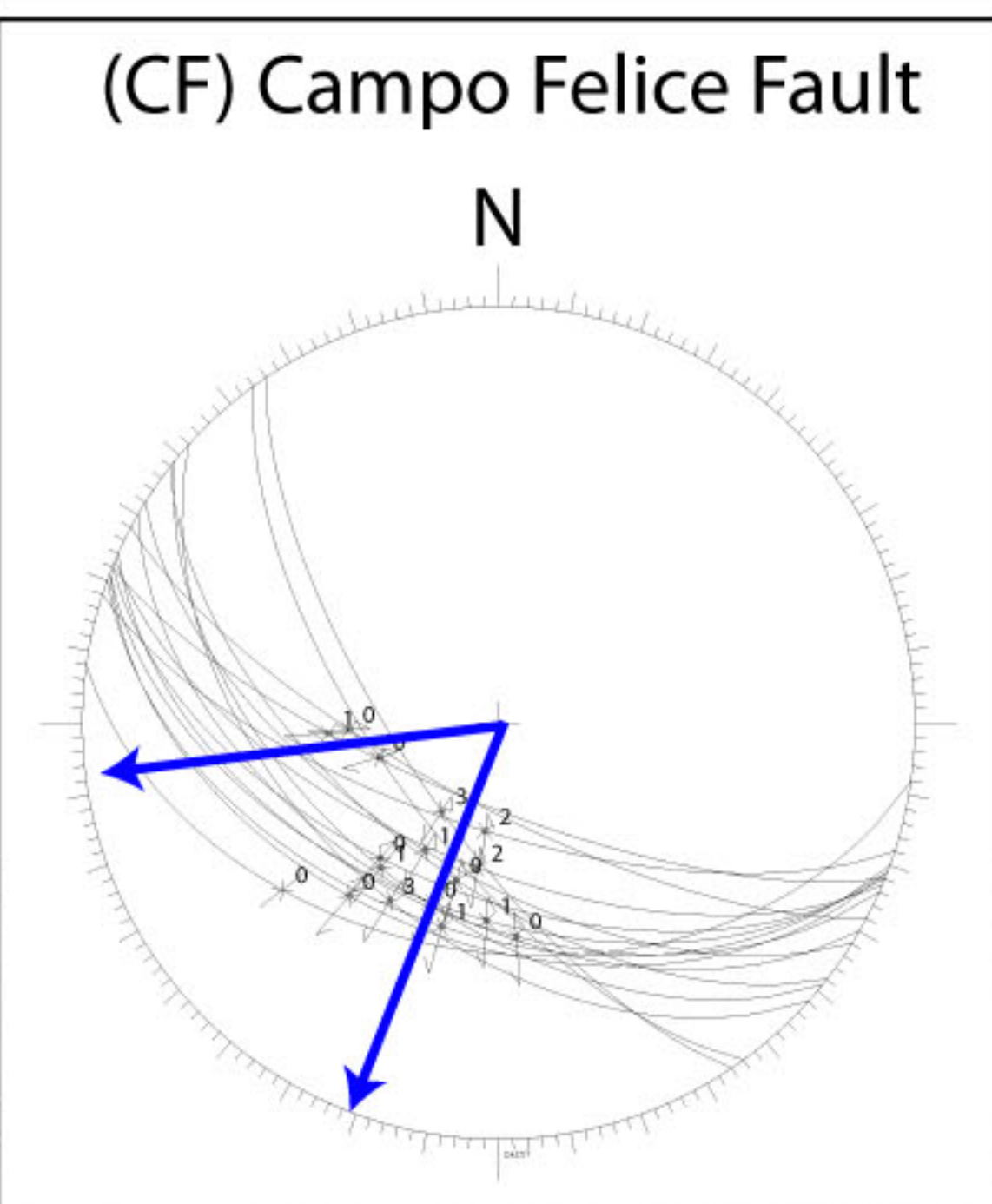
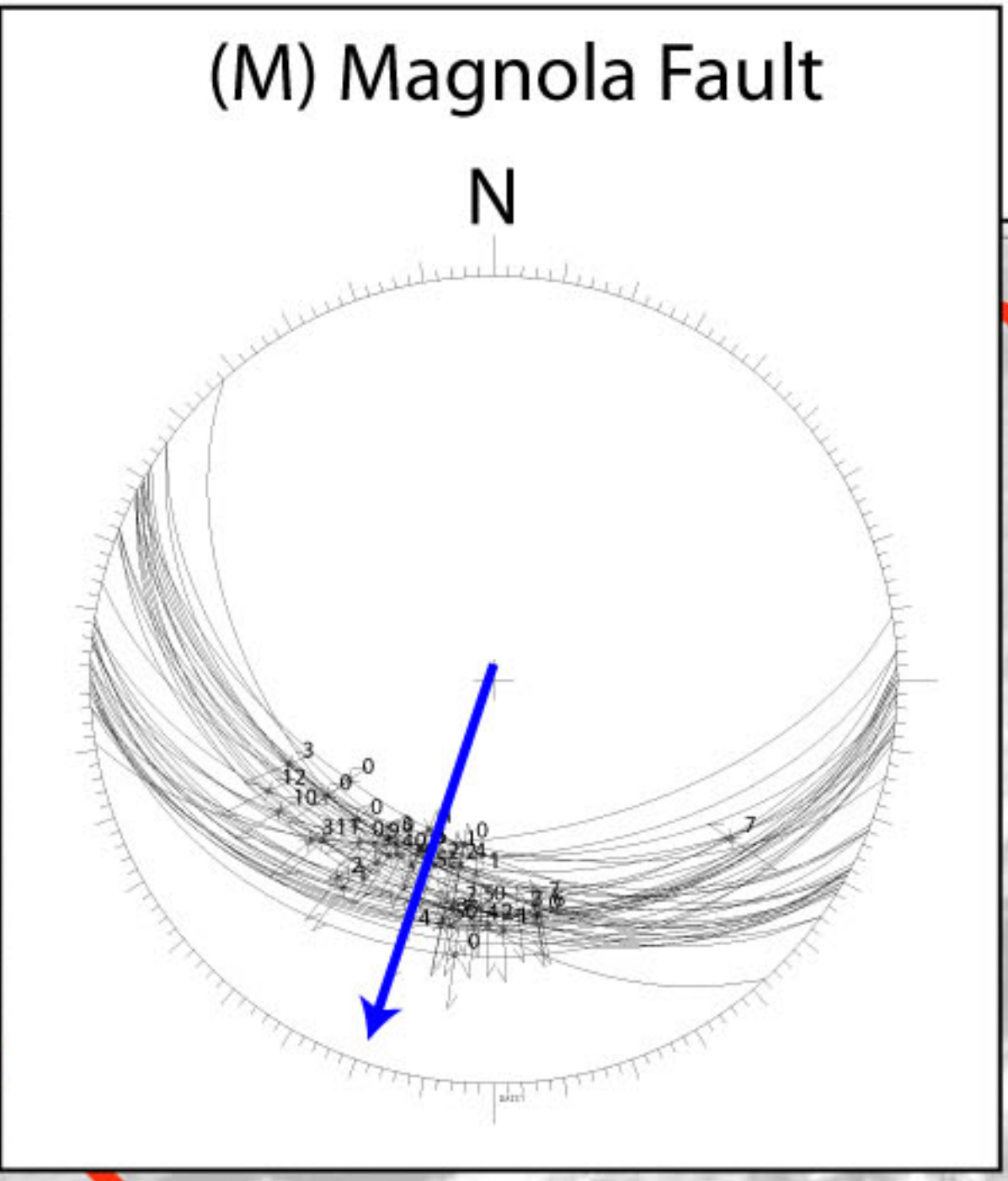
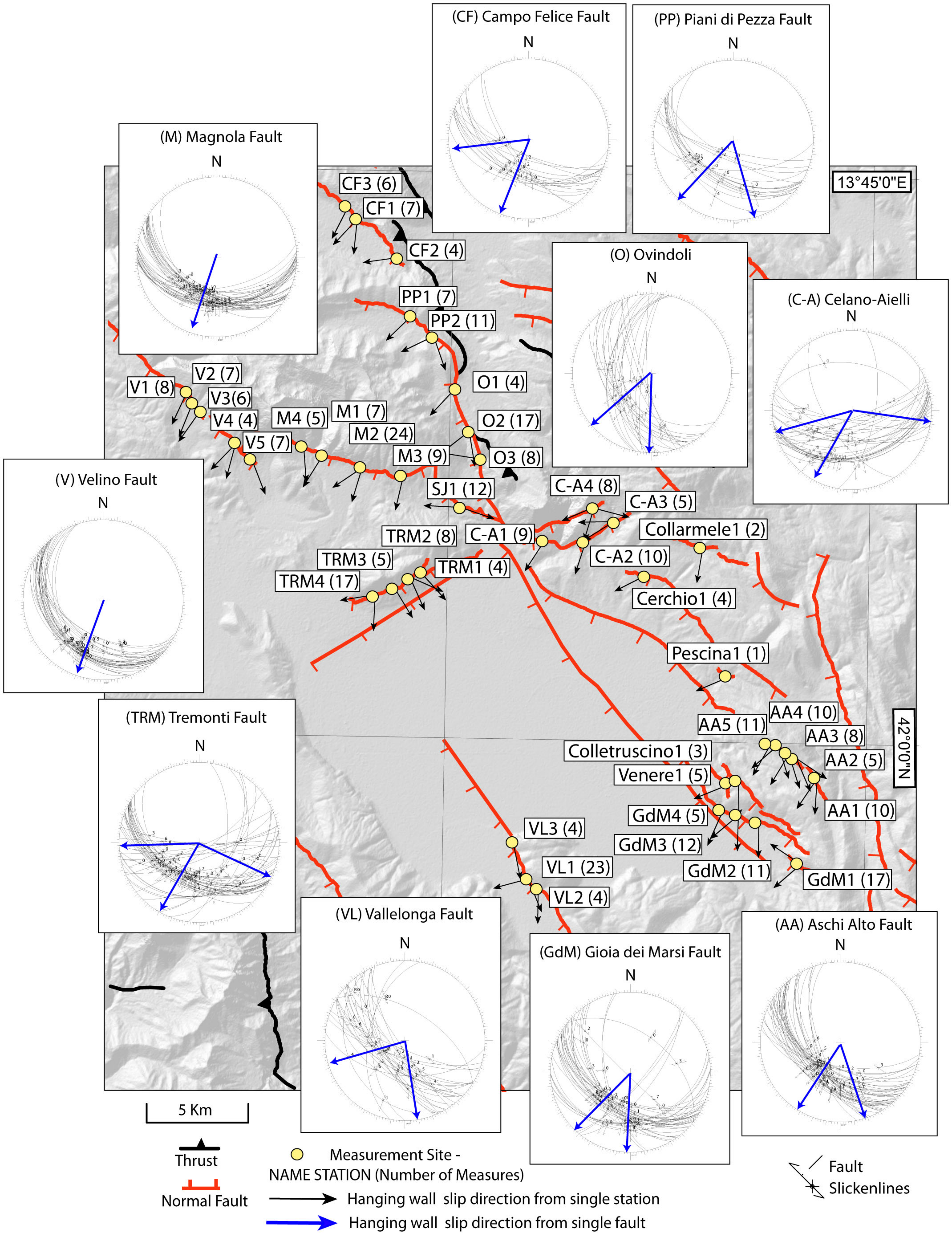
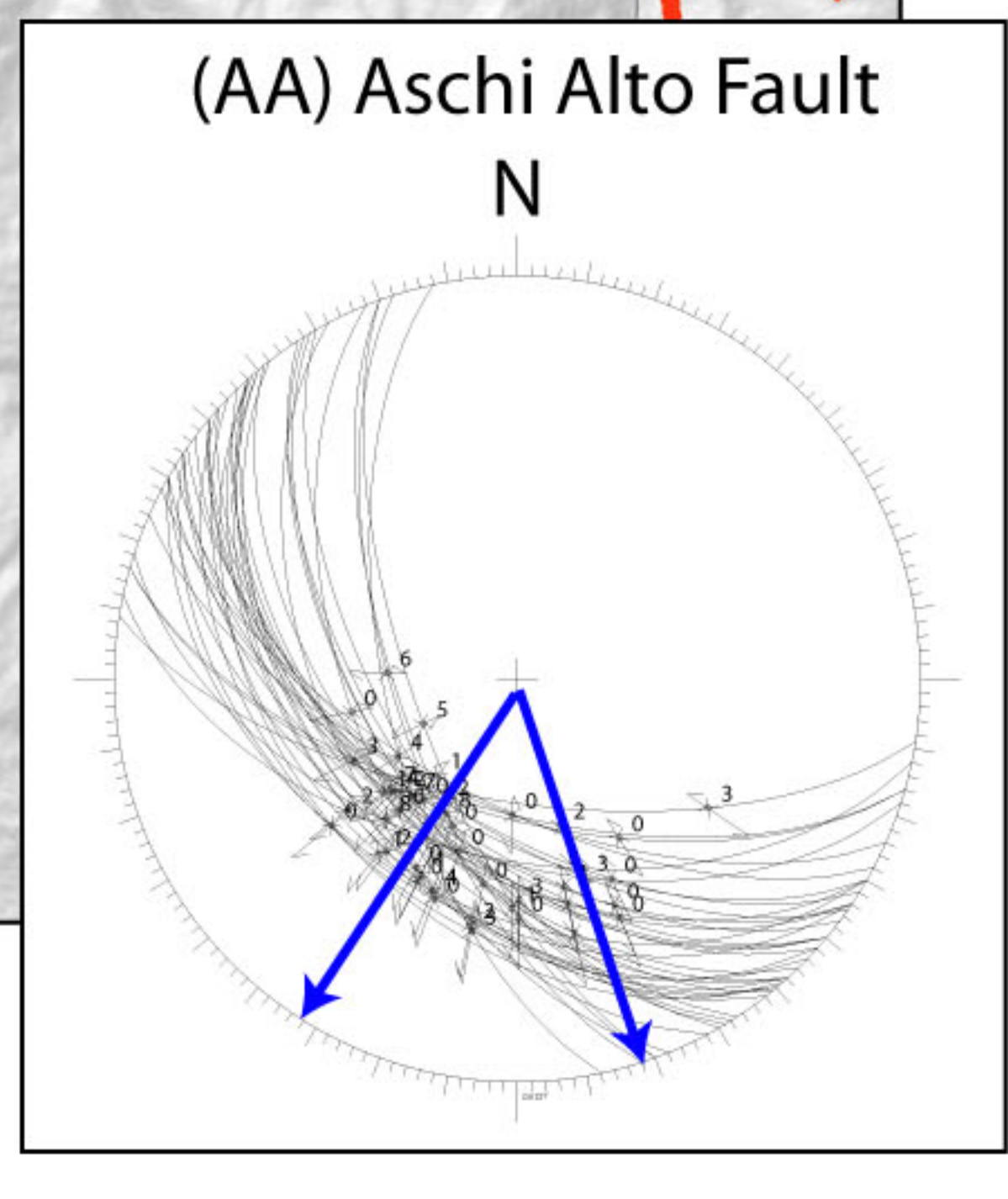
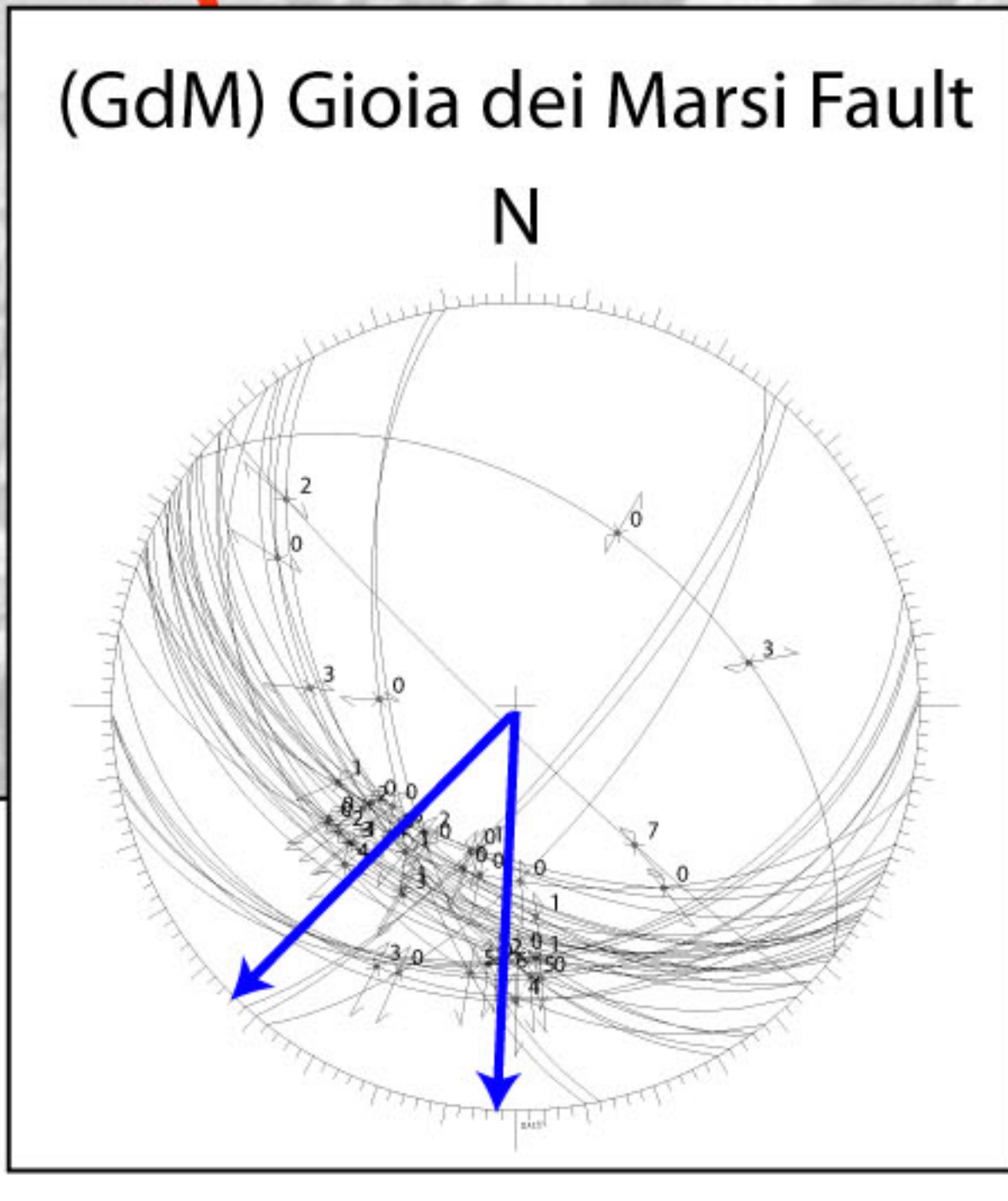
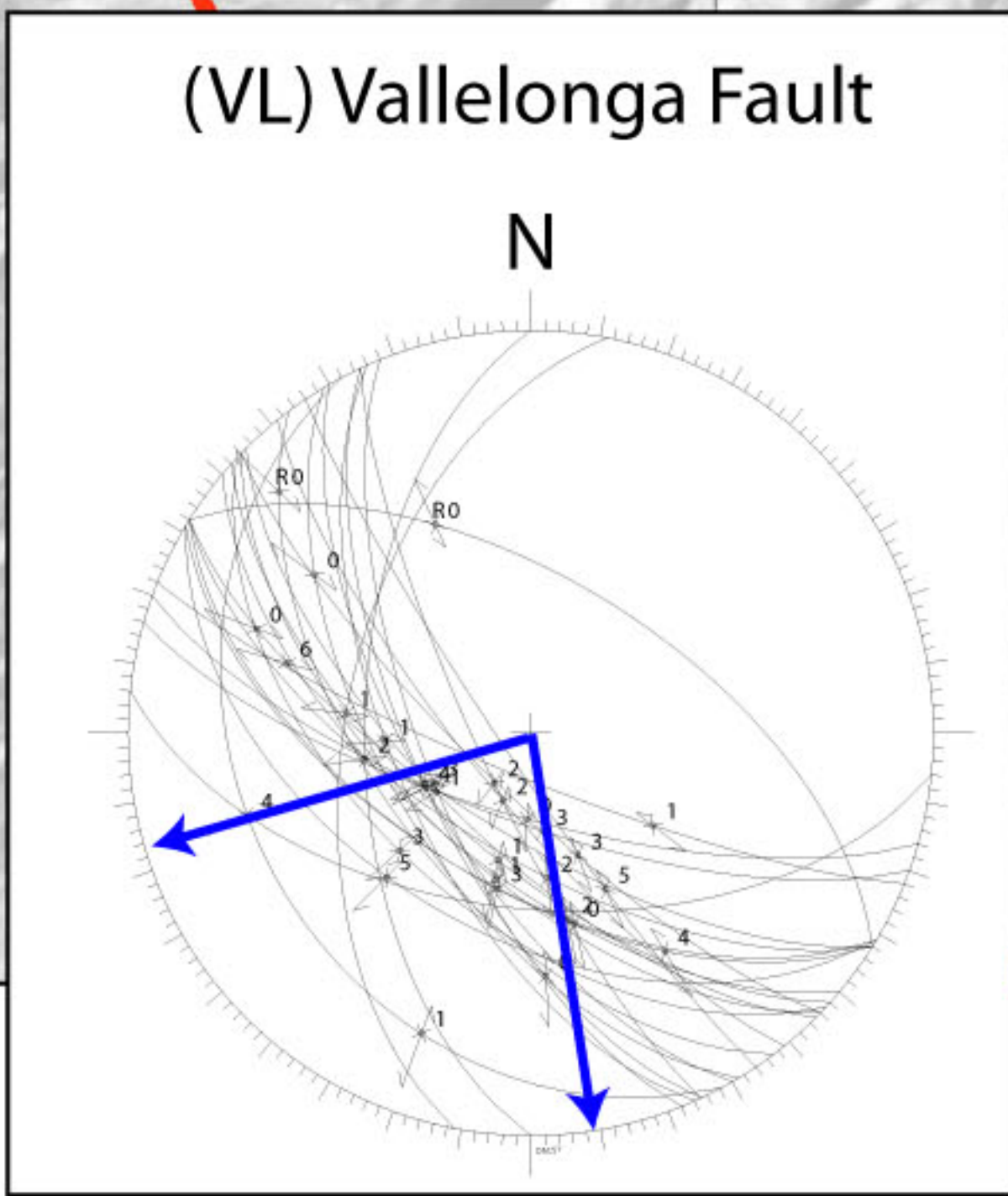
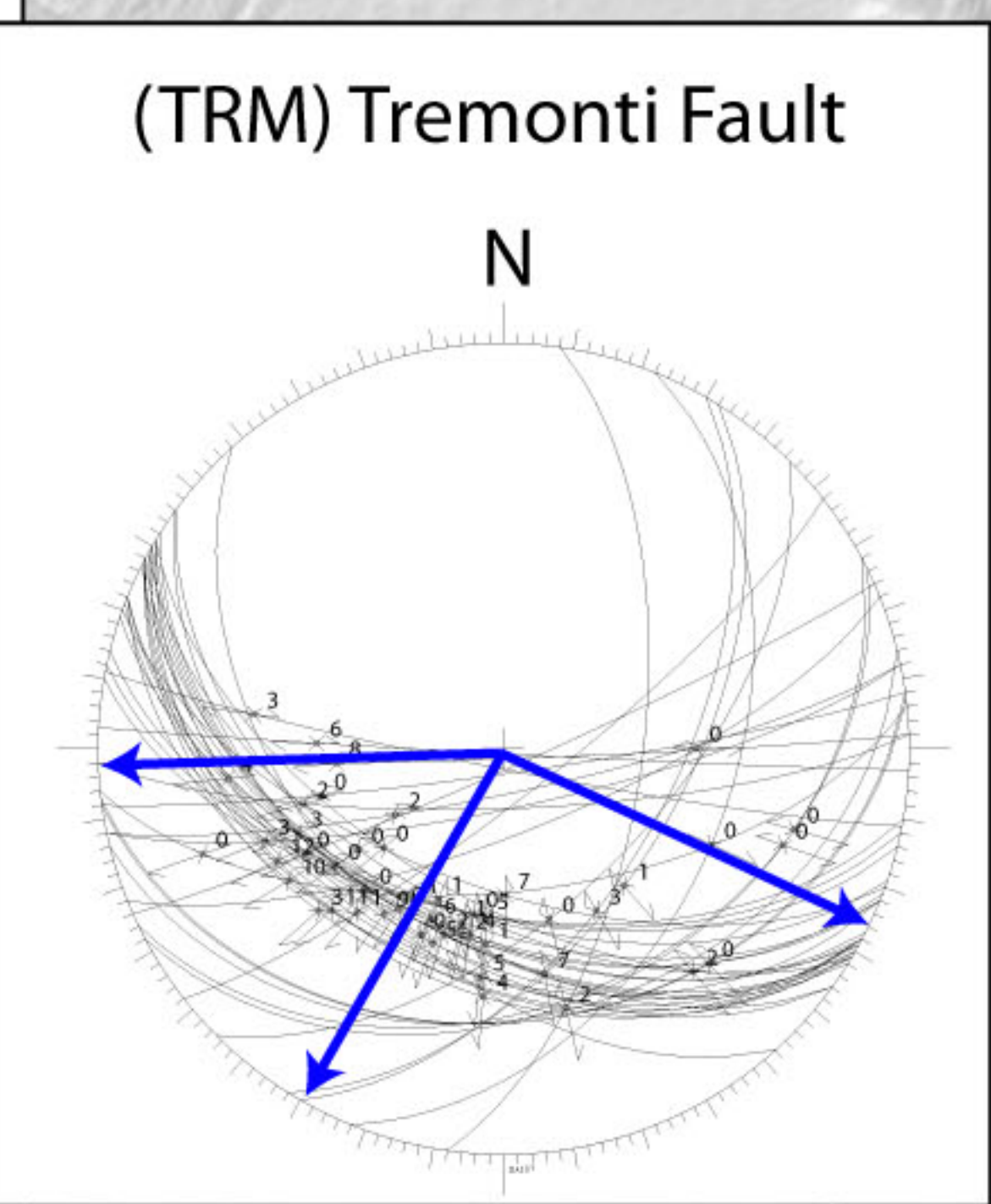
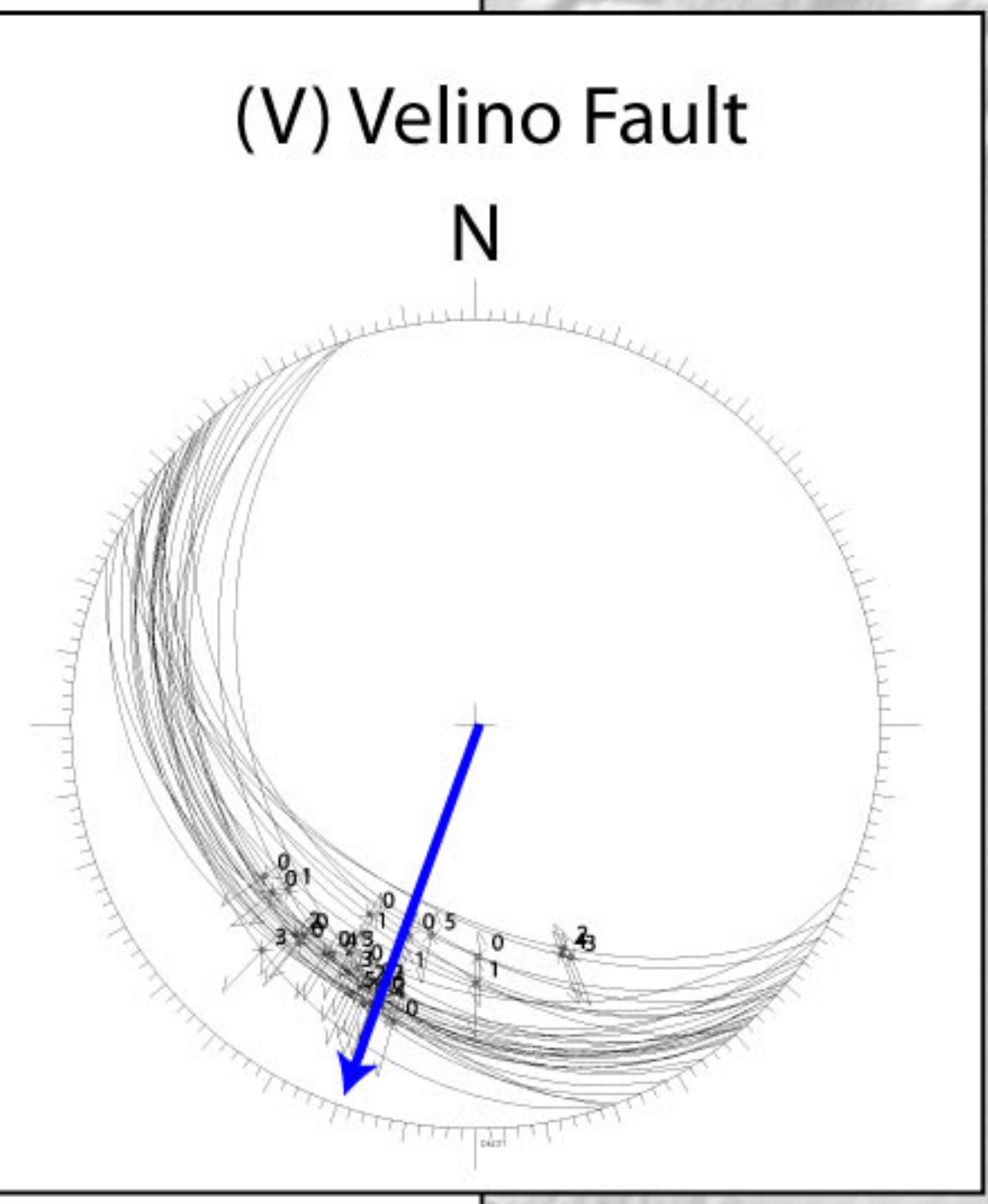
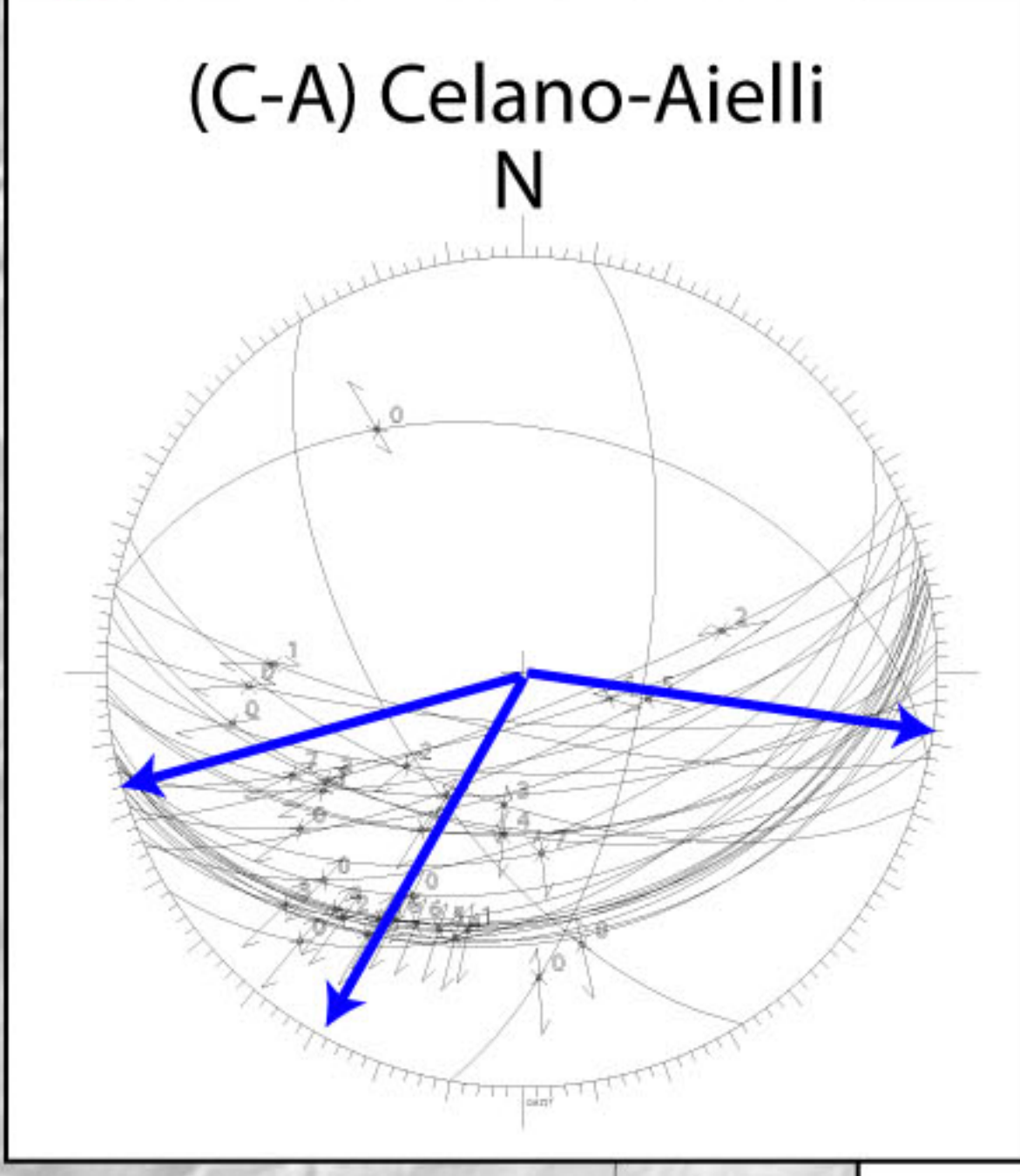
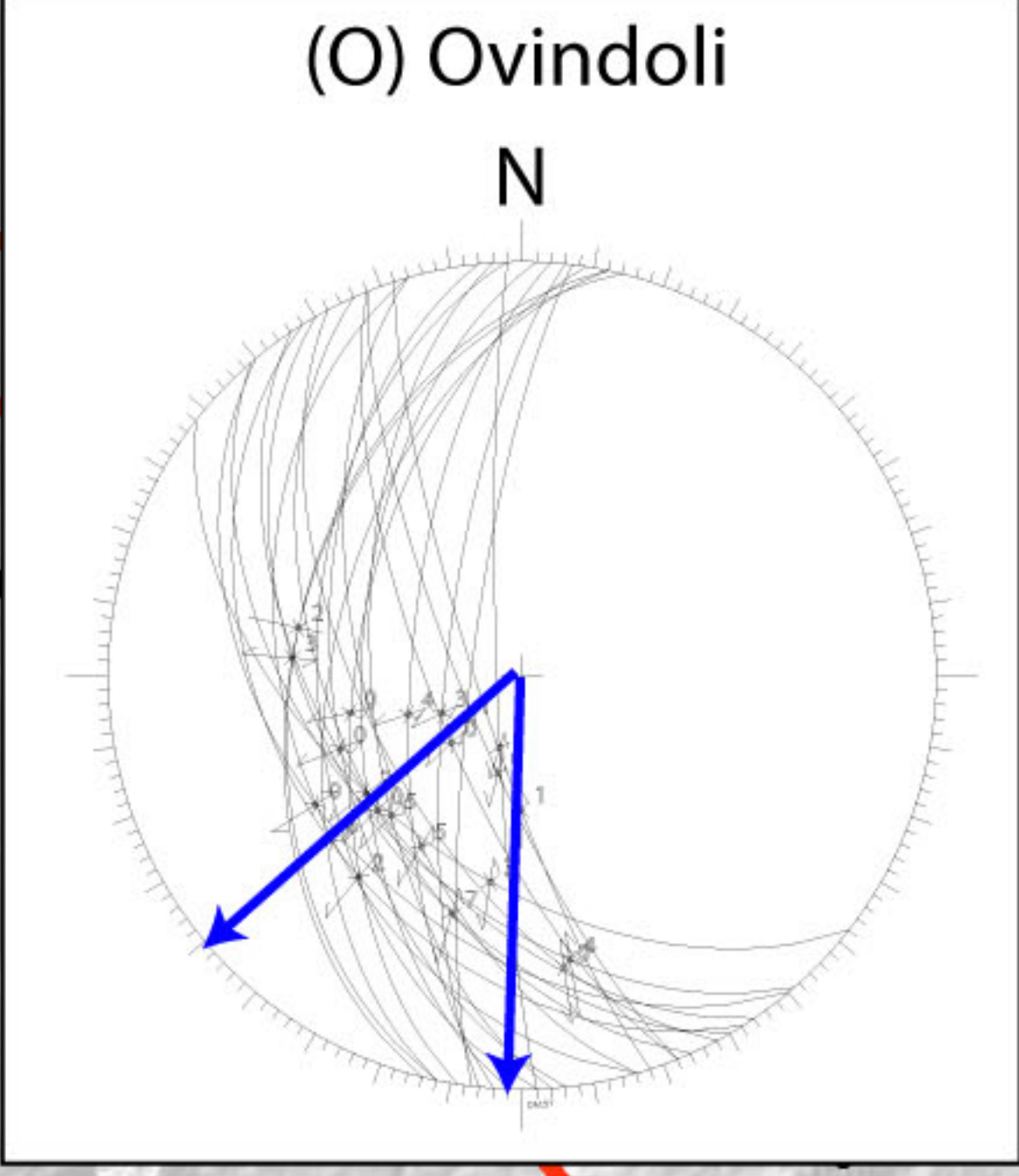


Figure3.



13°45'0"E



42°0'0"N

5 Km

Thrust
Normal Fault

● Measurement Site -
NAME STATION (Number of Measures)
→ Hanging wall slip direction from single station
→ Hanging wall slip direction from single fault

Fault
Slickenlines

Figure4.

A Piani di Pezza Fault

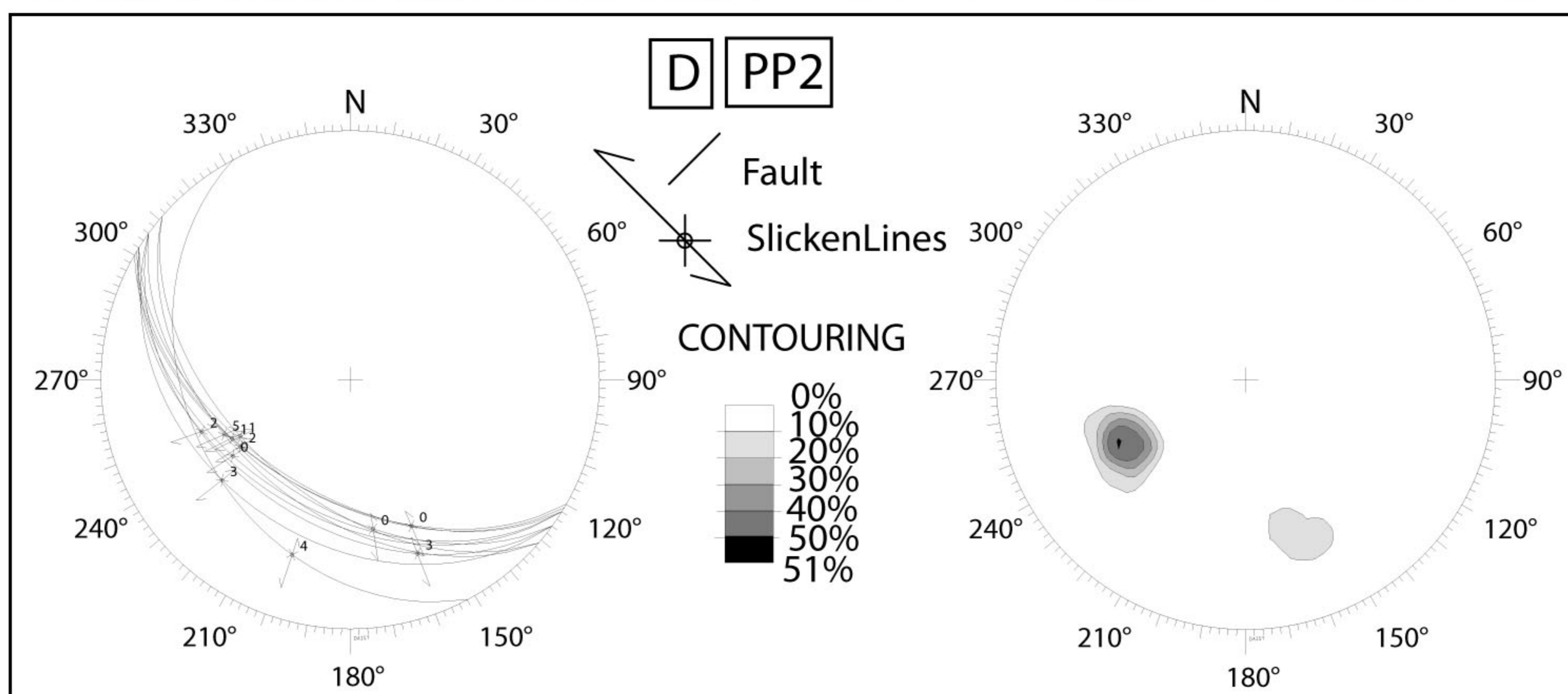
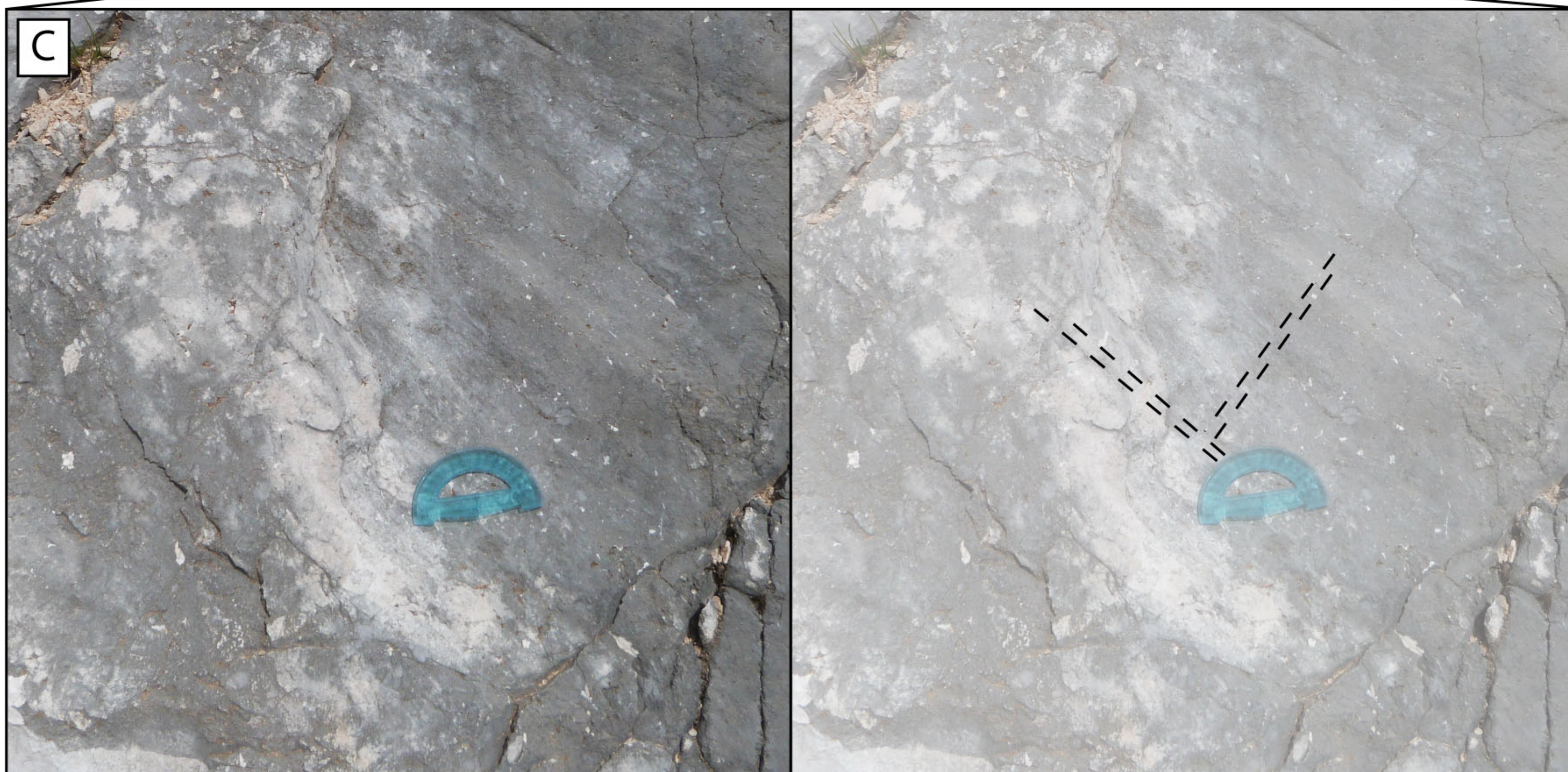
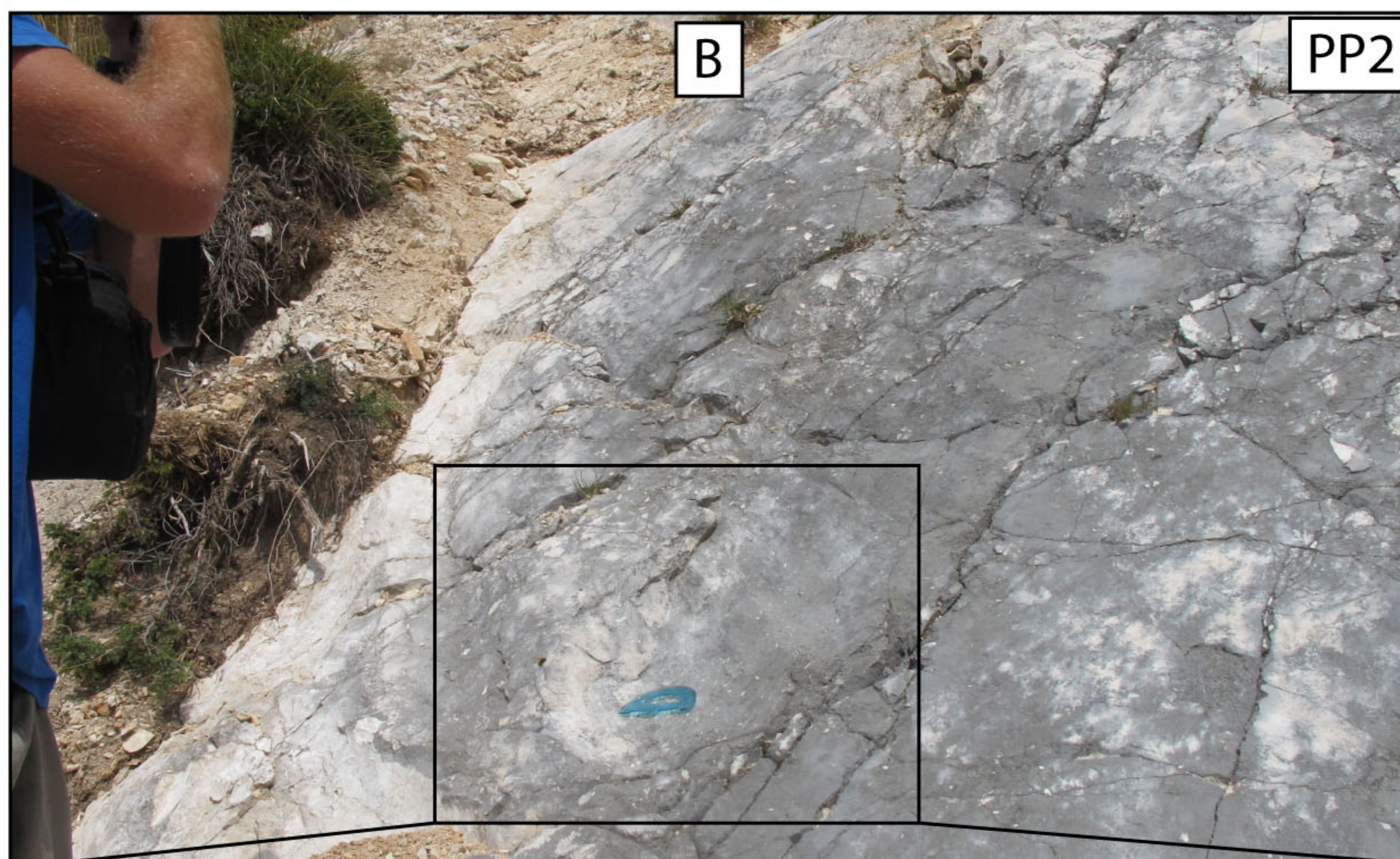
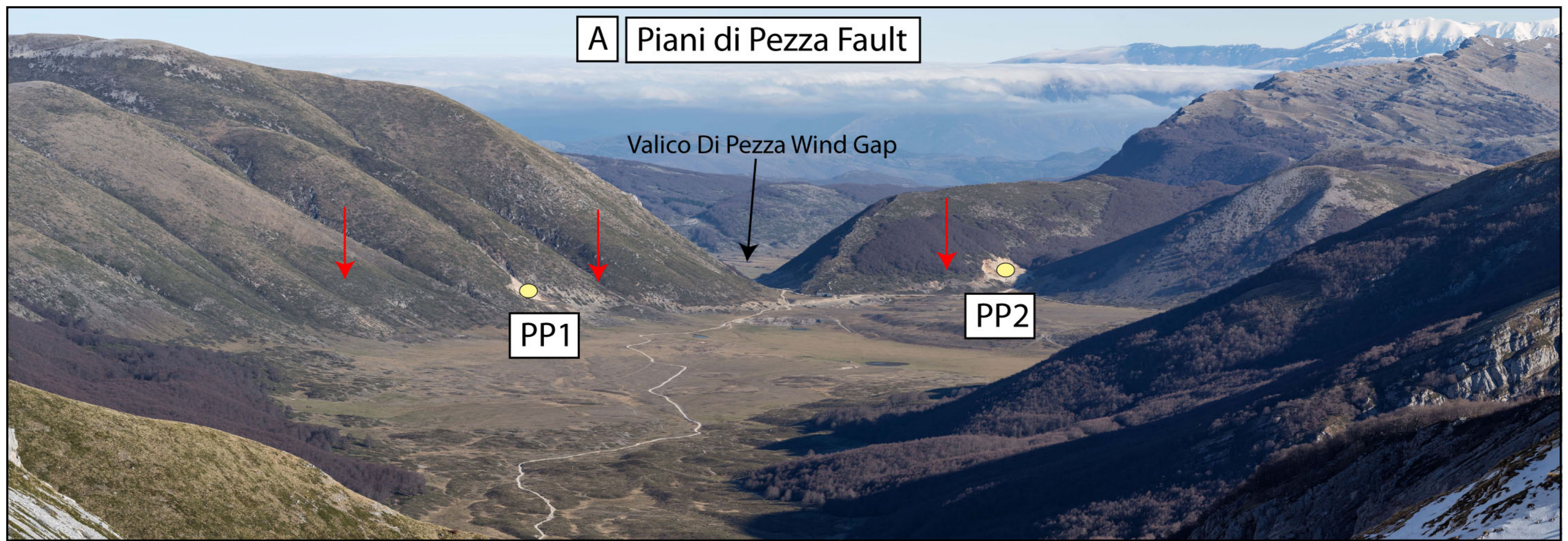


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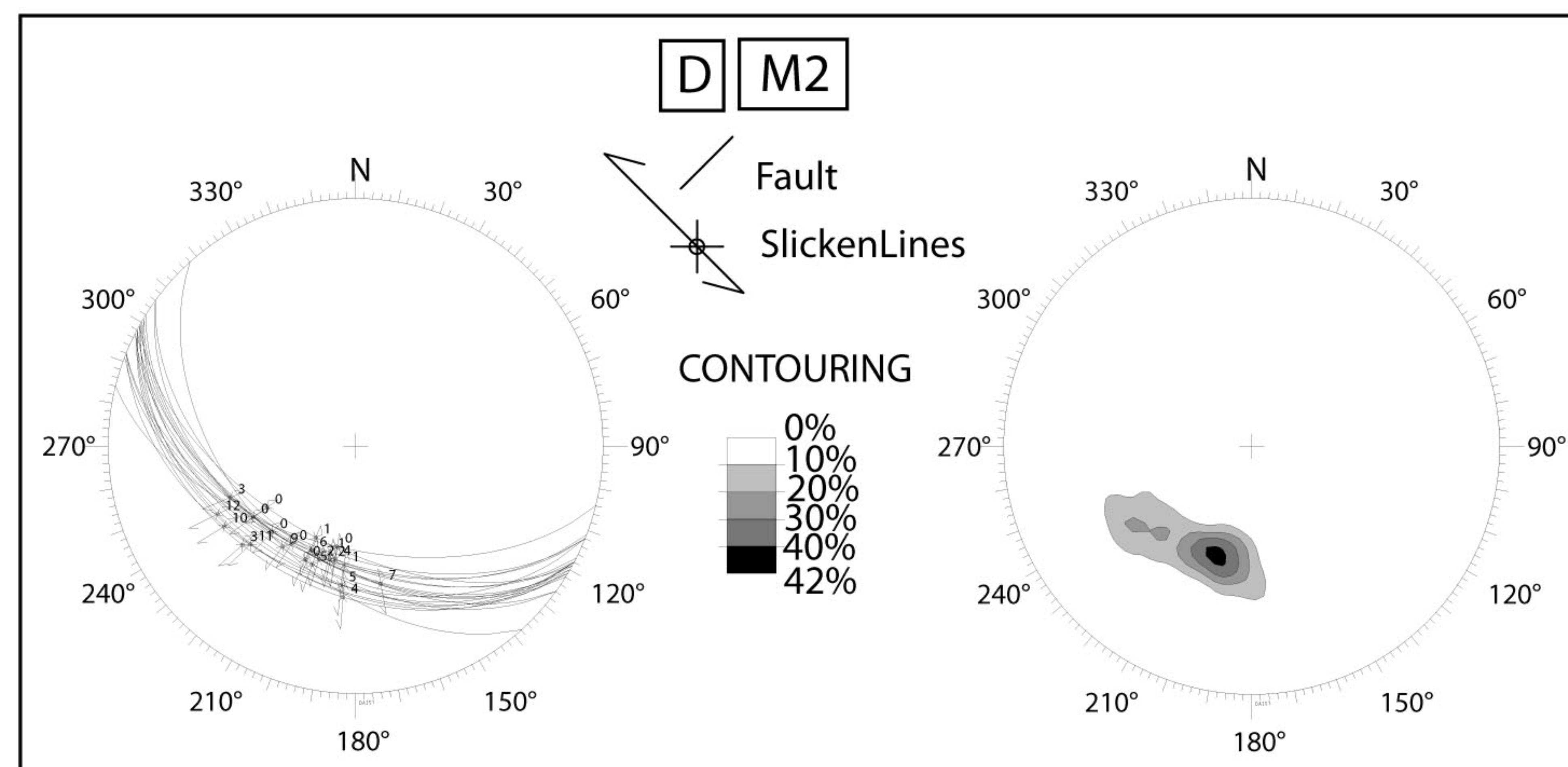
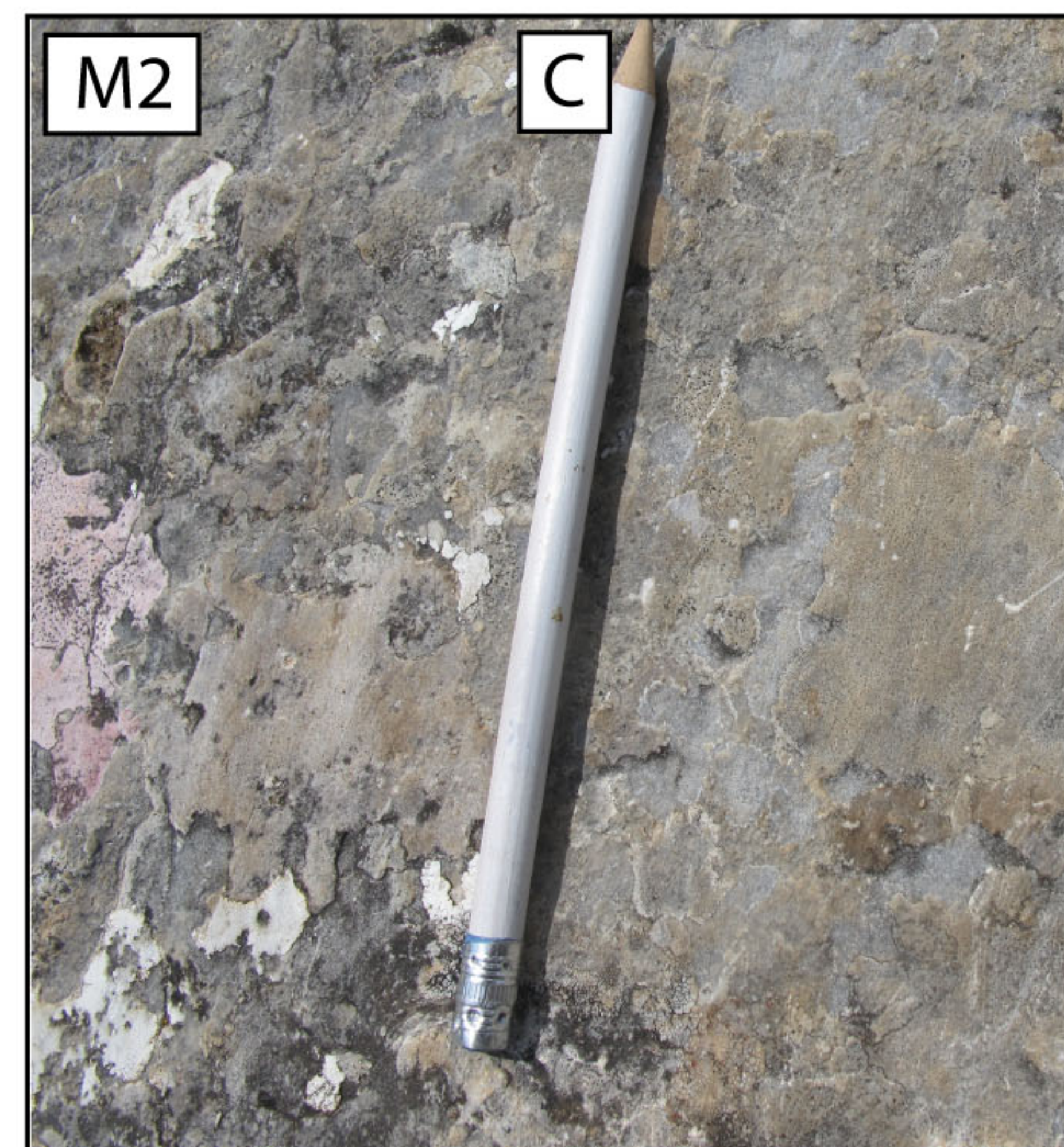
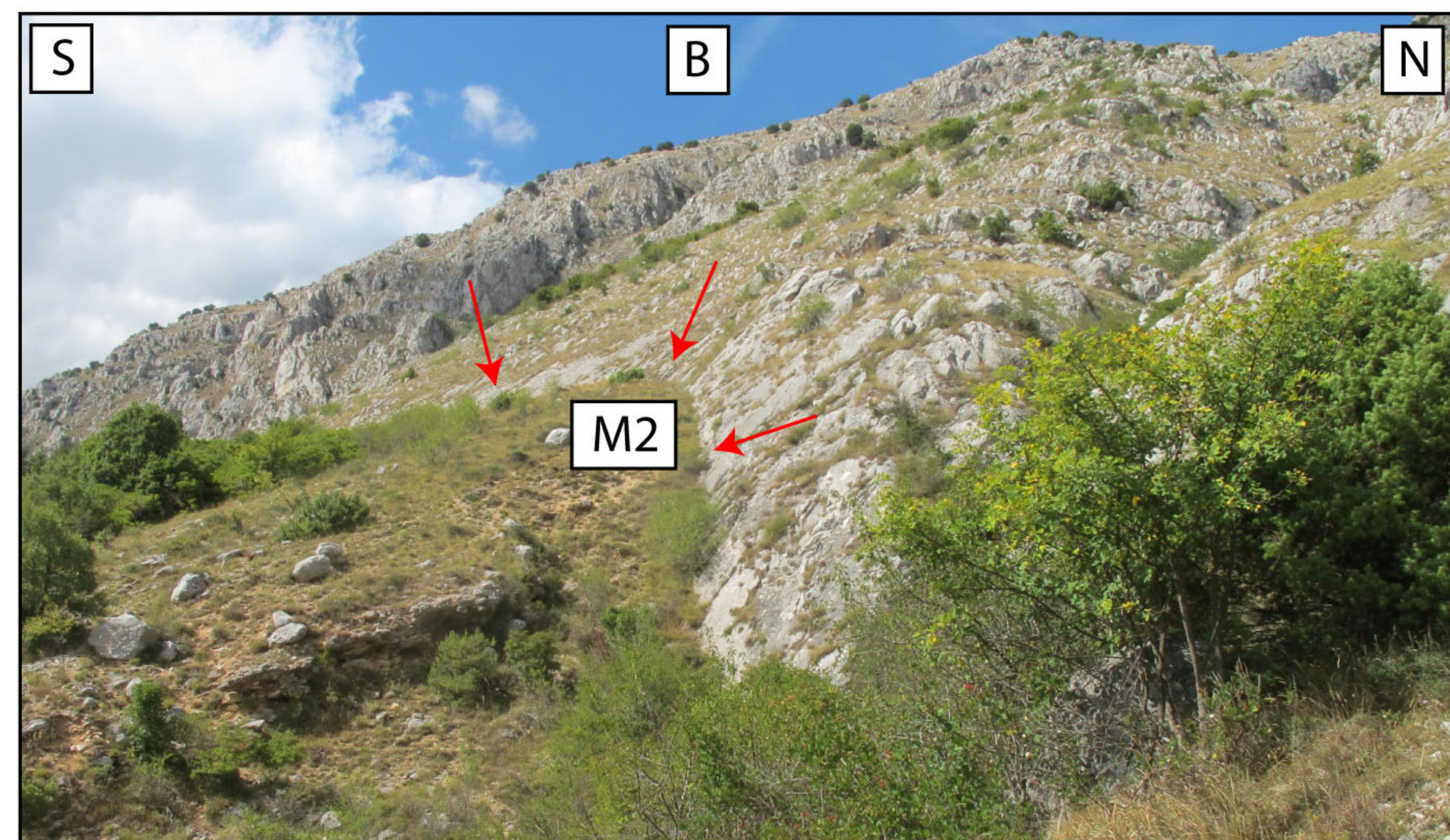
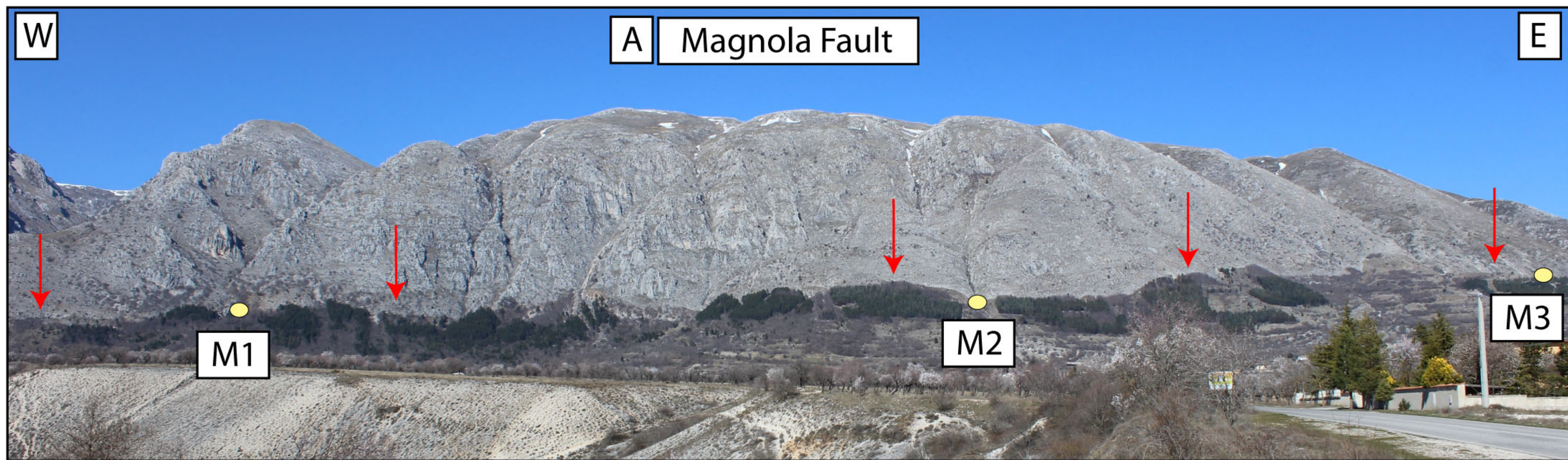


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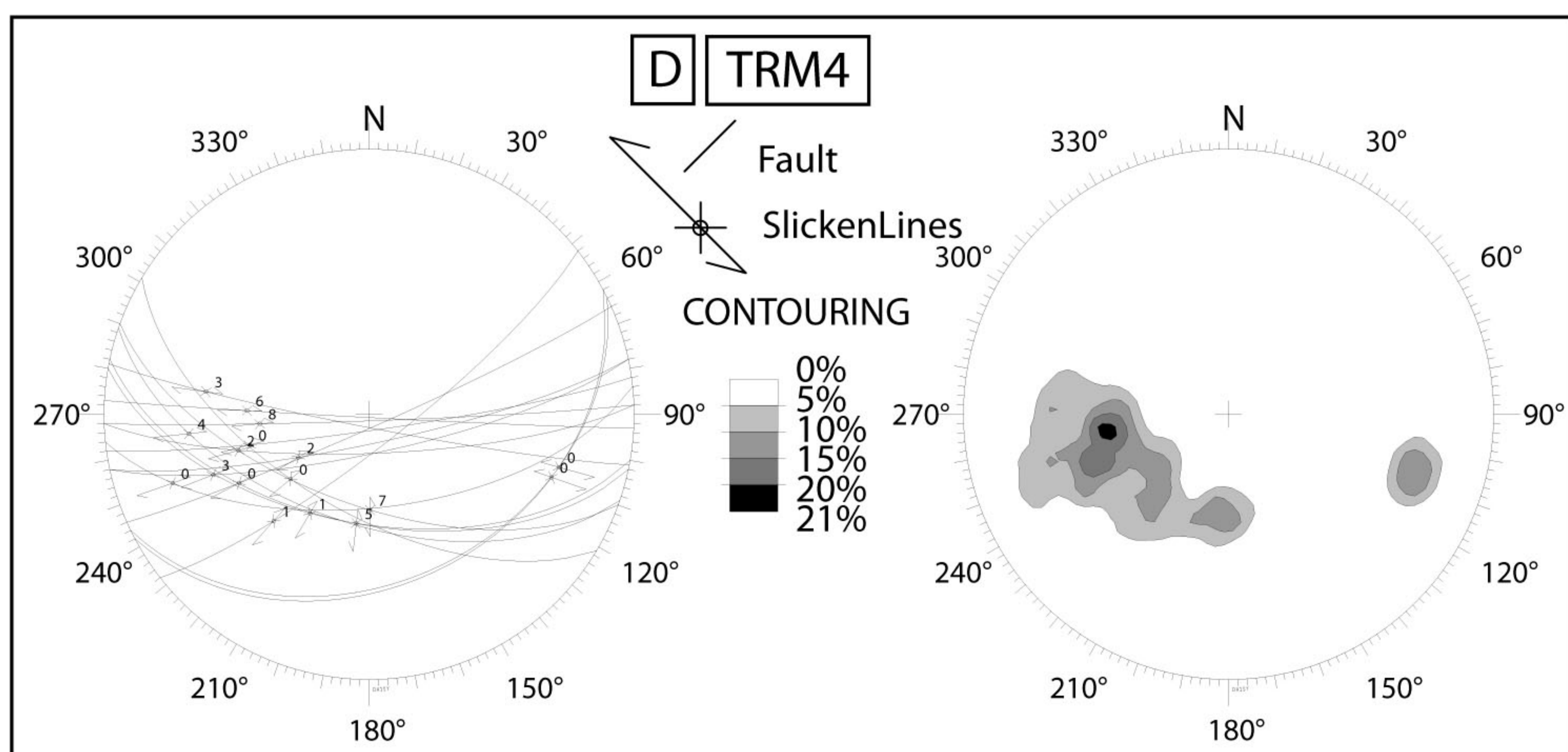
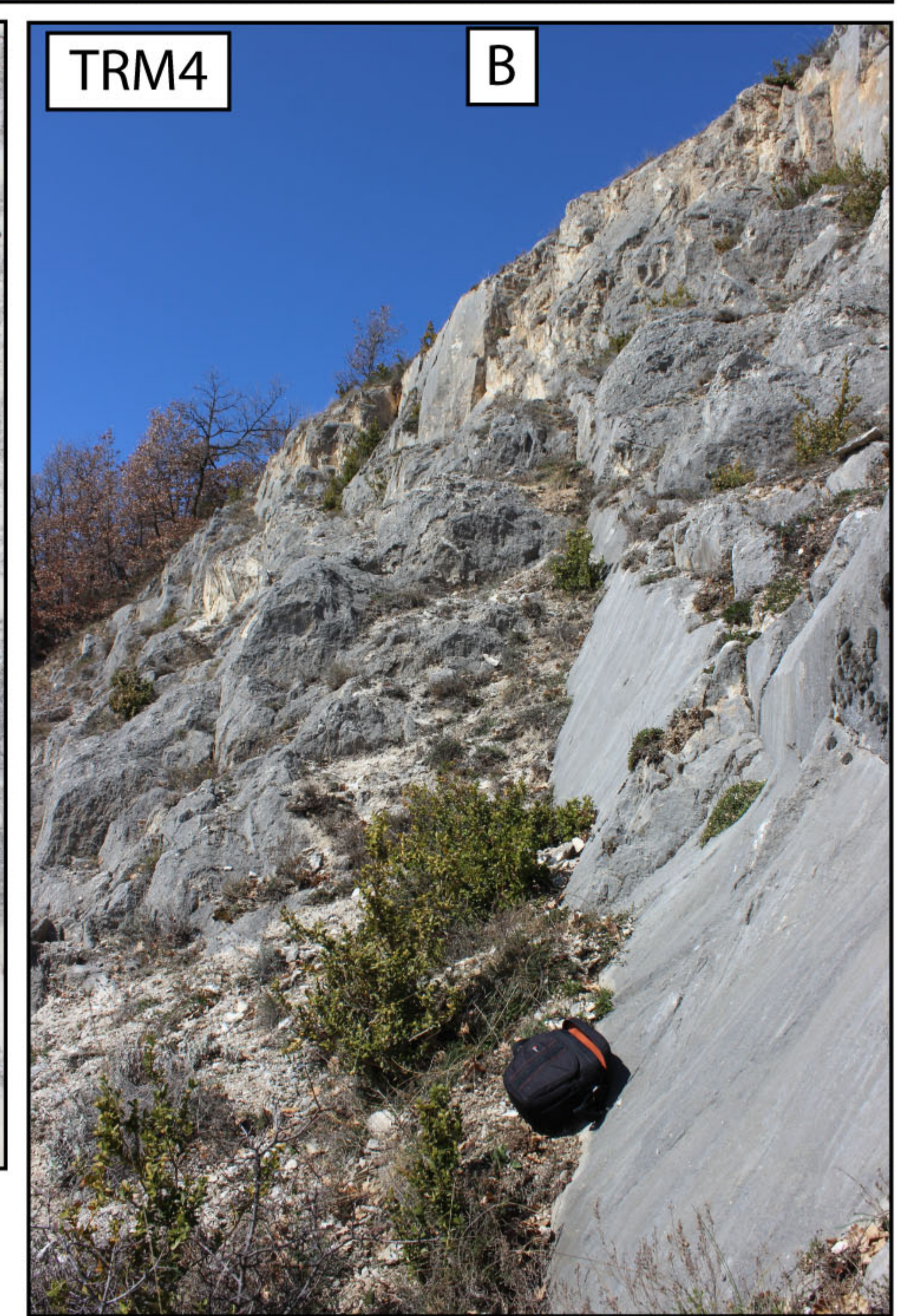
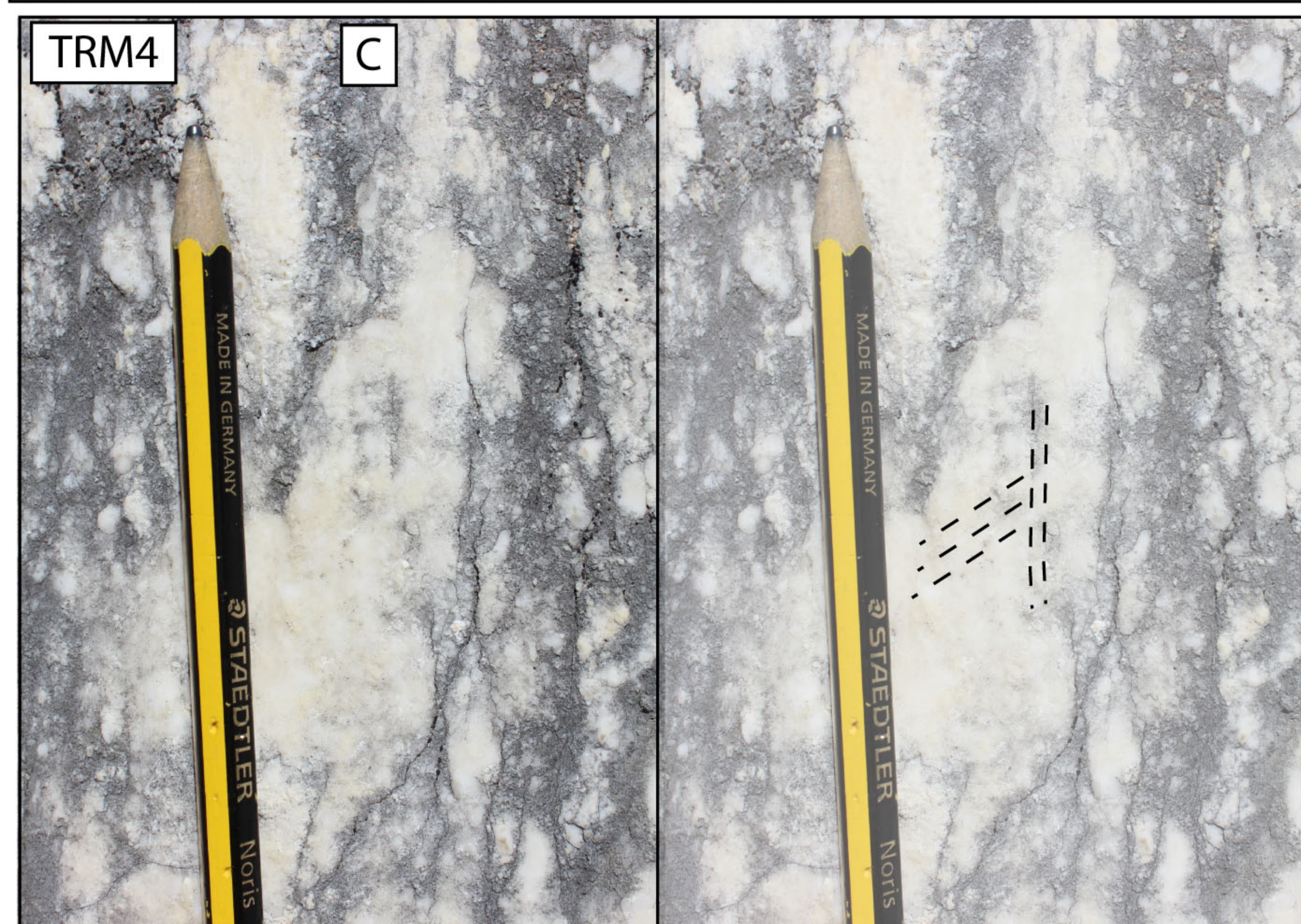
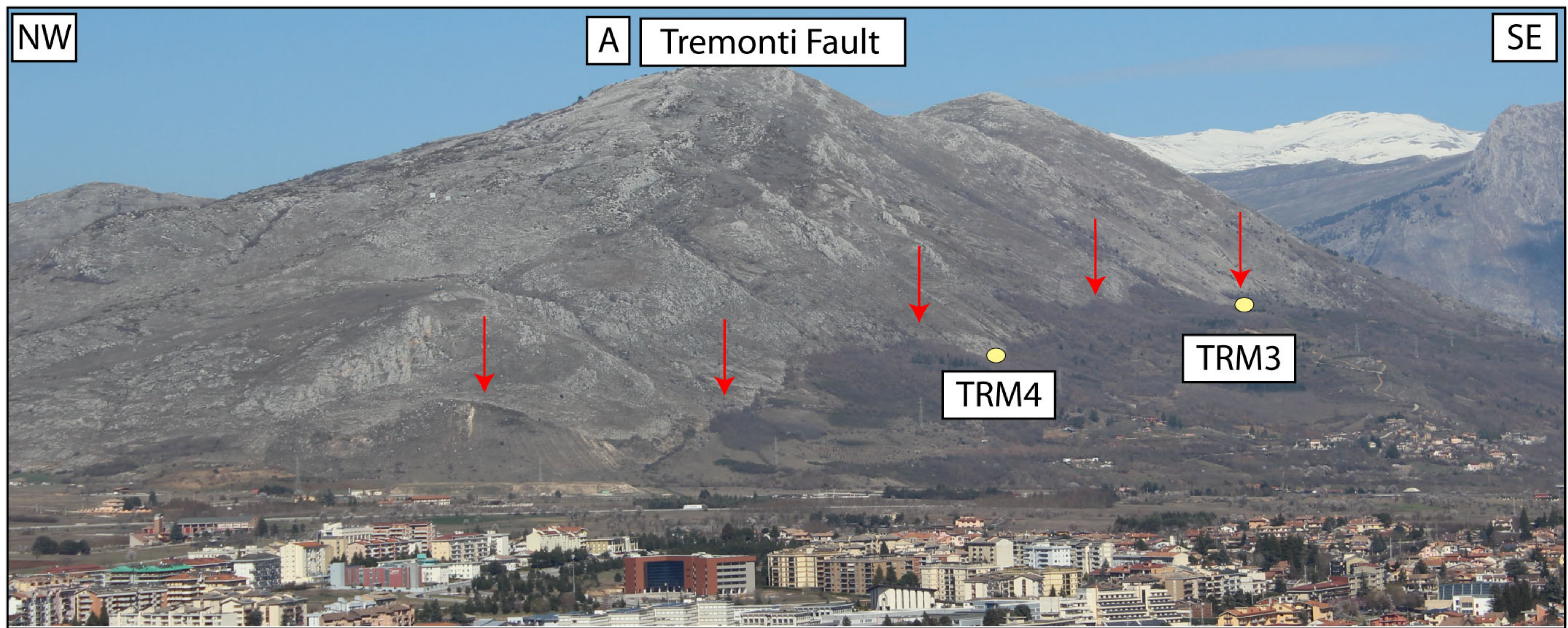


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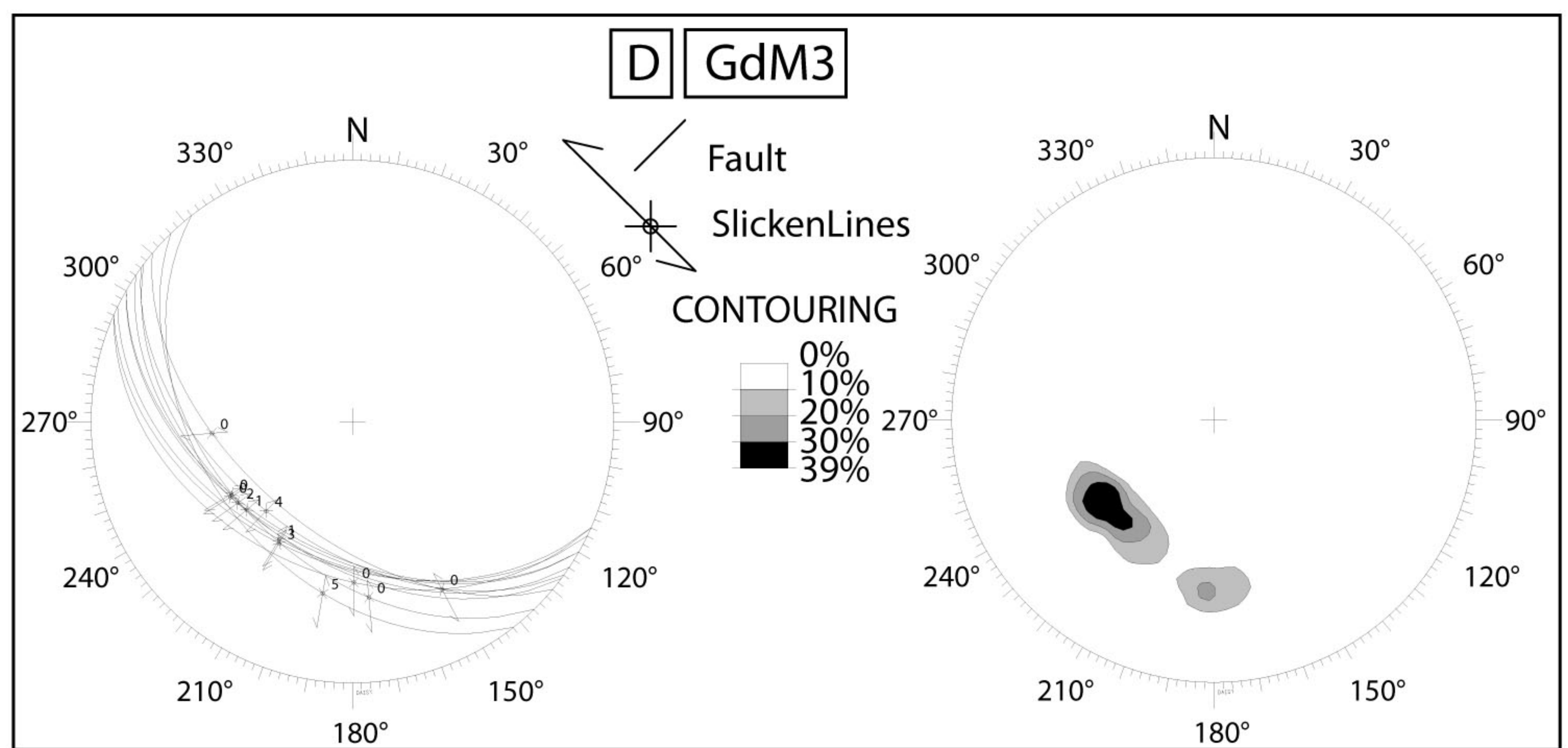
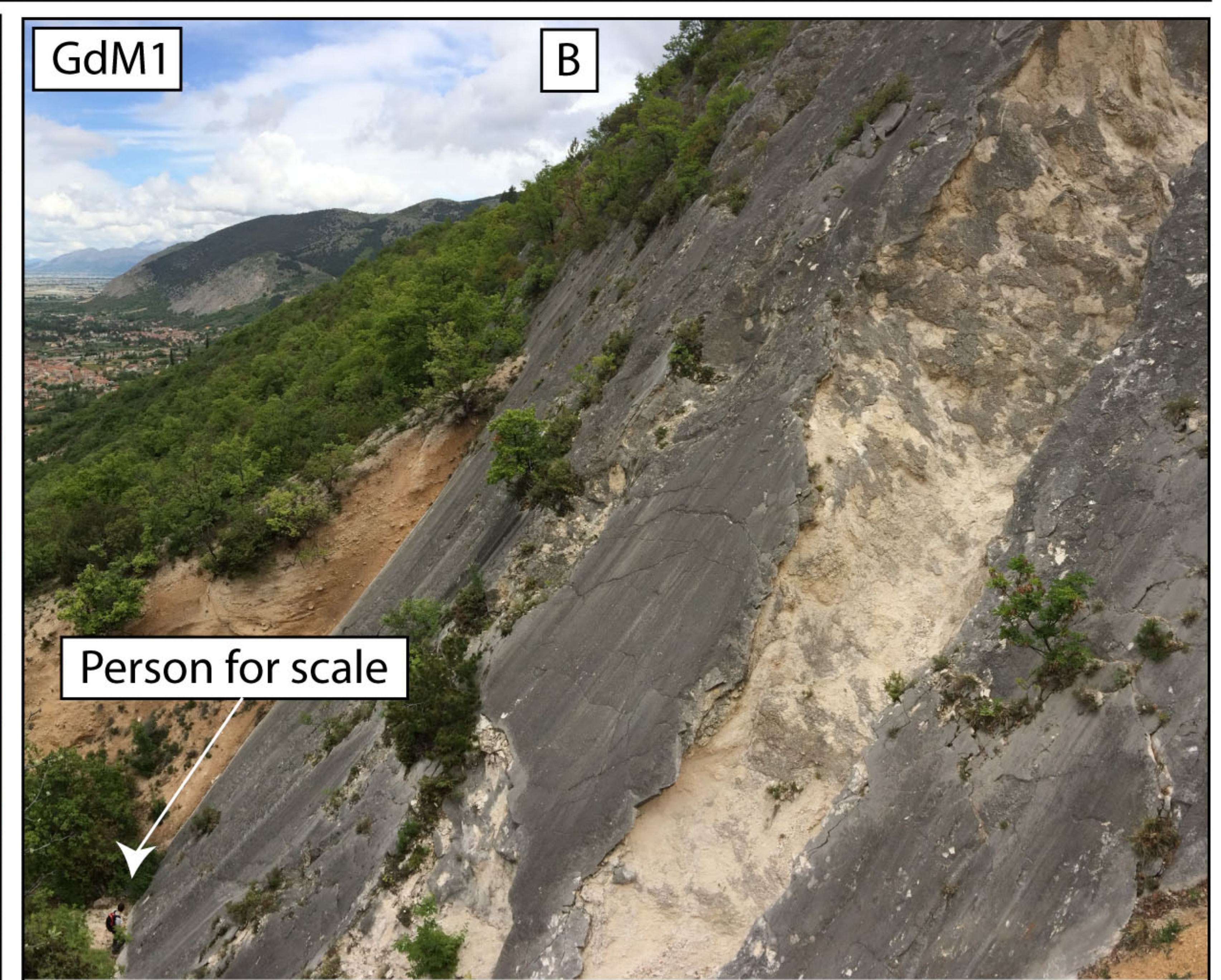
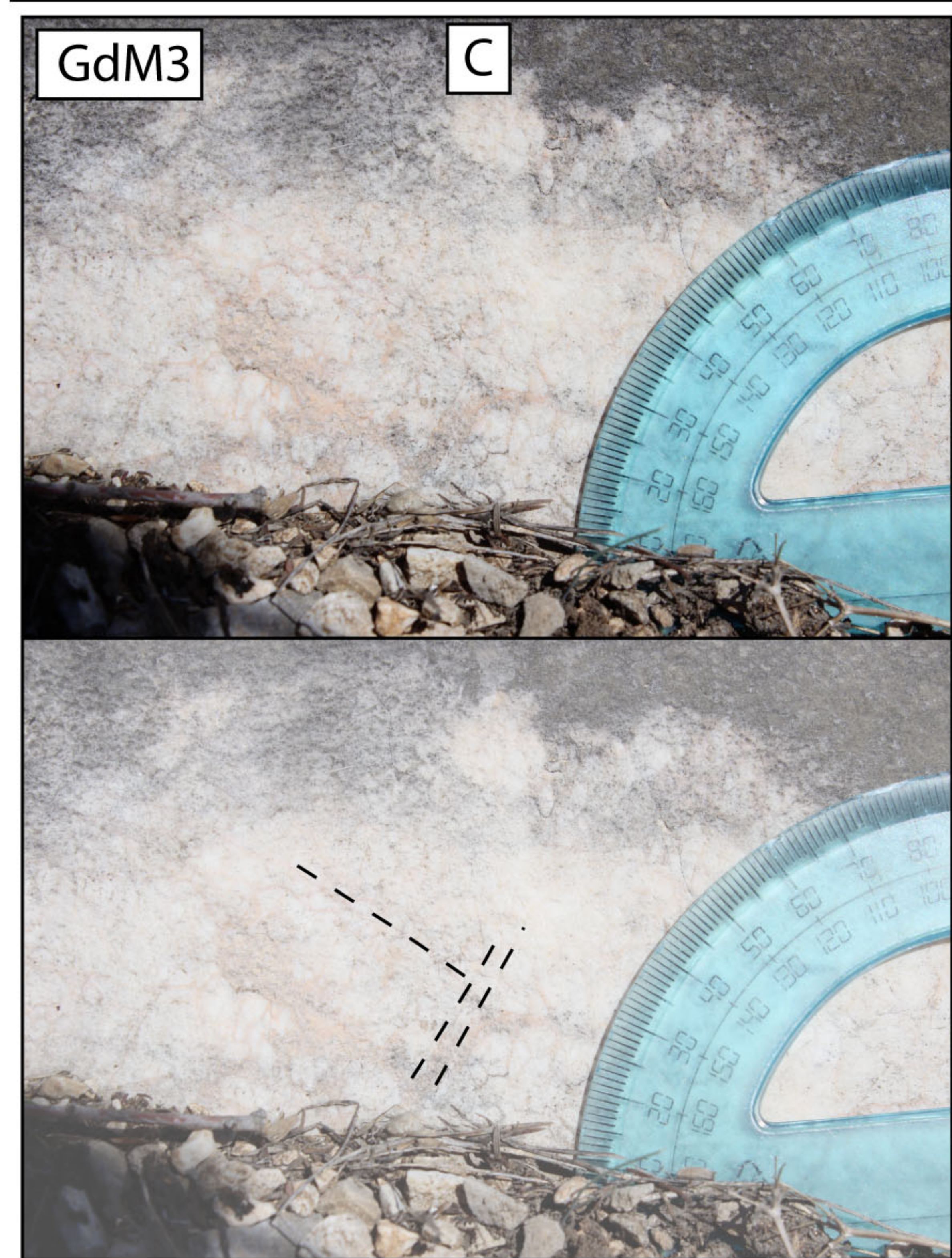
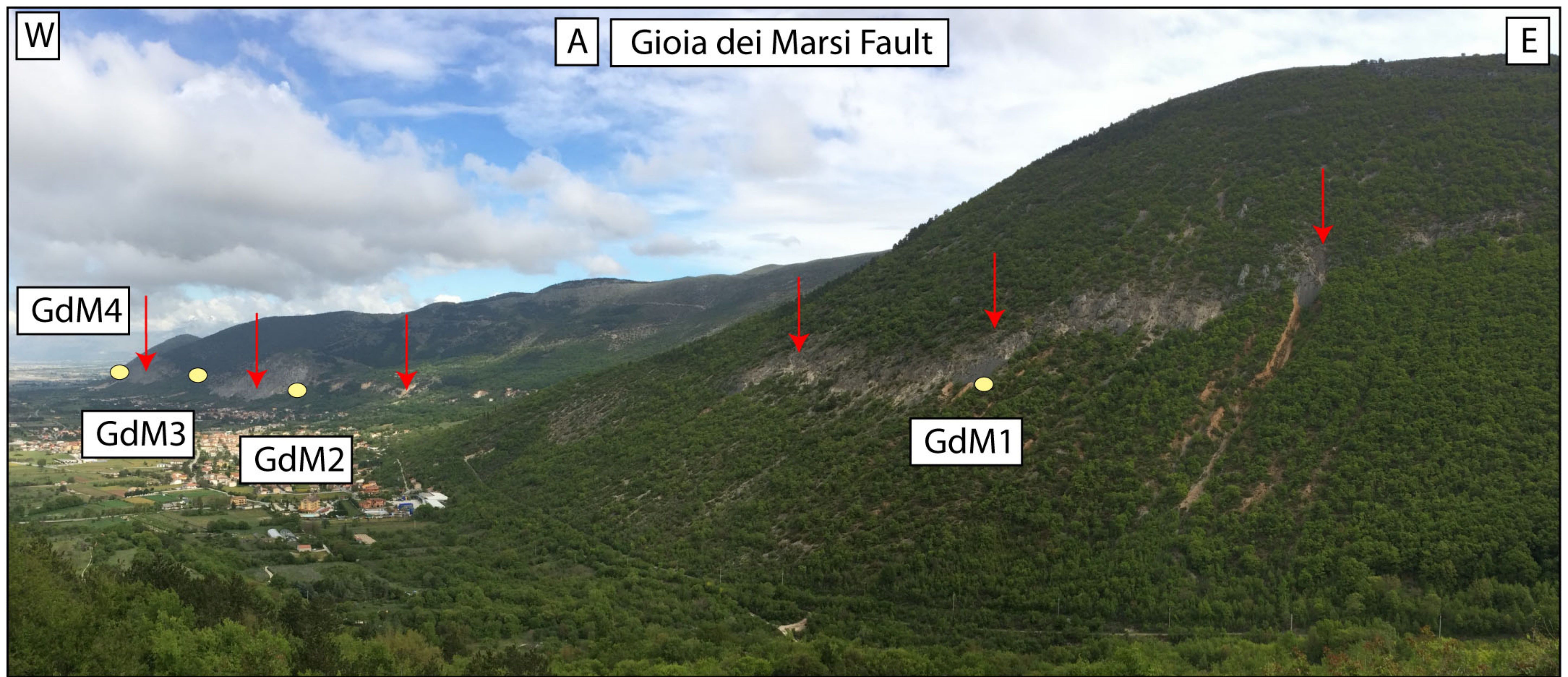
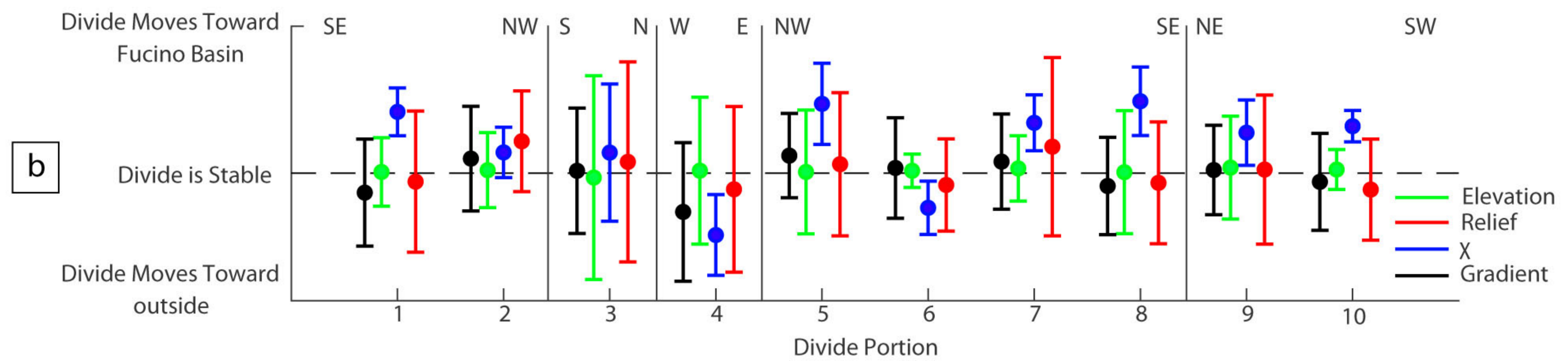
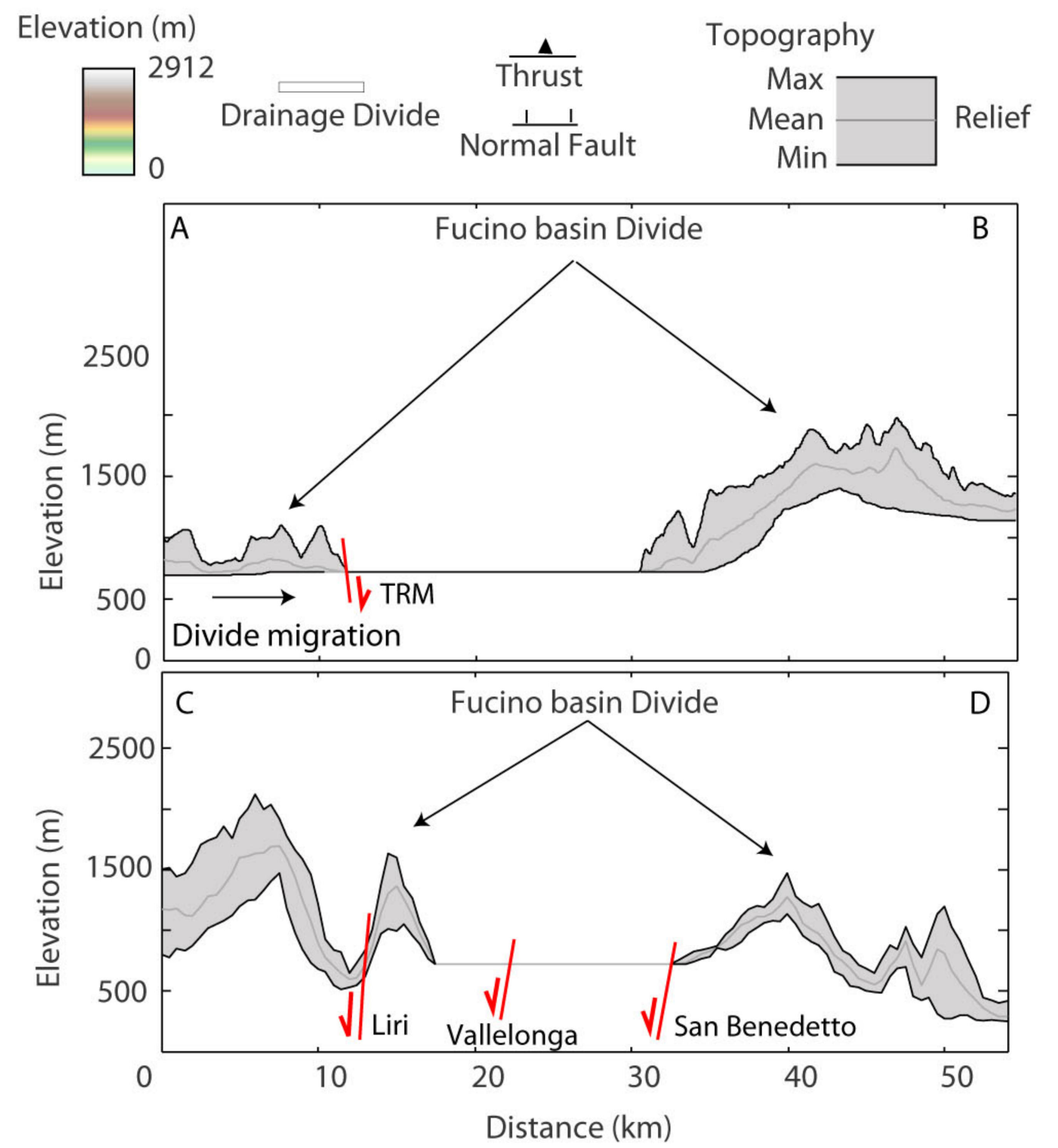
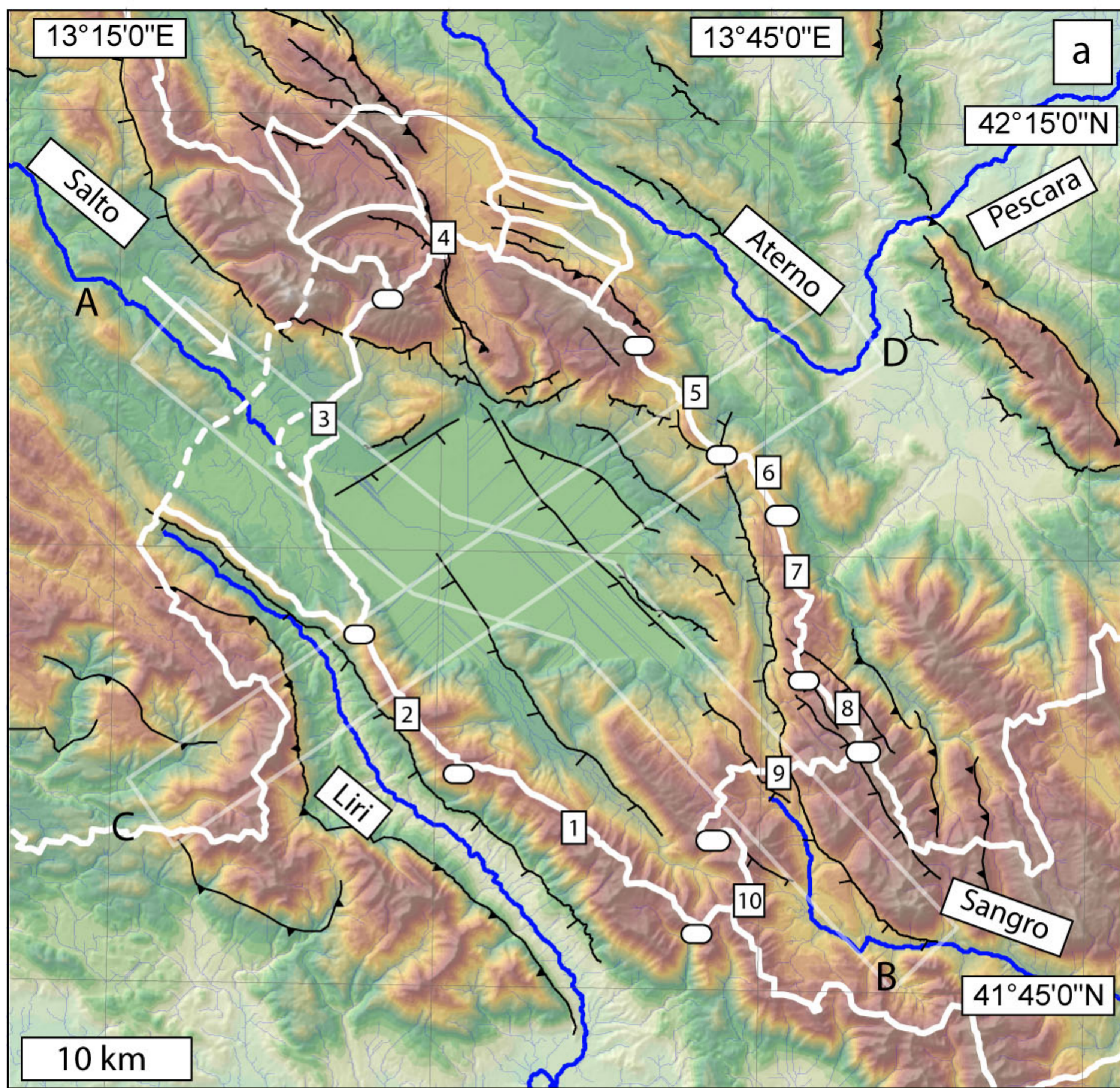


Figure8.



c) Local relief 5 km

d) Slope

e) ksn

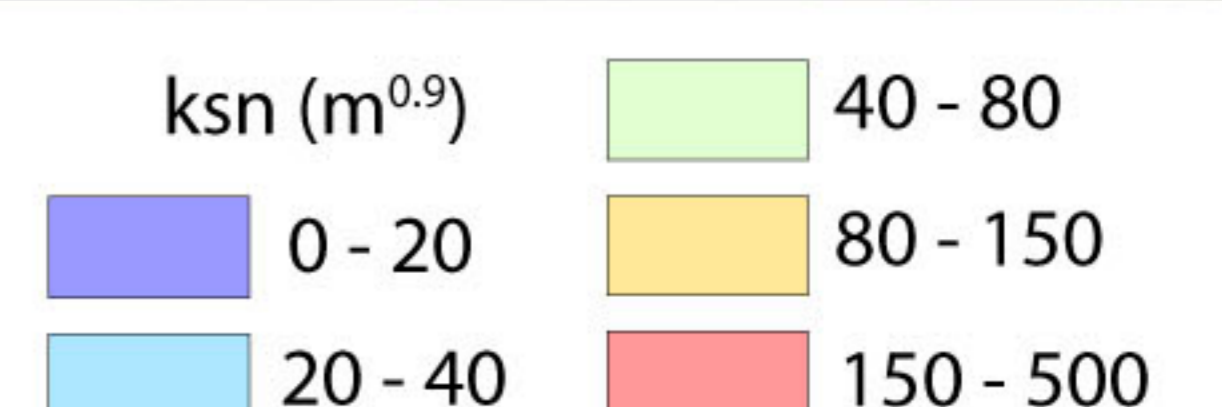
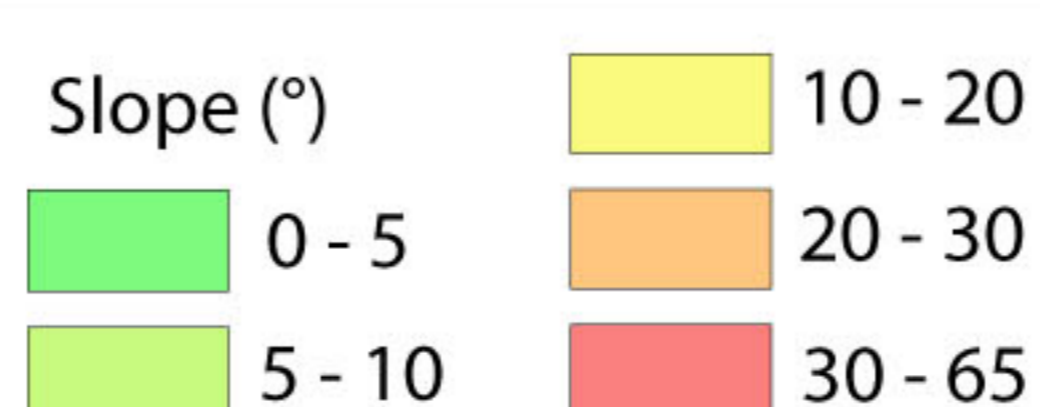
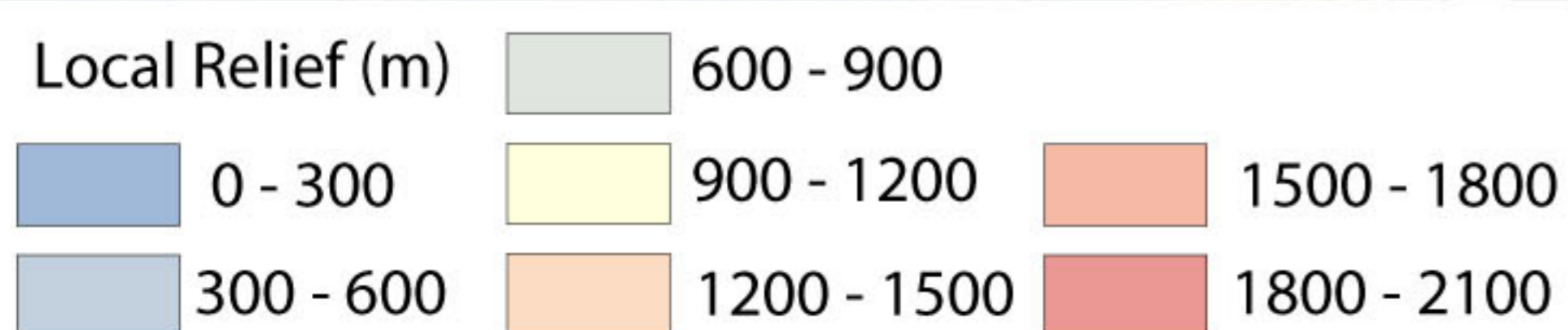
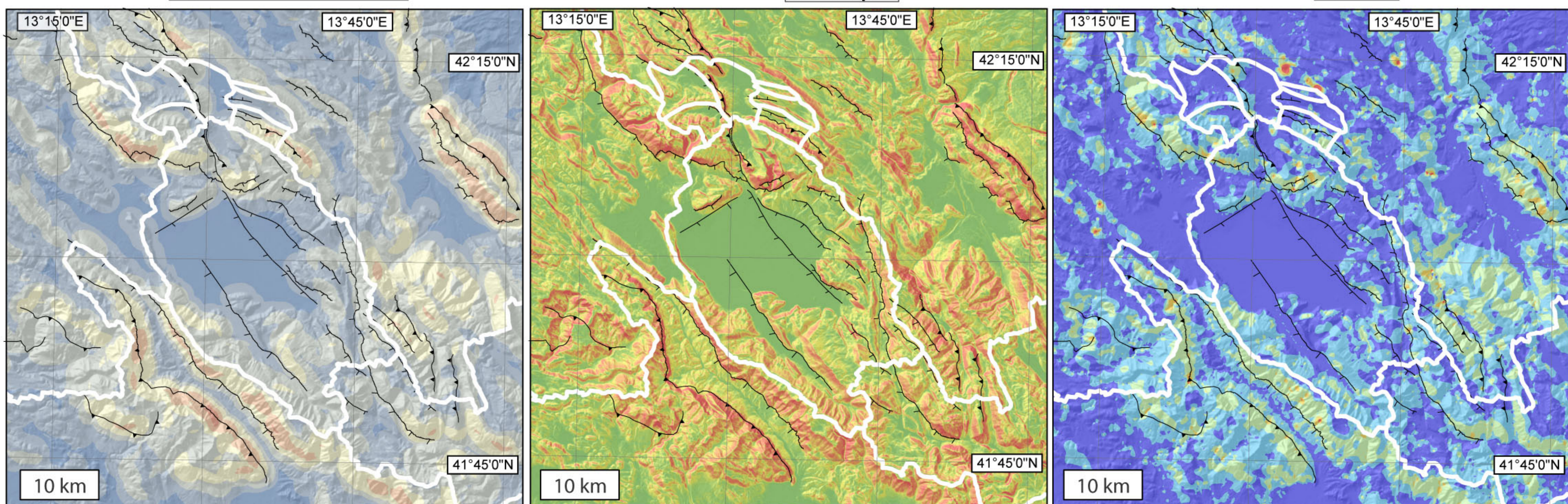
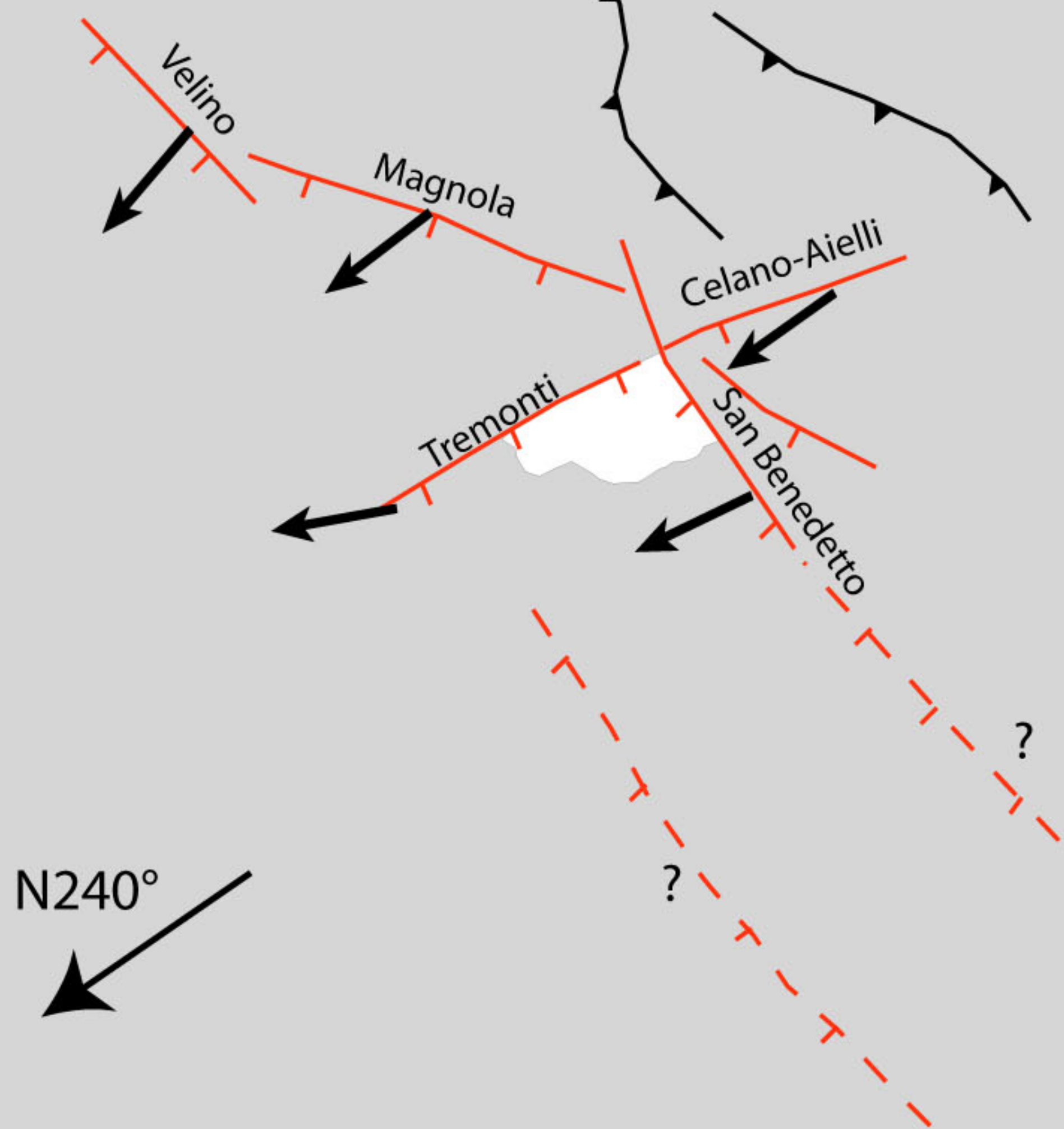


Figure9.

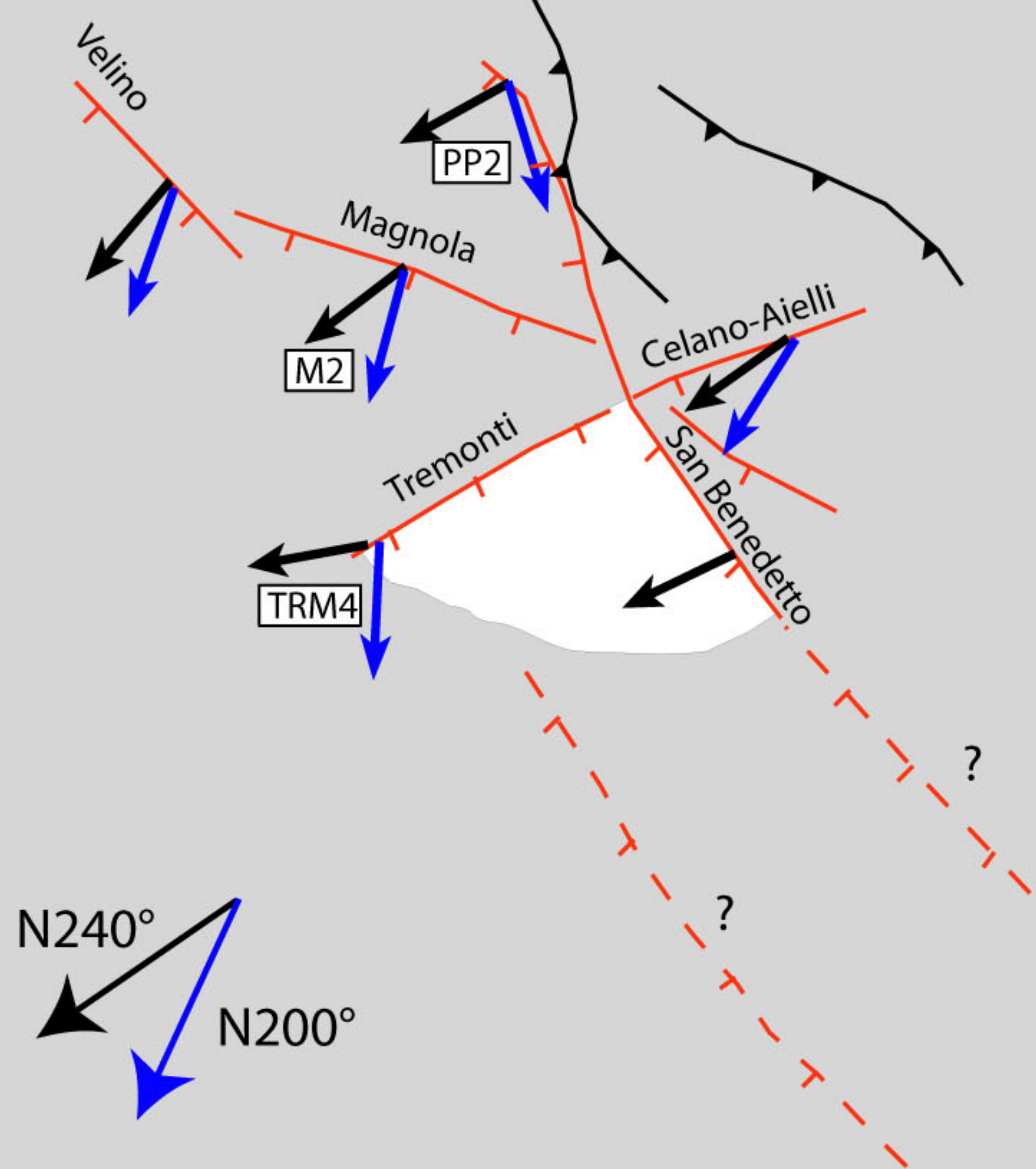
Thrust
Normal Fault

a) Late Pliocene

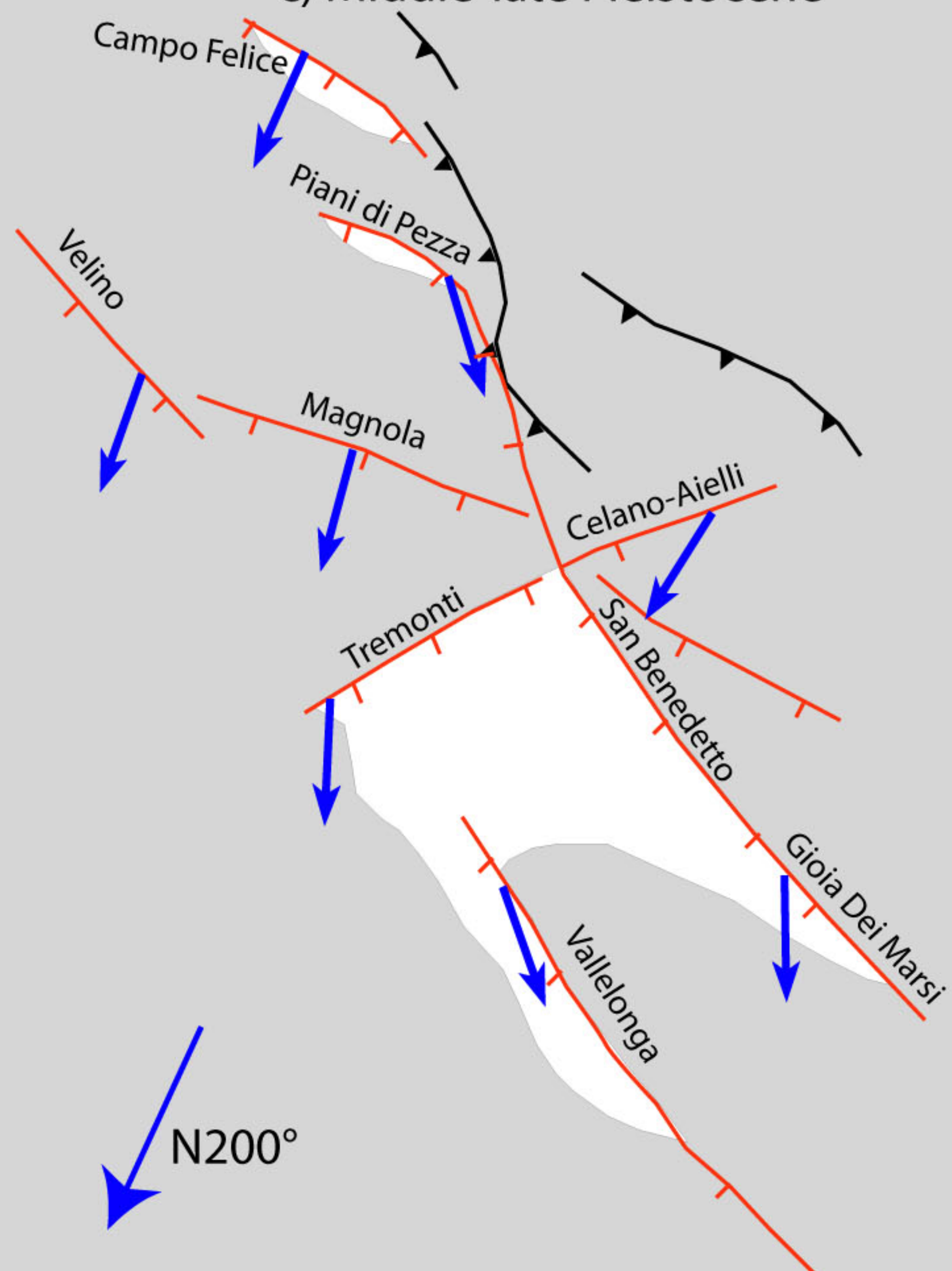
Sediments
Bedrock



b) Early-middle Pleistocene



c) Middle-late Pleistocene



d) Late Pleistocene-Holocene

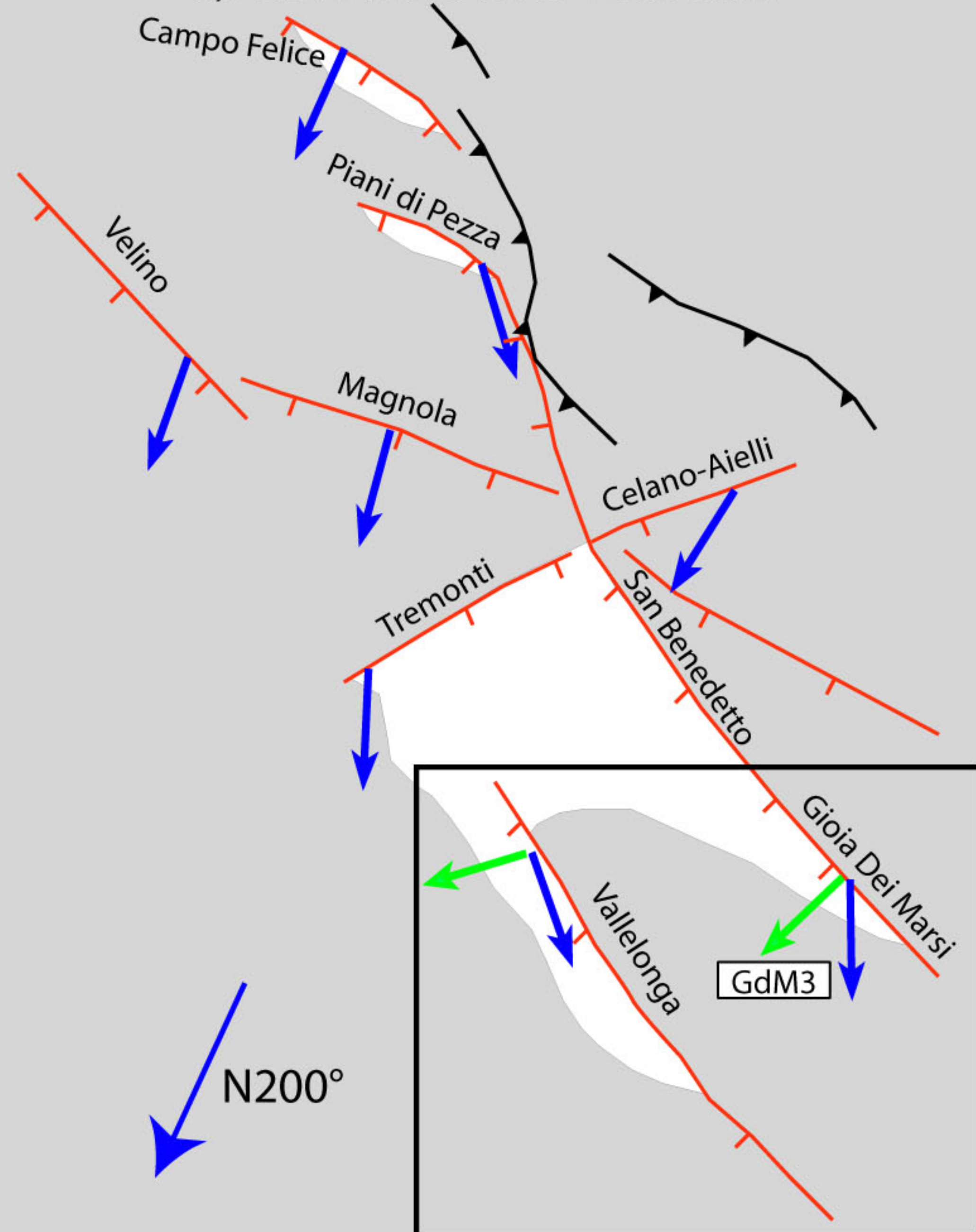


Figure10.

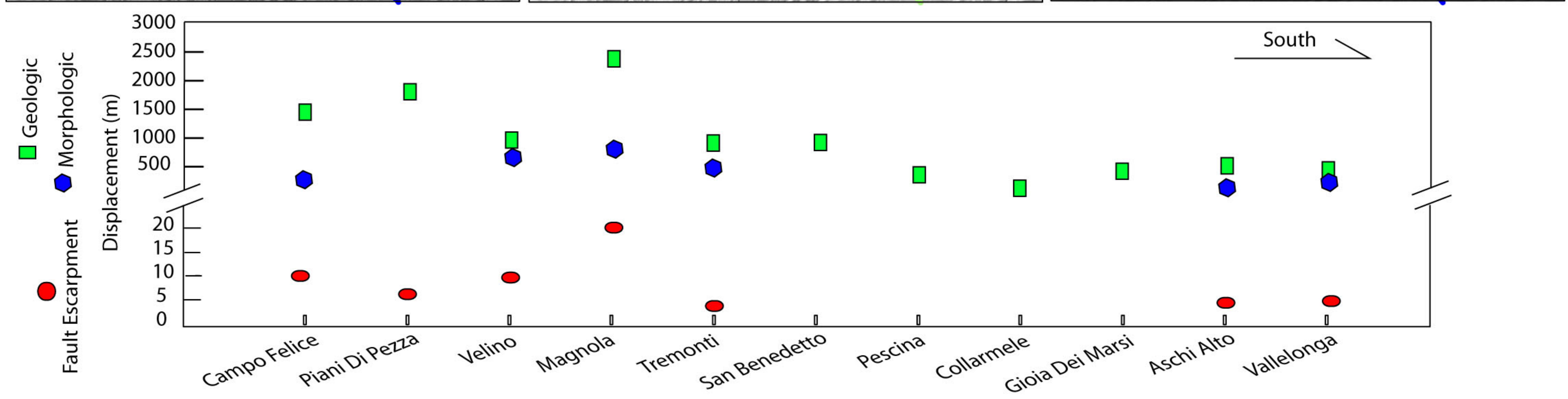
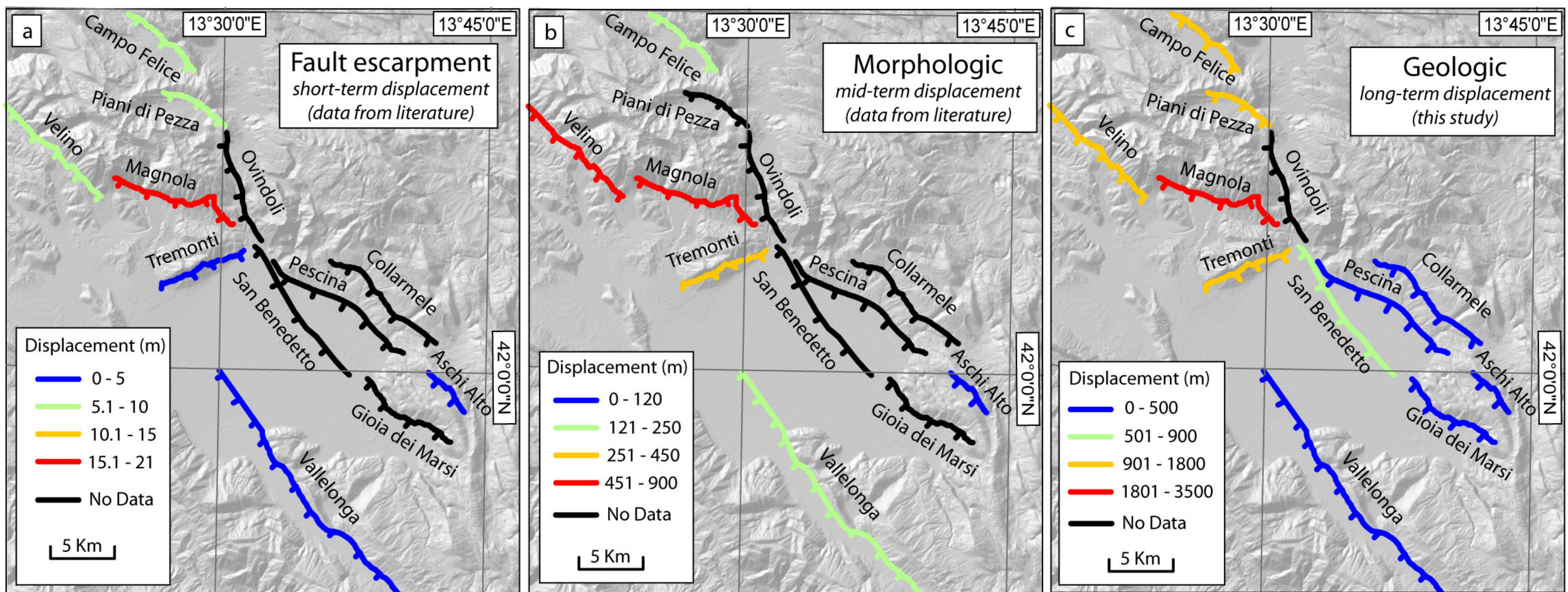
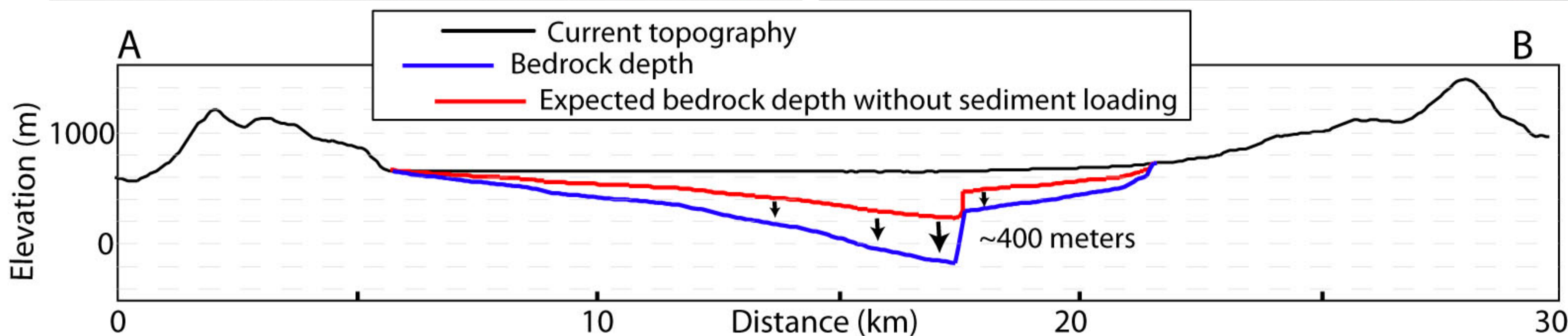
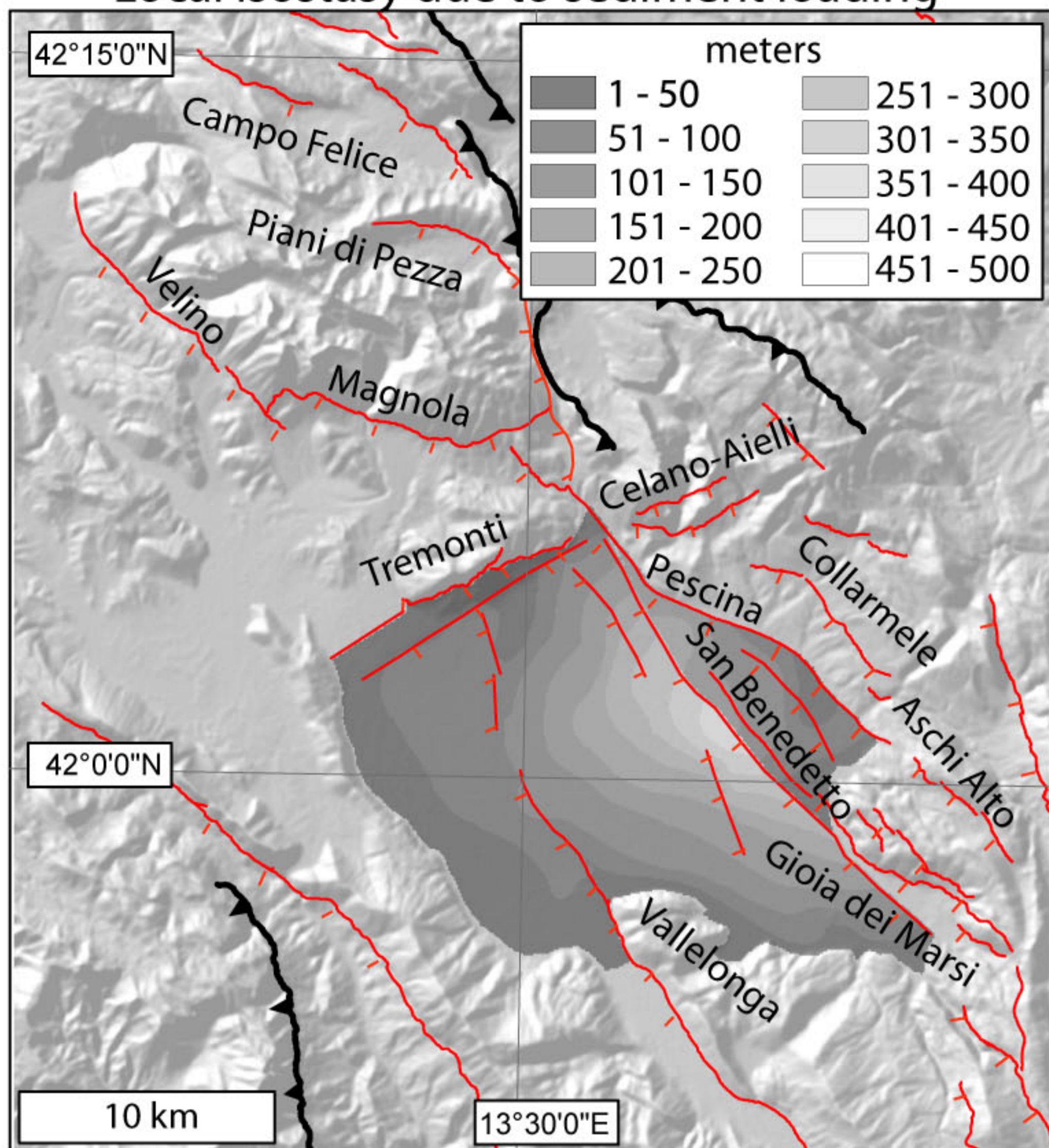
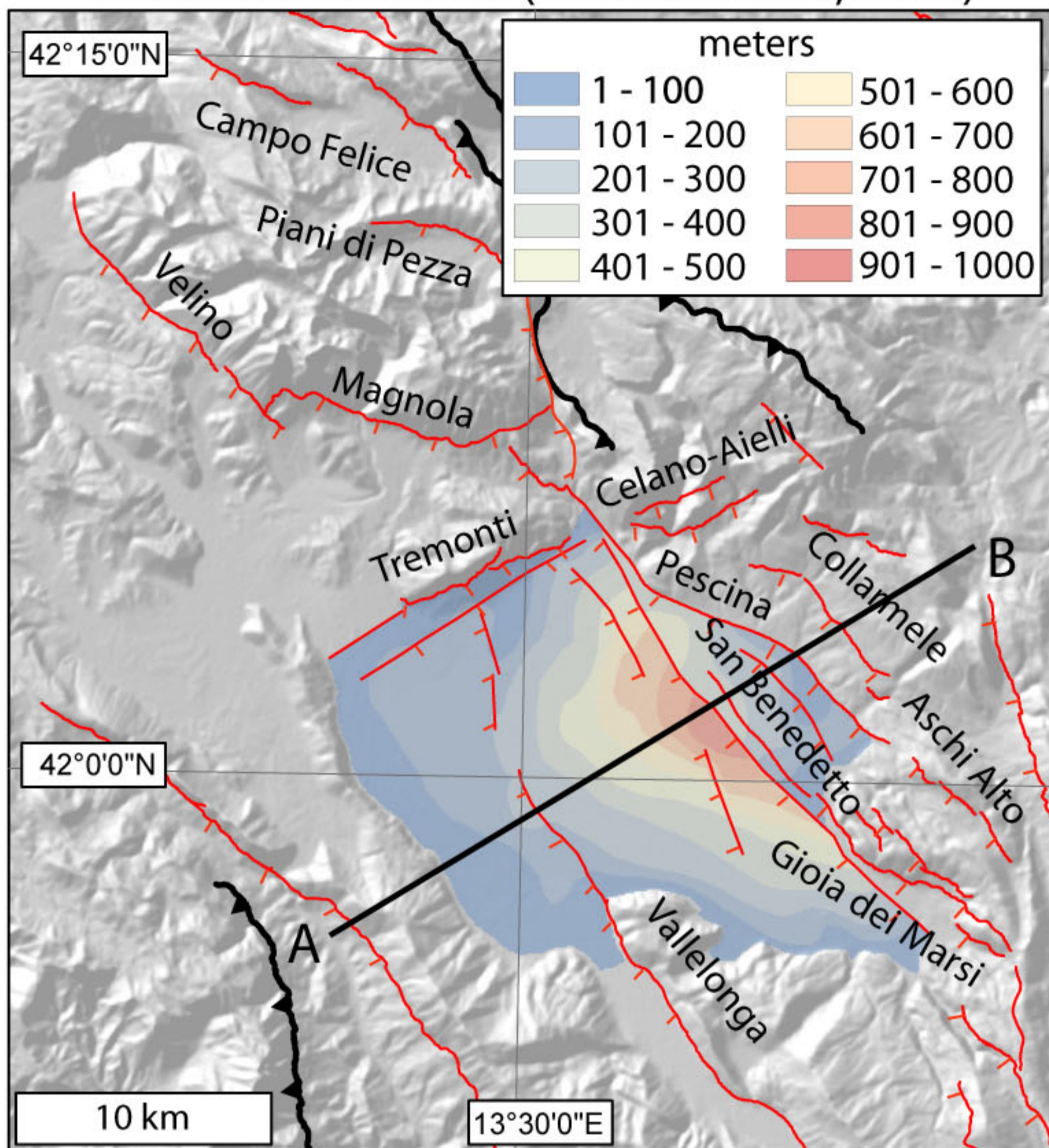


Figure11.

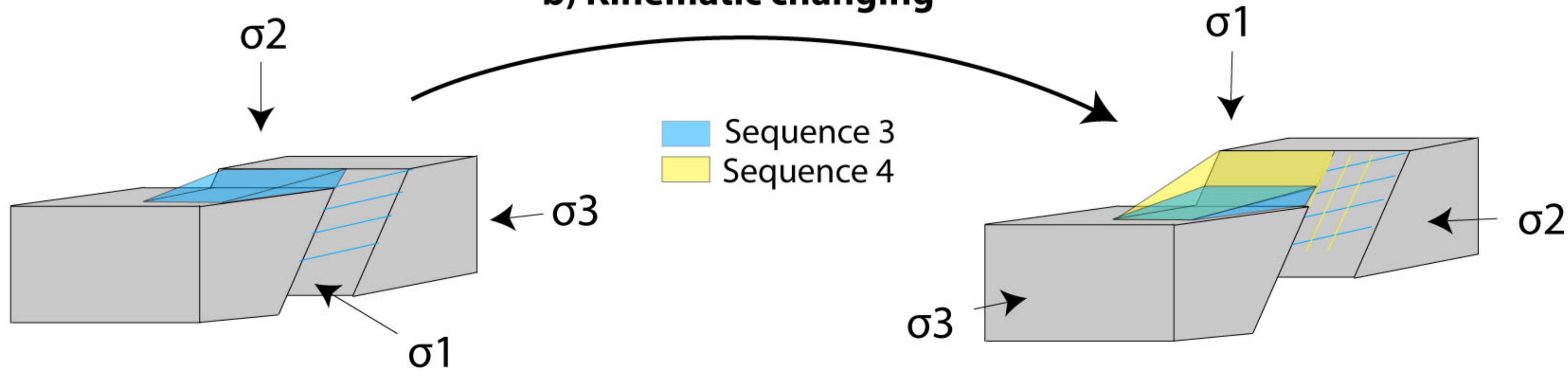
a) Increasing in vertical displacement

Sediment thickness (Cavinato et al., 2002)

Local Isostasy due to sediment loading

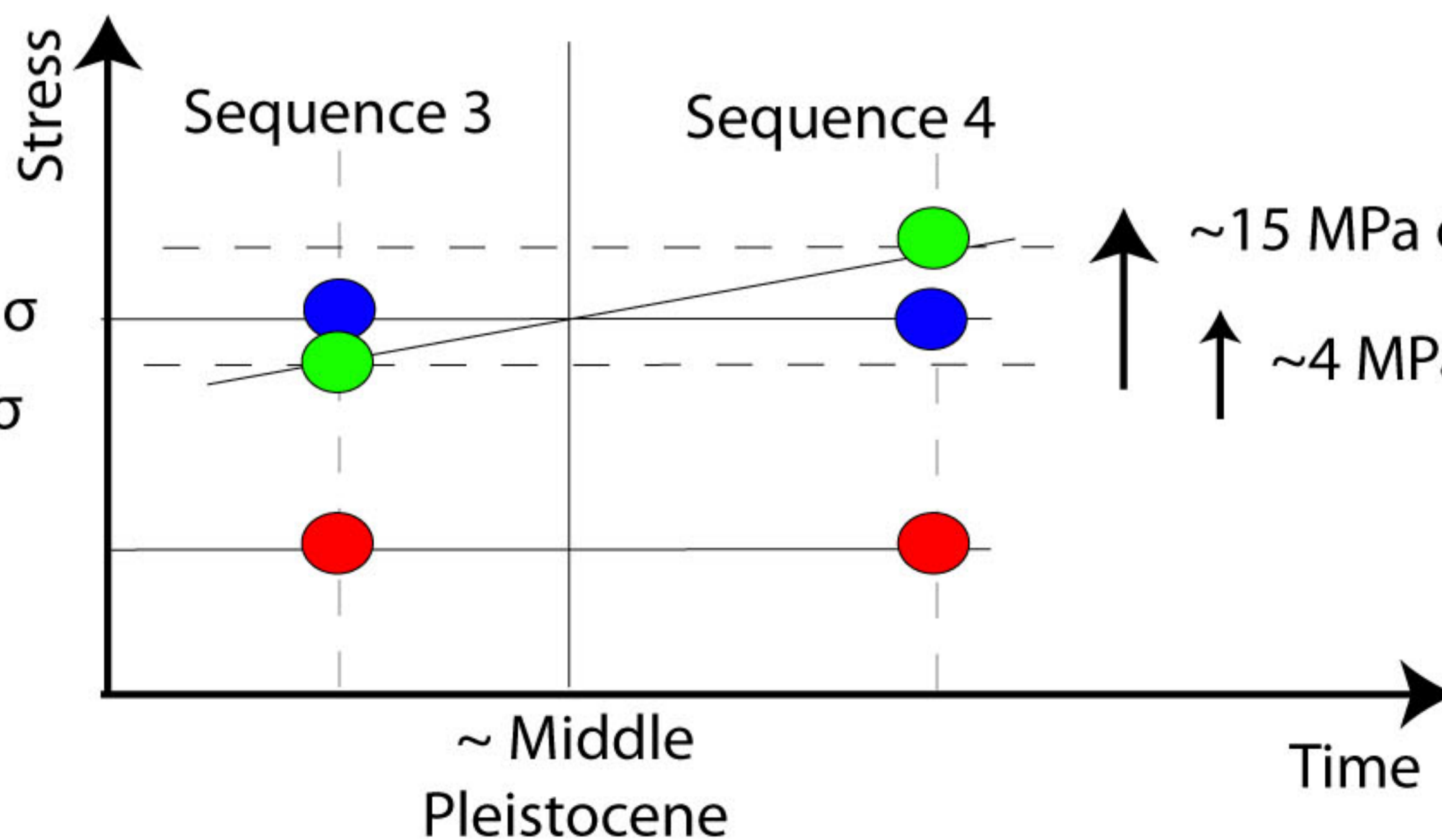


b) Kinematic changing



Strike/Oblique-slip stress regime

Normal dip-slip stress regime



- Maximum Horizontal σ
- Minimum Horizontal σ
- Vertical σ

~15 MPa of total increasing in Vertical σ
 ~4 MPa increasing to turn the Vertical σ into the maximum σ

~ Middle Pleistocene

Time

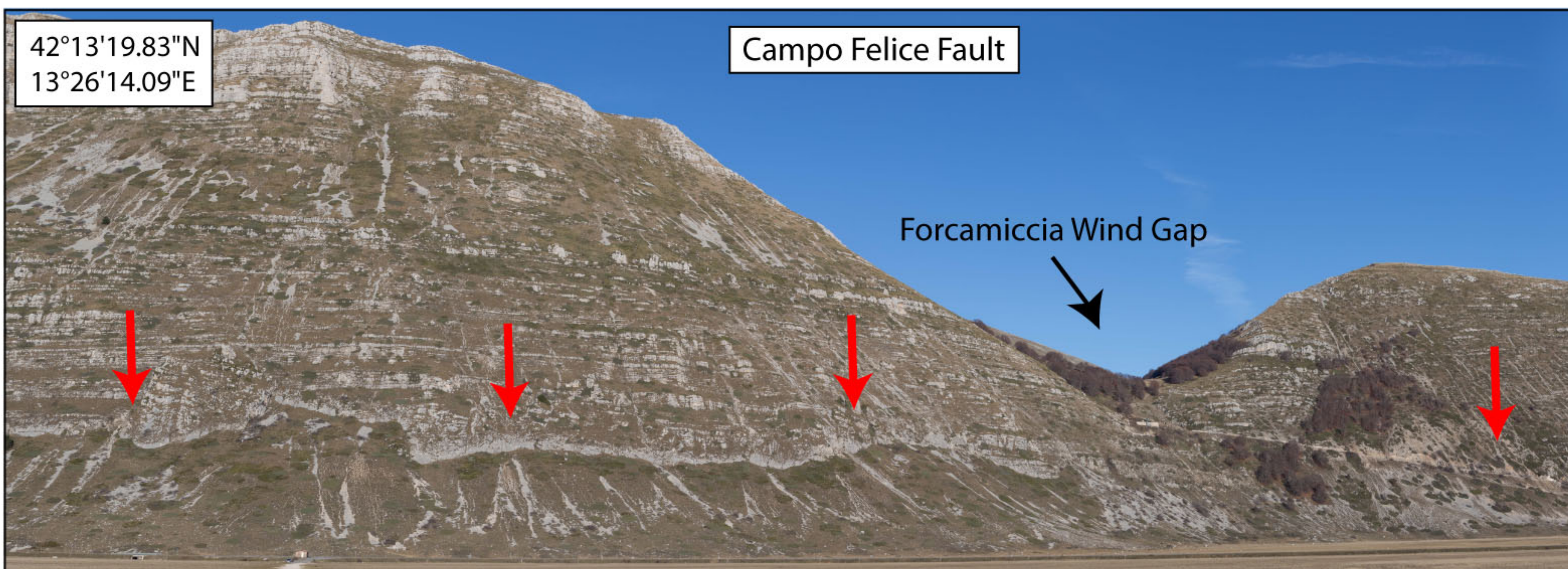
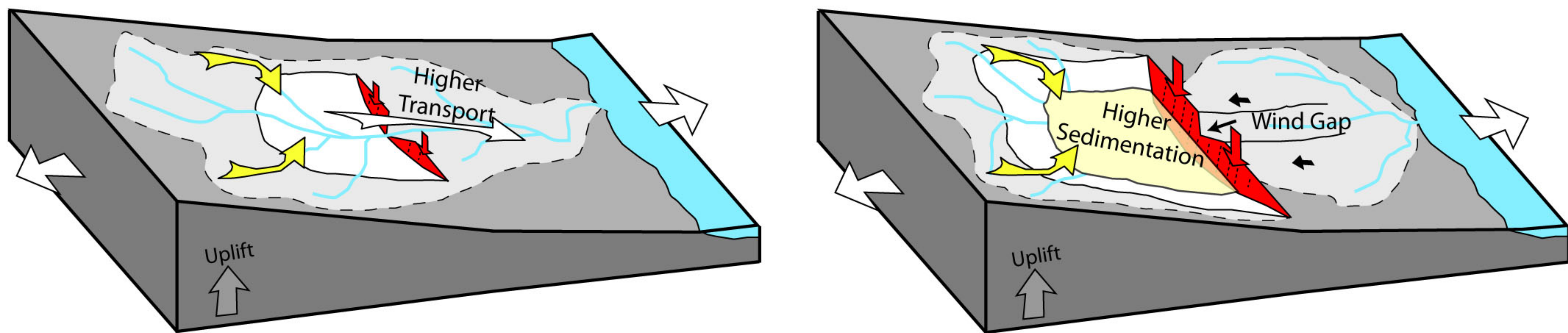
Figure12.

a)

From Exorheic to Endorheic

T1 - Erosion faster or equal than Faulting

T2 - Erosion slower than Faulting



b)

From Endorheic to Exorheic

T1 - Erosion slower than Faulting

T2 - Erosion faster than Faulting

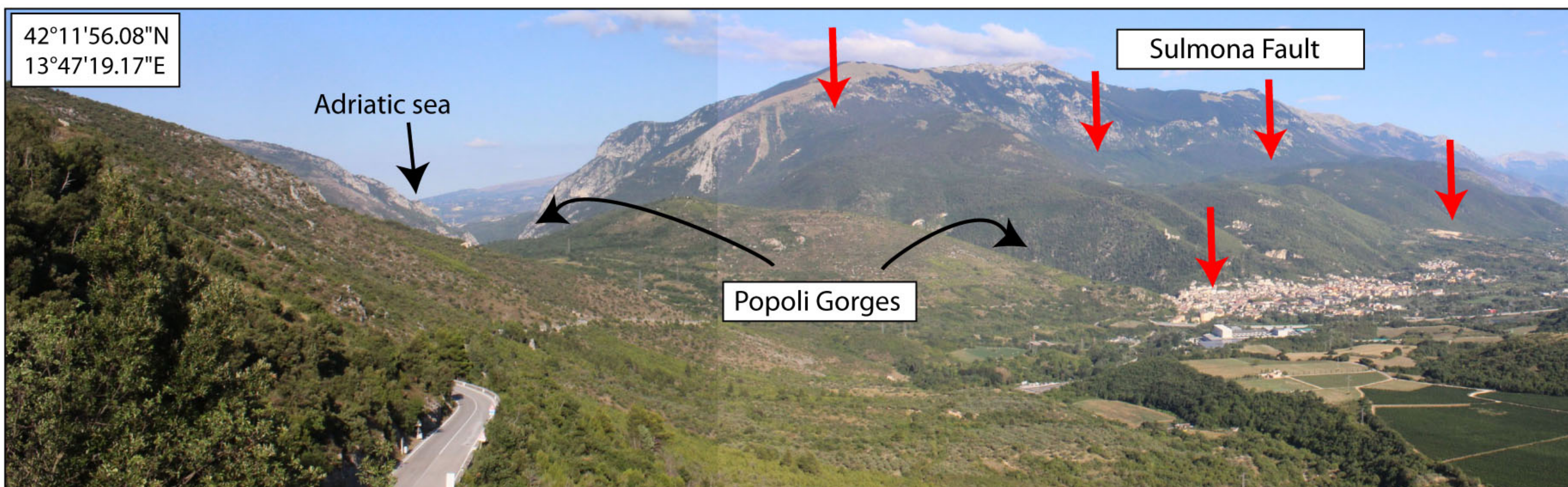
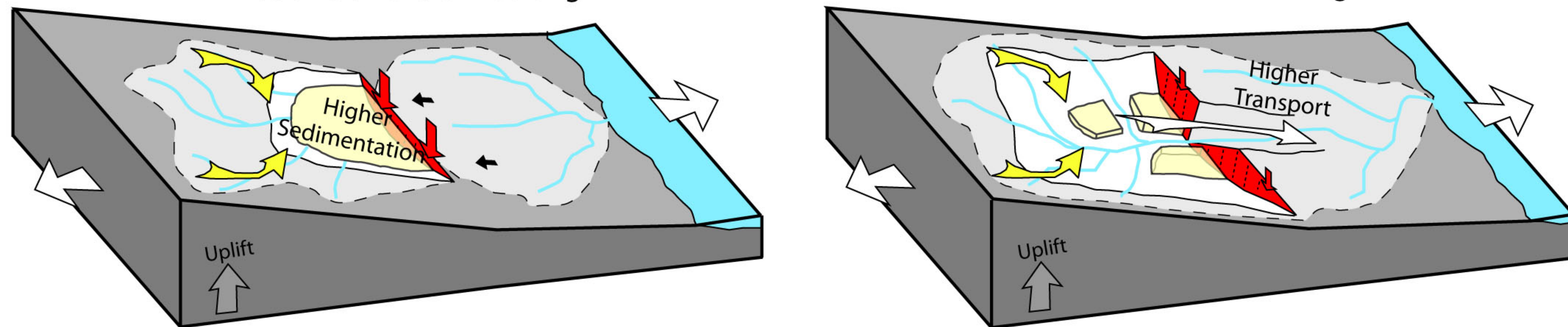
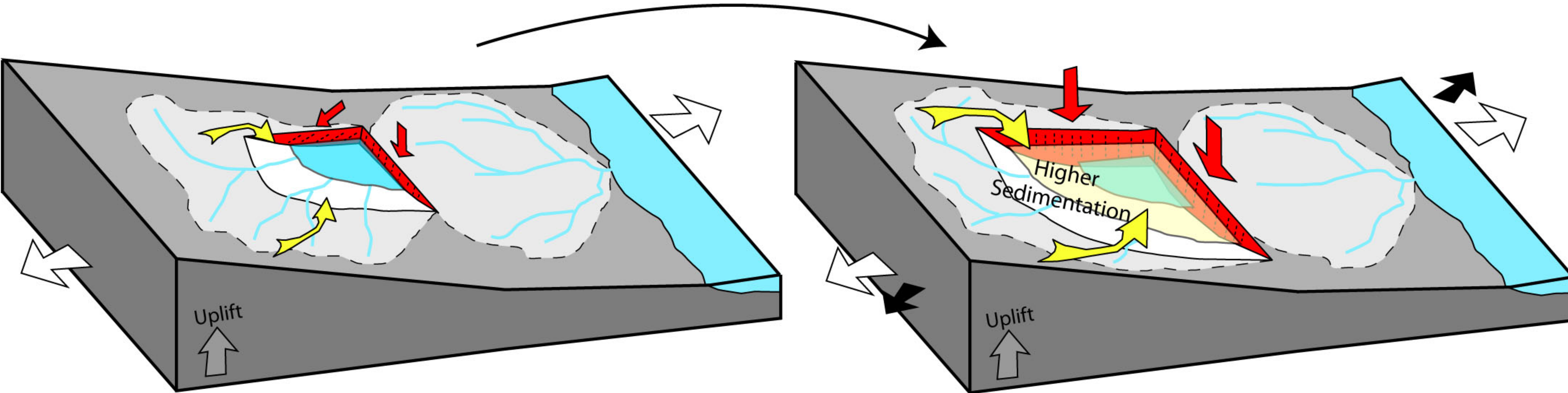


Figure13.

Schematic Structural evolution of the Fucino Basin

-Increasing in sedimentation
 -Kinematic changing from oblique- to dip-slip on the transfer fault



↔ Regional direction of extension (N240°)
 ↔ Local direction of extension (N200°)

Sequence 3
 Sequence 4
 From Cavinato et al., (2002)

