Unveiling ductile deformation during fast exhumation of a granitic pluton in a transfer zone

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Abstract
Exhumation and cooling of upper crustal plutons is generally assumed to develop in the brittle domain, thus determining an abrupt passage from crystallisation to faulting. To challenge this general statement, we have applied an integrated approach involving meso- and micro-structural studies, thermochronology, geochronology and rheological modeling. We have analyzed the Miocene syn-tectonic Porto Azzurro pluton on Elba (Tuscan archipelago – Italy), emplaced in an extensional setting, and have realized that its fast exhumation is accompanied by localized ductile shear zones, developing along dykes and veins, later affected by brittle deformation. This is unequivocally highlighted by field studies and the analysis of microstructures with EBSD. In order to constrain the emplacement and exhumation rate of the Porto Azzurro pluton we performed U-Pb zircon dating and (U+Th)/He apatite thermochronology. It results in a magma emplacement age of 6.4 ± 0.4 Ma and an exhumation rate of 3.4 to 3.9 mm/yr. By thermo-rheological modeling we were able to establish that localized ductile deformation occurred at two different time steps: within felsic dykes when the pluton first entered into the brittle field at 380 kyr, and along quartz-rich hydrothermal veins at c. 800 kyr after pluton emplacement. Hence, the major conclusion of our data is that ductile deformation can affect a granitic intrusion even when it is entered into the brittle domain in a fast exhuming extensional regime.

Key Words: Granite emplacement, fast exhumation, extensional tectonics, shearing, geochronology –thermochronology, rheological modeling.

1. Introduction

If strain rate is high (≥$1*10^{-14}$ s^{-1}, Ranalli, 1995; Rey et al., 2009), extensional tectonics can be accompanied by high heat flow and diffuse emplacement of magmatic plutons at different structural levels, from the base of of the ductile lower crust to the brittle upper crust. In this framework, the migration of fluids is favored by the kinematics of the shear zones, since the intermediate shear axis plays a significant role in defining the orientation of the structural channels where permeability is promoted (e.g.: Sibson, 2000; Rowland and Sibson, 2004; Liotta and Brogi, 2020). In particular, shear zones with a dominant transcurrent regime, such as the transfer zones (Lister and Davis, 1989, Gibbs, 1990), due to their geometrical and kinematic features, can produce the development of relatively restricted permeable crustal volumes where magmas can be channeled, to form shallow-

level plutonic bodies, that rapidly cool down during exhumation. It is generally believed that this fast exhumation is accompanied by brittle deformation during cooling of the magmatic body (e.g.: Moyen et al., 2003; Liu et al., 2020). With this idea in mind, we have analyzed the syn-tectonic Porto Azzurro pluton (Elba Island, **Fig.1**) and have realized that its exhumation is accompanied by localized ductile shear zones, developing along magmatic dykes and hydrothermal veins, later affected by brittle deformation. This sequence of events implies particular relationships among exhumation, cooling, stress and strain localization. To unravel the interaction between the structural history and exhumation-driven cooling, an integrated approach involving meso- and micro-structural studies, geochronology and rheological modeling is here adopted.
In this view, the inner zone of the Northern Apennines (Fig. 1) represents an ideal area for investigating the above described relationships, i.e. extensional structures, transfer zones (Liotta, 1991; Acocella and Funiciello, 2006; Dini et al., 2008; Liotta and Brogi, 2020) and magmatic bodies (Westerman et al., 2004; Dini et al., 2005; Dini et al., 2008) in a thinned continental crust (22-24 km, Di Stefano et al., 2011, with references therein). This area, in fact, is affected by extensional tectonics since early-middle Miocene (Carmignani et al., 1995), after having experienced overthickening during the Tertiary Alpine orogenesis (Carmignani et al., 2001; Molli, 2008; Rossetti et al., 2015; Bianco et al., 2019).
The methodological approach is based on the integration of: (a) field-mapping at different scales, in order to frame the (i) geometrical relationships between granitic pluton and hosting rocks, and (ii) fracture network, discrete shear zones, kinematics and cross-cutting relationships ??; (b) micro-structural studies by EBSD, in order to define the deformation temperatures on selected samples from ductile shear zones; (c) geochronological analyses and rheological modeling, to define the timing of deformational events during exhumation.

2. Geological setting

Elba (**Figs 1 and 2**) is part of the inner zone of the Northern Apennines, an alpine Alpine? belt deriving from the convergence and collision (Cretaceous-early Miocene) between the Adria microplate, belonging to the Africa plate, and the Sardinia-Corsica Massif of European pertinence (Molli, 2008 for a review; Handy et al., 2010). Collision determined the stacking of oceanic and continental tectonic units deriving from the palaeo-geographic domains of the Northern Apennines (Carmignani et al., 1994; Bianco et al., 2015). Since early-middle Miocene, inner Northern Apennines is affected by eastwards migrating extensional tectonics that consists in two main events (Brogi and Liotta, 2008; Barchi, 2010): (a) the first (early to late Miocene), characterized by an extension of at least 120% (Carmignani et al., 1994; Dallmeyer, and Liotta, 1998; Brogi, 2006; Brogi and Liotta, 2008), gave rise to low-angle normal faults; this event produced the lateral segmentation of the previously stacked tectonic units and the exhumation of mid-crustal rocks (Dallmeyer and Liotta, 1998; Brogi, 2008); (b) the second (Pliocene to Present) is defined by high-angle normal faults, crosscutting the previous structures (Liotta et al., 2010), and determining tectonic depressions where Pliocene to Quaternary continental and marine sediments deposited (Martini and Sagri, 1993; Pascucci et al., 2007; Brogi, 2011). The amount of extension related to these faults is estimated in about 6–7% (Carmignani et al., 1994). The opening of the Tyrrhenian basin (Bartole, 1995; le Breton et al, 2017) and the present crustal and lithospheric thicknesses, (22–24 and 30–50 km, respectively, Calcagnile and Panza, 1981; Locardi and Nicolich, 1992; Di Stefano et al., 2011), are the clearest evidence of the extensional evolution.
Since Langhian, the migration of extension is accompanied by magmatism, mostly of hybrid mantle-crust signature, with an eastward younging direction (Serri et al., 1993; Peccerillo, 2003). Thermal perturbation related to the emplacement into the upper crust of late Miocene-Pliocene plutons,

such as the Porto Azzurro monzogranite **(Fig. 2)** at Elba (Westerman et al., 2004; Caggianelli et al., 2014), produced contact metamorphic aureoles (Zucchi et al., 2017; Caggianelli et al., 2018; Pandeli et al., 2018) and widespread epithermal and mesothermal mineralization through Tuscany and Elba (Dini, 2003) where ore deposits have been exploited for centuries. The emplacement of the Porto Azzurro Pluton is interpreted as controlled by the activity of the Capoliveri-Porto Azzurro transfer zone (Liotta et al., 2015), affecting the southern part of the Elba (**Fig. 2).**

Exhumation of the Elba magmatic bodies is accompanied by faults activity (Westerman et al., 2008): among these, one of the most representative is the Zuccale extensional detachment fault zone, well exposed at Punta Zuccale **(Fig. 2)**, juxtaposing the Ligurian Units on the early–middle Triassic quartzite (Quarziti di Barabarca Fm., in Garfagnoli et al., 2005). The boundary is marked by a flat-lying mineralized extensional shear zone, up to 5 m thick and with a top-to-the-east sense of shear, regionally dipping to the East (Pertusati et al., 1993; Collettini and Holdsworth, 2004; Liotta et al., 2015). The study area is located in the footwall of the Zuccale detachment **(Fig. 3)**.

The Porto Azzurro pluton is a coarse-grained monzogranitic body with K-feldspar megacrysts (**Fig. 4**), poorly exposed in the eastern Elba. The best outcrops are located at Capo Bianco (Marinelli, 1959), flanking the bay of Barbarossa **(Fig. 5)**. The dimension of this pluton is unknown, although gravity data (Milano et al., 2019) suggest an elongated?? shape within the Capoliveri-Porto Azzurro transfer zone (Liotta et al., 2015) and revealing an intrusion even larger than the Monte Capanne pluton **(Fig. 6)**, this latter located in the western Elba (Fig.2) and showing a diameter of *c.* 10 km and a thickness of *c.* 2.5 km (Farina et al., 2010).

The age of the Porto Azzurro monzogranite is estimated at 6.53 ± 0.39 Ma by U-Pb zircon dating analyses (Gagnevin et al., 2011). In addition, ~~Bt~~ biotite Ar-Ar dating of 5.9 ± 0.2 Ma was obtained by Maineri et al. (2003). Magma emplacement took place at pressures of 200 - 175 MPa, as determined from mineral assemblages in the contact aureole (Duranti et al., 1992, Caggianelli et al., 2018). The wall and roof rocks (largely exposed in the Mt. Calamita promontory, **Fig. 2)** are the structurally deepest outcropping rocks of the Elba (Porto Azzurro Fm., in: Garfagnoli et al., 2005). These are mainly represented by micaschist, quartzitic phyllite, quartzite and minor amphibolite levels (Barberi et al.,1967; Garfagnoli et al., 2005). In the whole Mt. Calamita Promontory, and especially in the Barbarossa bay, the micaschist and the monzogranite itself are injected by leucogranite dykes, quartz and tourmaline veins (**Fig. 7**), dissected by later faults (Dini et al., 2002; Dini et al., 2008; Musumeci et al., 2011; Viti et al., 2016; Zucchi et al., 2017).

3. Dataset

We present here the collected dataset following the planned methodological approach, finalized to the reconstruction of the deformation and exhumation history of the pluton. In the Capo Bianco area **(Fig. 5)**, the monzogranite and its hosting rocks are structurally located in the footwall of the Zuccale extensional detachment. The hosting rocks are micaschist and quartzite (see Spina et al., 2019 for information on the protolith), characterized by a well-developed schistosity, showing a general westward to north-westward gently dipping attitude. Such a schistosity is defined? by low-P parageneses, generated during contact metamorphism, up to muscovite-out conditions (Garfagnoli et al., 2005; Zucchi et al., 2017; Papeschi et al., 2017; Caggianelli et al., 2018). For sake of clarity, the data are presented in distinct sections, starting with the petrographical description of the monzogranite.

3.1 Petrography of the monzogranite and felsic dykes

The texture of monzogranite is characterized by a coarse grain size, heterogeneous for the presence of K-feldspar megacrysts up to 10 cm in length. The mineral composition includes plagioclase, K-feldspar, quartz, biotite and minor white mica. Accessory phases are represented by tourmaline, ilmenite, zircon, monazite, xenotime and apatite. K-feldspar frequently shows Carlsbad twinning and perthitic exsolutions (**Fig. 8a**) while plagioclase sometimes displays Albite-Carlsbad twinning and oscillatory zoning (**Fig. 8b**), reflecting the relatively fast magma cooling. SEM/EDS analyses (Appendix: **Table A1 and A2**) indicate that plagioclase is generally zoned with andesine cores (X_{An} up to 0.39) and oligoclase rims (X_{An} down to 0.13). Biotite shows a marked pleochroism with a dark brown tone and the composition is characterized by an average Fe/(Fe+Mg) ratio of 0.56. Biotite frequently includes apatite, monazite and zircon, the latter two being surrounded by sharp metamictic halos (**Fig. 8c).** Quartz - K-feldspar intergrowths, typical of micrographic texture, reflect last melt crystallization **(Fig. 8d)**. Tourmaline is mostly of post-magmatic crystallization, being found in veins or along thin branched fractures. The possible former presence of cordierite (Marinelli, 1959) is revealed by the presence of clots made up of sericite and chlorite.

The felsic dykes, intruding the monzogranite and wall rocks, are characterized by a texture with a medium to fine grain size, heterogeneous for the presence of larger K-feldspars, up to 5 mm in length. The mineral composition includes K-feldspar, plagioclase (An14-16), quartz, tourmaline ± biotite ± white mica ± cordierite, with accessory phases represented by ilmenite, apatite, zircon and monazite. Thus, felsic dykes can be classified as tourmaline leuco-monzogranite. K-feldspar displays Carlsbad twinning and perthitic exsolutions. Plagioclase is present in minor amounts with respect to monzogranite and albite polysynthetic twinning is barely visible. Tourmaline is generally more abundant in the felsic dykes with respect to the monzogranite. SEM/EDS analyses reveal that tourmaline can be classified as Schörl. SEM analyses on rare biotite crystals reveal an elevated content in Fe with an average Fe/(Fe+Mg) of 0.74, distinctly higher than the value found in biotite of the monzogranite. White mica ± chlorite occurs as a common product of alteration at the expense of biotite, felspars and cordierite. Late magmatic muscovite is very rare. Accessory phases are represented by ilmenite, zircon, monazite and apatite. Analysis of zircon in one sample revealed an extremely high concentration of U (UO_2 up to 5.68 wt.%) that can be ascribed to disequilibrium partitioning of U between crystal and melt (Wang et al., 2011).

The preferred orientation of the euhedral K-feldspar megacrysts in monzogranite, well visible in the field (**Fig. 4**) and related to melt-present conditions (Paterson et al., 1989), is barely recognizable at the scale of the optical microscope. Differently, it can be easily recognized in the felsic dykes, even though the K-feldspars are rarely euhedral, owing to subsolidus deformation. All the examined rocks were affected by intense deformation in solid-state conditions both in the ductile and brittle domains.

3.2 - Structural features

Micaschist and monzogranite are deformed by superposed three different faulting episodes **(Fig. 5)**. Tourmaline is present in the damage zone and along the slip-surface of the first two generations of faults, thus suggesting their coeval development, in the frame of a progressive deformation. The faults of the last episode are without hydrothermal mineralization in their damage zones.

(i)?? The first generation of sub-vertical faults, N-S and NNE-striking crossed the monzogranite, related dykes and hosting rocks. Magmatic melt was channeled along these faults, thus resulting in decimeters-thick tourmaline-rich felsic dykes. Within these dykes, a pervasive foliation, parallel to the contact with the hosting rocks, is recognizable (Fig. 7). Tourmaline lineation, formed during the shearing on the main foliation, is more evident close to the boundary between the dykes and the hosting monzogranite (Fig. 7d). Finally, cm-thick quartz-rich veins postdated the felsic dykes (Fig. 7e).

(ii) The second generation of faults consists of sub vertical NE-trending, left-lateral oblique-slip faults exposed in the western part of the Barbarossa ~~Gulf~~ bay (**Fig. 9a,b**) and of sub-horizontal to gently E-dipping normal faults (**Fig. 9c**). Both types of faults dissect all previous structures and are characterized by tourmaline in their shear zones and by offsets ranging from few to tens of meters (**Fig. 3).**

The left-lateral oblique-slip faults are interpreted as minor faults associated to the Capoliveri-Porto Azzurro transfer zone (**Fig. 2**). The low-angle normal faults are subsidiary structures of the Zuccale normal fault **(Fig. 3), affecting** its footwall (Liotta et al., 2015).

Low-angle normal faults are characterized by the presence of up to 30 cm thick mineralized cataclasite: tourmaline mineralization occurs in the monzogranite (**Fig. 9d**), whereas graphite and tourmaline are typically present in micaschist (**Fig. 9e**). Dilatational shear veins filled by tourmaline and subsequent Fe-oxides and/or Fe-hydroxides, typified the damage zone of these latter structures. Kinematic indicators are given by the relationships between the tourmaline and/or graphite-bearing fault zones and associated minor structures, and by mesostructures on the fault-slip plane. These latter are represented by lunate structures, grooves, mega-grooves, slickenlines, quartz-fiber steps, with a top-to-the-east sense of shear, defining a clear normal movement, consistent with the kinematics of the Zuccale normal fault (Pertusati et al., 1993; Keller and Pialli, 1990; Collettini and Holdsworth, 2004).

A significant low-angle normal fault characterizes the Capo Bianco promontory, and affects the roof of the monzogranite (**Fig. 10a,b**). It defines a more than 3 m thick cataclasite level, mostly made up of comminuted micaschist and monzogranite, mineralized by tourmaline and Fe-oxides and/or Fe-hydroxides. Along the main slip surface a mm-thick shear vein, made up of Fe-oxides and/or Fe-hydroxides is recognizable. Along this slip surface a cm-thick felsic dykelet, parallel to the shearing plane, is also hosted **(Fig. 10c).** Furthermore, a similar dykelet is deformed in the cataclasite **(Fig. 10d)**, indicating fault activity during dyke injection. Kinematic indicators are consistent with the Zuccale detachment (**Fig. 10a).**

The occurrence of tourmaline and quartz in both NE-striking left-lateral oblique-slip faults and low-angle normal faults, as well as the deposition of graphite within the cataclasite, indicates that both fault systems were coeval with mineralization. Nevertheless, the comminution of both tourmaline and quartz within the fault zones, strongly suggests that the activity of both sub-vertical and sub-horizontal faults continued after the mineralization event, too.

(iii) The third generation of faults consists of mostly NW-striking subvertical structures. These faults show up to 50 cm thick damage zones with shear fractures and minor faults. The cataclasite is composed of comminuted rock elements, ranging in size from 0.1 to 3 cm. The angular relationships between minor fractures and the main slip surfaces indicate a dominant right-lateral shear component (**Fig. 11a,b**), as it is also suggested by the occurrence of extensional jogs (**Fig. 11c**). Slickenlines on fault planes are oblique, with pitches ranging from 120° to 140° (**Fig. 11d,e**).

3.3 Microstructural analysis

On the basis of the structural setting described above, we have selected the rock-samples targets? to be studied microstructurally. These are felsic dykes and quartz-rich hydrothermal veins, as well as the monzogranite they intruded along first fault generation, and where ductile deformation concentrated. We present EBSD data of two key samples to highlight microstructural features constraining the distinct deformation conditions.

The first sample (Fig. 7c, sample RG2) consists of a foliated felsic dyke with a maximum thickness of c. 20 cm, crosscutting the monzogranite. The dyke shows a clear mylonitic fabric with evidence of dextral extensional shear, resulting in strongly boudinaged K-feldspar. In **Fig. 12a**, a boudinaged Kfs-clast with recrystallized necks and deformed tails, is enveloped by dynamically recrystallized 10 to 50 μm large quartz crystals. Plagioclase recrystallized to significantly smaller grains (< 10 μm) that are interlayered between the Kfs clasts and the quartz tails.

EBSD analysis was applied on five selected sites of sample RG2 to document the deformation microstructures, in terms of the type of crystallographic preferred orientations (CPOs). This approach will allow us to get an insight into the activated slip systems within quartz and Kfs during plastic deformation and hence the deformation temperatures.

Figure 12a shows that boudinaged Kfs clasts are twinned according to the Carlsbad law, and the recrystallized grains within the neck have variable misorientations.

Four of the five selected sites for quartz CPO analysis are in the Kfs-rich parts of the dyke, whereas the fifth is located in a quartz-dominated vein containing boudinaged tourmaline and few strongly deformed Kfs crystals. **Figure 12b** reveals that most of the recrystallized quartz grains have a reddish color and hence a strong attitude to have their *c*-axes aligned towards the observer. The crystallographic orientations of all quartz grains within the analyzed sites give identical indications.

Pryer (1993) suggests that Kfs plastic deformation is only expected at temperatures exceeding 600 °C. We have tried to reconstruct from the EBSD map the possible slip systems that may have operated during the ductile necking of the Kfs clast. For this purpose, we have constructed a pole figure plot that considers just one analytical point for every recrystallized Kfs-grain in the neck (**Fig. 13**). The most likely operating slip system during Kfs boudinage was (110)1/2[1-12]. Dispersion of the slip directions in an ill-defined girdle may suggests dislocation creep to have been accommodated by grain boundary sliding and possibly by diffusion controlled mass transfer process. The latter is suggested by the growth of Kfs-fringes in the strain shadow of the large grey undeformed plagioclase crystal (on the lower right of the **Fig. 12a**), which must have been in a hard slip position.

TEM analysis of experimentally deformed Kfs crystals (Willaime et al., 1979; Scandale et al., 1983), deformed in the temperature range between 700 and 900 °C and at strain rates of $2x10^{-6}sec^{-1}$, shows that the easiest slip systems in Kfs are those with glide along the cleavage planes (010)[001] and (010)[101]. Another possible slip system is (110)1/2[1-12], coherent with our data.

Deformation temperatures consistent with 600 °C are also supported by the recrystallized quartz grains. This is confirmed by the pole figures shown in **Fig. 13,** where quartz c-axes maxima are concentrated in the central part, and poles to a-prisms {11-20} and m-prisms {10-10} are aligned along the primitive great circle. Schmid and Casey (1986) have shown that in metamorphic rocks deformed at normal geological strain rates, these quartz pole figures are characteristic of the upper amphibolite facies conditions, and are consistent with temperatures of about 600 °C, already suggested by the plastically deformed Kfs clasts.

Concerning the second sample (RG4), we have studied the transition between the foliated monzogranite and a tourmaline-bearing quartz-rich hydrothermal vein where an evolution from a mylonitic to ultramylonitic fabric is observed. **Figure 14a** shows the areas analysed in the scanned thin section (at plane polarized light) with a closely flat running foliation in the ultramylonitic zone (F**ig. 14b**). In the upper third of the thin section, larger biotite flakes highlight the foliated monzogranite. In the central mylonitic part, biotite has become strongly deformed and altered. We present data from the two microstructurally distinct areas (red squares, in **Fig. 14a,b)**, that progressively accommodate increasing amounts of deformation in the mylonitic part (Fig. 15a,b), which allow us to understand the deformation mechanisms and temperature conditions affecting this micro shear zone. The CPOs of these recrystallized quartz grains will then be compared with those of sample RG4, previously described and which deformed at about 600 °C.

Figure 15a, an Inverse Pole Figure (IPF) color coded orientation map, shows how quartz of the foliated monzogranite deforms when it meets the upper shear zone boundary. Large grains of the quartz layer show only limited internal deformation in form of arranged subgrain-boundaries (upper part of **Fig. 15a**). Grain size reduction is confined to high stress zones, distinguished by fracture propagation. Here, recrystallization results in a typical grain size smaller than 10 µm. The EBSD orientation map shows that recrystallized grains deriving from different quartz domains mix up and assume misorientations among grains from less than 2° to > 60°. Mixing up of recrystallized quartz grains originally belonging to distinct large quartz grains suggest that grain boundary sliding has been important during this incipient deformation (Ishii et al., 2007).

In the central part of the mylonitic shear zone, grain size reduction of quartz is pronounced (**Fig. 15b**), even though few larger remnant grains, showing core-mantle microstructures, are still preserved. Plots of the EBSD data in pole figures **(Fig. 13e**) show a distribution of c-axes characteristic for a type-I crossed girdle according to Lister (1981), suggesting the activation of basal <a>, rhombs <a> and prism <a> slip systems, typically operating at greenschist facies conditions during metamorphism and therefore at temperatures of about 450 °C. The intense strain gradient that evolved within the shear zone, resulting in complete recrystallization and progressive grain size reduction towards the ultramylonitic part of the shear zone, suggests activation of a strain softening mechanism. Considered that the analyzed shear zone nucleated on a tourmaline-bearing quartz-rich hydrothermal vein, it is reasonable that fluids conveyed along the vein, acting as the driving force for strain softening. Hence, ductile deformation associated to these veins and the adjacent monzogranite clearly occurred later and at lower temperature than that having affected the felsic dykes.

3.4. Geochronological and thermo-chronological data

After having defined deformation characteristics and their related temperatures, in order to unravel cooling history and exhumation after monzogranite emplacement, we now focus on the time constraints by U/Pb geochronological data. We consider previous U/Pb SIMS analyses on three magmatic zircon grains from the Porto Azzurro monzogranite (Gagnevin et al., 2011) not robust from a statistical point of view. So, we decided to analyze 79 out of 132 inspected zircon grains from two samples of the monzogranite exposed in Capo Bianco promontory and La Serra locality (**Fig. 2).** We have analyzed them with LA-ICP-MS for U/Pb zircon dating (see section 3.3).

U-Pb results

U/Pb datings were obtained for sample RG14 from the La Serra locality and sample RG12A from the Capo Bianco locality (**Table A3**).

Sixty-one zircon grains were initially studied from RG14 using CL?? imaging in the SEM. Zircon grains are characterized by complex internal features, with rare continuous oscillatory zoning (**Fig. 16a**). Instead, zircon grains show usually domains or well-defined cores characterized by different CL features with respect to the surrounding rims or rim-domains (**Fig. 16a-d**). Brighter irregular surfaces can occur between these different domains (**Fig. 16c,d**).

Thirty-nine U-Pb analyses were performed on thirty-eight selected zircon grains. Spot analyses were located mainly at the outer rims with oscillatory zoning. U-Pb results are mainly discordant with only seven concordant data. $^{206}Pb/^{238}U$ apparent ages range from 5.6 to 7.2 Ma.

Seventy-two zircon grains were inspected for sample RG12A at the SEM for internal features. The images revealed that zircon grains are commonly characterized by two domains with different CL features, locally separated by a brighter thin domain (**Fig. 16e**). Oscillatory zoning is common for the external domains (**Fig. 16e**), whereas the inner portions are characterized by more complex zoning features (**Fig. 16f-h**). Convoluted zoning is common and can locally interest entire grains (**Fig. 16f,g**), whereas sector zoning is less frequent but occupies large portions of the zircon grains **(Fig. 16h).**

Forty U-Pb analyses were performed on thirty-nine selected zircon grains. Spot analyses were located mainly at the outer rims with oscillatory zoning. U-Pb results are mainly discordant with only four concordant data. $^{206}Pb/^{238}U$ apparent ages range from 5.9 to 7.9 Ma.

The isotopic data obtained from the two samples are mainly discordant and the $^{206}Pb/^{238}U$ data define a main peak between 6 and 7 Ma. The large occurrence of discordant data can be due to limitations of the LA-ICP-MS technique, unable to detect low concentrations of ^{207}Pb in young zircon grains. Another possible cause of discordance can be due to the analyses of distinct domains. Zircon shows complex zoning features clearly associated to different growth stages. Although the analyses were mainly located at the outermost rims, likely associated with the last growth stage, we obtained a large age interval.

Combining the U-Pb data with CL images, we observed that the oldest ages **(see Fig. 16h)** are commonly associated with zircon cores showing CL features that are different and discordant with those of the surrounding thin rims, having oscillatory zoning. This observation suggests the presence of xenocrysts and inherited grains, in agreement with Gagnevin et al. (2011). Thus, excluding outlayers (5 on 39 and 8 on 40 data, from sample RG12A and RG14, respectively), the average ages of the two samples are 6.4 ± 0.4 Ma and 6.4 ± 0.3 Ma **(Fig. 17)**, respectively and very close to the Gagnevin et al. (2011) results. A similar U/Pb geochronological result was obtained from the Calamita schists, affected by contact metamorphism (Musumeci et al., 2011).

4. Post-emplacement thermo-rheological evolution

In order to constrain the deformational evolution with time of the above analysed meso- and micro-structures, we simulated thermal and rheological evolution of the Porto Azzurro pluton for 1 Ma after its emplacement, taking into account the thermo-chronological constraints provided by zircon U-Pb and biotite Ar/Ar datings. In the following, we present the thermal and rheological evolution both in terms of static and dynamic approaches, with the aim to highlight the contribution of extensional tectonics in favoring cooling and migration of the brittle/ductile transition towards shallower structural levels.

4.1 Thermal evolution

The early cooling history of the Porto Azzurro monzogranite has been firstly simulated in static conditions through a unidimensional thermal model (Caggianelli et al., 2018) by numerically solving the differential equation:

$$\frac{\partial T}{\partial t} = \frac{K}{\rho C_p}\left(\frac{\partial^2 T}{\partial z^2}\right) + \frac{A}{\rho C_p} \qquad (1)$$

where *T*, *t* and *z* are temperature, time and depth, respectively; *K*, *ρ*, *Cp* and *A* are thermal conductivity, density, specific heat and radiogenic heat production, respectively **(Table 1)**. Crust and lithosphere thicknesses are fixed at 28 and 56 km respectively, in agreement with the supposed condition in Tuscany during late Miocene (Caggianelli et al., 2014). About the pluton, it was assumed a tabular shape with a thickness of 3 km and an initial temperature of 850 °C. The effect of latent heat for magma crystallization was considered in the temperature interval of 850-650 °C **(Table 1)**.

The results obtained by a program in Stella® code are here proposed through a T-t diagram for the first 1 Ma after magma emplacement fixed at 6.4 Ma (**Fig. 18**). In the same diagram, it is presented a second cooling history, ongoing dynamically during pluton unroofing at an initial rate of 5 mm/yr, decreasing exponentially according to a decay constant (*c*) of c. 10^{-6} yr^{-1}, similarly to the model proposed for the Monte Capanne intrusion (Caggianelli et al., 2014), reducing in 1 Ma the crust and lithosphere thicknesses to c. 25 and 53 km, respectively. The second cooling history incorporates the effect of heat advection due to rock exhumation and thus the differential equation, to be solved numerically, becomes:

$$\frac{\partial T}{\partial t} = \frac{K}{\rho C_p}\left(\frac{\partial^2 T}{\partial z^2}\right) + \frac{A}{\rho C_p} - v_z \frac{\partial T}{\partial z} \qquad (2)$$

with

$$v_z = \frac{dz}{dt} = -cz \qquad (3)$$

A slight discrepancy between the two cooling histories is discernible (Fig. 18) once about 0.7 Ma are elapsed from the time of emplacement, since the dynamical model reproduces a slightly faster cooling rate. The available datings of zircon by U-Pb and of biotite by Ar/Ar (see section 2) were plotted on the same T-t diagram. It is assumed that zircon age corresponds to crystallization at *c.* 800 °C, and that biotite age corresponds to a closure temperature of 430 °C, obtainable by Dodson (1973) formula with Ar diffusion parameters by Harrison et al. (1985). It may appear that the biotite closure temperature is too high with respect to the normally adopted values (e.g. 280-345 °C in Harrison et al., 1985), but the resulting number is mostly an effect of the size of the biotite laminae (width of 1-2 mm) and, above all, of the elevated cooling rate (at least 150 °C/Myr) expected for the top of the shallow Porto Azzurro magmatic body. Ar diffusion in biotite was recently simulated through numerical models by Skipton et al. (2018). It results that at 450 °C, Ar retention in biotite is sensibly controlled by the cooling rate. Anyway, modeled cooling histories are compatible with thermochronological data.

6.2 Rheological evolution

Numerical results of the cooling histories have been used to construct simplified rheological evolutions. For the estimation of the brittle strength we used the equation by Sibson (1974):

$$\sigma = \sigma_1 - \sigma_3 = \beta(1-\lambda)\rho g z \qquad (4)$$

where σ is differential stress, β is a dimensionless parameter depending on the frictional coefficient and tectonic regime and λ is the pore fluid pressure value.

For the estimation of the ductile strength we used the power-law dislocation creep equation (see Ranalli, 1995):

$$\frac{d\varepsilon}{dt} = A_c \sigma^{\mathrm{n}} exp\left(-\frac{E}{RT}\right) \qquad (5)$$

where $d\varepsilon/dt$ is the strain rate, σ is differential stress, A_c, *n* and *E* are creep parameters, *R* is the gas constant and *T* is the temperature.

It was assumed a pore fluid pressure factor (λ) of 0.9, a β value of 0.75 adequate for an extensional tectonic context, and a strain rate of 1×10^{-14} s^{-1} during ductile deformation by dislocation creep

mechanism. The flow law parameters of quartzite and granite (Ranalli, 1995) have been used for roof rocks and Porto Azzurro monzogranite, respectively (**Table 2**). Depth-strength diagrams at 300, 600 and 1000 kyr are provided in **Fig. 19**. According to the static model, the B/D (brittle/ductile) transition at 300 kyr is in the roofing rocks at z = 5 km, well above the contact with the underlying monzogranite, located at c. 6.4 km. At 600 kyr the B/D transition deepens (z = 5.6 km) but still remains above the wall rock - pluton contact. However, the lithological change from wall rock to granite generates a passage to the brittle domain and the appearance of a second B/D transition at a depth of 6.8 km. Consequently, at this time the top of the pluton is already entered into the brittle domain. At 1 Myr, the shallower B/D transition disappears and the deeper one drifts to a depth of 7.8 km well within the plutonic body. The dynamic model (**Fig. 19b**) differs in producing shallower B/D transitions and in anticipating the passage to the brittle domain owing to the faster cooling of the pluton and roof rocks.
The evolution of the B/D transitions for 1 Ma starting from magma emplacement is portrayed in the diagram of **Fig. 20** for both the static and dynamic conditions. In the static case, it can be seen that the minimum depth of the B/D transition (c. 5 km) occurs at c. 250 kyr (point x in **Fig. 20a**). Instead, the genesis of the second B/D transition takes place at c. 500 kyr (point z). Finally, at 900 kyr the shallower B/D transition (point y) disappears and all the wall rocks pass to the brittle domain. In the dynamic case, the passage of the top of the monzogranite to the brittle domain (point z in **Fig. 20b**) is anticipated by about 120 kyr (i.e. at c. 380 kyr), whereas the shallower B/D transition culminates after c. 500 kyr at a minimum depth of c. 3.6 km (point x) and disappears after c. 750 kyr (point y). Afterwards, the whole plutonic body, with the exception of the deepest part, migrates into the brittle domain.

5. Final exhumation: (U-Th)/He dating

For the sake of completing the exhumation history of the Porto Azzurro monzogranite (sample RG12), the apatite He ages provide the best constraints on when the exhumation event ceased. In this view, we provided (U-Th)/He thermo-chronology on apatite (**Table A4, in Appendix)** in order to furnish time constraints on the final part of exhumation, after the time interval considered by the model.
The pluton apatite He age is 1.4 ± 0.6 Ma younger than the zircon U-Pb age suggesting that within less than 2 million years after emplacement the pluton had exhumed and cooled. It is likely that over this time the geothermal gradient (g) was still high. Assuming g = 50 °C per km can be estimated the average exhumation rate of c. 3.4 mm/yr is close to the average value determined for the first 1 million years in the modelled cooling history derived in section 4.

Considering that the Miocene crust should be still hot, a thermal gradient of 60 °C per km can be assumed. In this view, an averaged exhumation rate of c. 3.9 mm/yr is derived, higher than the average value of c. 3.4 mm/yr during the first 1 Myr in the modeled cooling history. This suggests an acceleration of the exhumation history in its last part.

6. Discussion

The meso-structural analysis, together with the kinematic study, indicate that the tectonic context in which deformation took place is extensional. In the same framework, injections of felsic dykes occurred in dilatational fractures, affecting the monzogranite and hosting rocks.
Fast exhumation implies two main regional factors: (i) high heat flux, promoting a decrease of rock density and, (ii) relatively high extensional strain rate ($\geq 1\times10^{-14} s^{-1}$, Ranalli, 1995). Since the first factor

can be easily linked to magma emplacement, extensional strain rate is conversely computed considering restored balanced regional geological sections (Carmignani et al.,1994; Dallmeyer and Liotta, 1998), resulting in $3x10^{-14}$ s^{-1}, during Miocene. Locally, in Elba, even greater values can be indicated referring to the Zuccale fault activity, encompassed between 7 and 5 Ma (Westerman et al., 2004; Dini et al., 2008) and resulting in a total throw of 6 km (Pertusati et al., 1993). Considering these data, an average slip rate of about 3 mm/yr at least, corresponding to an extensional strain rate in the order of 10^{-9} s^{-1},?? can be assumed. Collettini et al. (2009) explain these relevant values considering metamorphic reactions caused by fluid-rock interactions and determining talc inducing fault-weakening. In the same line, we have evidence for syn-tectonic injection of melt in the slip zone of the fault at the roof of the monzogranite, thus promoting fault-weakening, too. A further element acting for fault weakening is represented by the occurrence of graphite along the slip surfaces and within the damage zone of normal faults affecting the hosting rocks (Liotta et al., 2015) and unroofing the Porto Azzurro monzogranite.

In the time interval from 6.4 to 5 Ma, exhumation rate is comprised between 3.4 and 3.9 mm/yr, as estimated on the basis of geochronological, thermo-chronological data and on the modeling. It exceeds the so far determined values for inland Tuscany (Balestrieri et al., 2003, 2011; Fellin et al., 2007; Thomson et al., 2010; Abbate et al., 1994; Carminati et al., 1999; Coli, 1989), where the highest values (1.3 - 1.8 mm/yr) have been determined, on the basis of fission tracks analysis on apatite and zircon, during late Miocene, after and before time ranges?? where exhumation rate is lesser than 1 mm/yr (Balestrieri et al., 2003; Fellin et al., 2007; Thomson et al., 2010).

The difference between our exhumation rate and those from literature, is explained as a consequence of: (i) significant enhancement of lithospheric stretching and high heat flux; (ii) localized magma emplacement; (iii) rapid granite unroofing, as suggested by the slip rate of Zuccale fault, and consequent fast cooling rate. Significantly, this localized high exhumation value is framed in a context still subjected to an acceleration of the regional exhumation, during late Miocene (Balestrieri et al., 2003; Fellin et al., 2007; Thomson et al., 2010).

Reasonably, the above mentioned processes are linked to the migration of the brittle/ductile transition toward shallower crustal levels, due to the thermal perturbation induced by monzogranite emplacement. This promoted the weakening of the upper crust, and together with melt and hydrothermal fluid migrations, acceleration of fault slip rates.

Field evidence indicates that the brittle/ductile migration was a quick event, since ductile deformational features are limited to localized melt-assisted shear zones, followed by high temperature fluid flow within discrete zones.

In fact, microstructural analysis of the felsic dyke **(Fig. 7c**) suggests high deformation temperatures. These allowed boudinaged Kfs to recrystallize ductilely, quartz to deform by prism <a> slip and hence consistent with ductile Kfs-recrystallization. Prism <c> slip at geological strain rates is typical for granulite facies conditions (Mainprice et al., 1986), whereas prism <a> slip is consistent with upper amphibolite facies conditions (Schmid and Casey, 1986) and therefore at about 600 °C.

The microstructural interrelationships support that deformation of the felsic dyke occurred after melt-present monzogranite deformation. Strain softening due to fluid flow along the dykes will have allowed faster strain rates as that typically registered during main geological deformation events, and will have been in the order of $1x10^{-9}s^{-1}$ (e.g. Tullis et al., 1973; Kruhl, 1998; Okudaira et al., 1998), a value that is comparable with the one from the Zuccale fault??.

Quartz-rich hydrothermal vein sample RG4 shows deformation microstructures formed during lower temperature conditions than those preserved within sample RG2. In sample RG4 we have studied the transition from foliated monzogranite to the tourmaline-bearing quartz-rich vein which is accompanied by the evolution from a mylonitic to an ultramylonitic fabric. **Figure 7e** suggests that mylonitic deformation of the monzogranite next to the vein, as well as of the vein itself, has been

triggered by fluid-induced weakening. Since dyke and vein are close to another (just 10 meters), deformation along both discrete shear zones must have occurred at different times after the monzogranite crossed the brittle-ductile transition boundary during cooling and exhumation. By considering the thermo-rheological evolution depicted in **Figures 18 and 20b**, it can be deduced that the monzogranite, close to the roof rocks, entered the brittle field at 480 °C when 380 kyr were elapsed after pluton emplacement. This is a minimum estimate for the timing of the dyke injection and dyke deformation, considered that a brittle behavior of the monzogranite is necessary for a dyking process.
Quartz CPOs of the mylonitic monzogranite next to the tourmaline-bearing quartz-rich vein suggest deformation temperatures of about 450 °C **(Fig. 15a).** This implies that hydrothermal vein deformation has occurred c. 550 kyr after monzogranite emplacement **(see Fig. 18).** A deformation temperature difference of >100 °C between the dyke and the vein is supported by the quartz pole figures that suggest rhombohedral <a> slip together with basal <a> slip and subordinated prism <a> slip (Passchier and Trouw, 2005) for deformation of the vein.
This reconstruction unveils that the fast transition into the brittle domain, still involves ductile deformation, in contrast with what is expected for quickly cooling upper crustal granitic plutons (Caggianelli et al., 2000; Moyen et al., 2003; de Saint-Blanquat et al., 2006; Gonzáles Guillot et al., 2018; Liu et al., 2020).

7. Conclusions

The integration among meso- and microstructural analyses, geochronological, thermo-chronological studies and modeling allows us to state that Porto Azzurro monzogranite emplaced, at 6.4 ± 0.4 Ma in the upper crust (about 6.5 km depth) in extensional setting. Consequently, it experienced fast cooling and a quick transition into the brittle regime. Nevertheless, this is marked by ductile deformation, within discrete melt- and fluid-assisted shear zones, as highlighted by EBSD data of ductilely deformed K-feldspar and quartz.
The thermo-rheological model indicates that the upper part of the monzogranite entered into the brittle domain *c.* 380 kyr after its emplacement, allowing dyke formation and melt-injection that triggered localized high strain ductile deformation. The same happened *c.* 400 kyr later, when even the roof rocks entered into the brittle domain, and quartz-rich veins deformed at lower temperatures, most~~ly~~ likely due to enhanced fluid-flow within the localized shear zones.
The cooling history of the pluton took place during fast exhumation, with an estimated rate of c. 3.4-3.9 mm/yr, and was promoted by extensional tectonics and correlated unroofing faults, during late Miocene.
The key-message of this paper is that ductile deformation can affect a granitic intrusion even when it is already entered into the brittle domain, due to localized thermo-rheological perturbations, caused by late magmatic events, in an extensional regime.

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Appendix: Analytical methods

During this investigation we used scanning electron microscope (SEM) for: (i) quantitative chemical analysis; (ii) electron backscattered diffraction (EBSD) analysis; (iii) backscattered electron (BSE) imaging, and (iv) CL-imaging; by using: an EVO Zeiss SEM at Bari University; a CamScan 2500 SEM equipped with EDX and NordlyseNano EBSD detectors at University of Padova; and, a Philips XL30 electron microscope equipped with a Centaurus CL detector at Pavia CNR-IGG-UOS, respectively. In situ zircon U-Pb geochronology was determined by LA-ICP-MS at CNR-IGG-UOS of Pavia.

In situ U-Pb geochronology was determined by excimer LA-ICP-MS at CNR-IGG-UOS of Pavia. Zircons were separated by conventional methods (crushing, heavy liquids, hand picking) from two samples (RG12; RG14). Prior to age determination, the internal structure of the zircon grains was investigated with backscattered electron (BSE) and Cathodoluminescence (CL) images using a Philips XL30 electron microscope equipped with a Centaurus CL detector. Images were obtained using 15 kV acceleration and a working distance of 26 mm.

Age determinations were performed using a 193 nm ArF excimer laser microprobe (GeoLas200QMicrolas) coupled to a magnetic sector ICP-MS (Element 1 from ThermoFinnigan). Analyses were carried out in single spot mode and with a spot size fixed at 25 mm. The laser was operated with a frequency of 5 Hz, and with a fluence of 8 J/cm2. Sixty seconds of background signal and at least 30 s of ablation signal were acquired. The signals of masses 202Hg, 204(PbHg), 206Pb, 207Pb, 208Pb, 232Th, and 238U were acquired in magnetic scan mode. 235U is calculated from 238U based on the mean ratio 238U/235U of 137.818, as recently proposed by Hiess et al. (2012). The 202 and 204 masses were collected in order to monitor the presence of common Pb in zircon. Mass bias and laser-induced fractionation were corrected by adopting external standards, the GJ-1 zircon standard (608.56 ± 0.4 Ma; Jackson et al., 2004). During an analytical run of zircon analyses, a reference zircon 91500 (Widenbeck et al., 1995) was analyzed together with unknowns for quality control. Data reduction was carried out through the GLITTER software package (Van Achterbergh et al., 2001). Time-resolved signals were carefully inspected to detect perturbation of the signal related to inclusions, cracks, or mixed-age domains. Within the same analytical run, the error associated with the reproducibility of the external standards was propagated to each analysis of sample (see Horstwood et al., 2003), and after this procedure each age determination was retained as accurate within the quoted error. The Concordia test was performed for each analytical spot from 206Pb/238U and 207Pb/235U ratios using the function in the software packyrge Isoplot/Ex 3.00 [Ludwig, 2003]. Percentage of discordance has been calculated as {[1-(206Pb/238U age/207Pb/235U age)] x 100}. Errors in the text and figures are reported as 2sigma. The Isoplot software was also used to draw diagrams of age data. U-Pb isotope analyses and calculated ages of zircons are reported in the data repository (Tables A1 and A2).
Single crystal apatite (U-Th)/He dating has been undertaken at Scottish Universities Environmental Research Centre. Individual apatite grains were screened based on their clarity and morphology, and hand-picked for (U-Th)/He analysis then packed into Pt tubes prior to analysis. Helium, U and Th analytical protocols adopted in this study follows those described by Foeken et al. (2006, 2007). Length and width measurements for alpha ejection correction (FT; Farley, 2002), were taken for each grain. (U-Th)/He dates were calculated using standard procedures developed by Meesters and Dunai (2002). Total analytical uncertainty, computed as a square root of squares of weighted uncertainties of U, Th and He measurements, and including the estimated additional variation of ± 7% determined on repeat analyses of Durango apatite.

APPENDIX CAPTIONS

Table A1 – Selected analyses of mineral phases in monzogranite (LCA 14) and in a felsic dyke (LCA 7).

Table A2 – Selected analyses of accessory phases in manozogranite (LCA 14) and felsic dyke (LCA 7).

MANUSCRIPT CAPTIONS

Fig. 1 - Structural sketch map of Northern Tyrrhenian Basin and Northern Apennines. The main Pliocene–Quaternary basins, transfer zones and Neogene–Quaternary intrusive bodies are indicated (after Dini et al., 2008).

Fig. 2 - Geological sketch map of Elba Island. The main faults, the study area and the location of the dated (^{206}Pb/^{238}U in zircon) samples (RG12 and RG14) are indicated.

Fig. 3 – a) Google Earth photograph of the Barbarossa ~~gulf~~ bay where the analysed monzogranite is exposed; the main structural elements and the mineralised volume corresponding to the Zuccale Fault is also indicated; b) geological sections across the bay with the main structural

elements; c) panoramic view of the eastern side of the bay with the main faults and the analysed monzogranite; d) detail of an extensional detachment fault accompanying the deformation in the footwall of the Zuccale Fault; the stereographic diagram (lower hemisphere, equiareal projection) illustrates the poles of the minor faults linked to the detachments and their kinematics.

Fig. 4 – a) Aligned K-feldspar megacrysts defining the magmatic foliation of the monzogranite; b) details of twinned K-felspar megacrysts; c) panoramic view of the Capo Bianco promontory with the tectonic relations of the monzogranite and the hosting rocks.

Fig. 5 - Geological map of the Capo Bianco promontory. Stations of structural analyses and relative data are reported in stereographic diagrams (lower hemisphere, equiareal projection). The location of the analysed samples for microstructural analyses is also indicated. Structural and kinematic data of each fault generation, as described in the text, are reported in stereographic and rose diagrams (lower emisphere, equiareal projection).

Fig. 6 – Bouguer gravity field after Milano et al. (2019). Isolines are in mGal. NE-trending faults are matching the low gravimetric trend, that is here related to a regional transfer zone where magmatic bodies emplaced.

Fig. 7 – Photographs illustrating felsic dykes intruded within mechanical discontinuities formed during first generation faulting: a) dm-thick felsic dykes intruded within micaschist affected by second generation extensional faults; b) felsic dyke intruded within the monzogranite sampled for microstructural analyses (sample RG2); c) detail of the felsic dyke shown in fig. (c) illustrating the internal foliation and s-c structures; d) L-tectonite consisting of tourmaline lineation developed on the felsic dyke-surface foliation.

Fig. 8 - Micrographs of petrographic features of the Porto Azzurro monzogranite in plane polarized (PPL) and crossed polars (CP) light. a) Karlsbad twinned K-feldspar with perthitic exsolution lamellae; b) plagioclase with oscillatory zoning (CP image); c) basal section of biotite including elongated apatites and tiny zircon and monazite crystals surrounded by metamictic halos (PPL image); d) K-feldspar-Quartz granophyric intergrowth (CP image).

Fig. 9 - Photographs illustrating details of the second generation faults in the Barbarossa bay and Capo Bianco promontory: a) meter-thick tourmaline-rich mineralised fault at the boundary between the monzogranite and the host rocks; b) shear veins parallel to cm-thick fault zones mineralised by quartz and tourmaline, dissecting a dm-thick felsic dyke; c) panoramic view of the low-angle normal faults characterizing the footwall of the Zuccale Fault; d) detail of a fault zone mineralized by tourmaline and quartz characterizing the footwall of the Zuccale Fault and affecting the monzogranite; e) detail of a fault zone mineralized by Fe-oxyhydroxides characterizing the footwall of the Zuccale Fault and affecting the micaschist in the fault system shown in (c).

Fig. 10 - Photographs illustrating faults of the second generation: a-b) panoramic view of the extensional fault system separating the monzogranite from the roof rocks (micaschist) characterizing the footwall of the Zuccale Fault in the Capo Bianco promontory; c) detail of a cm-thick felsic dykelet assisting deformation along the slip surface of the fault zone illustrated in (a) and (b); d) felsic dykelet injected in the cataclasite and ductilely deformed during faulting.

Fig. 11) - Photographs illustrating details of the third generation faults affecting the monzogranite in the Capo Bianco promontory: a-b) horsetail fractures developed in the tip-damage zone of a right-lateral strike-slip fault; c) dm-sized extensional jog formed in the linking?? damage zone of two overstepping right-lateral strike-slip faults; d-e) detail of a right-lateral strike-slip fault zone and minor (i.e. splay) structures.

Fig. 12– EBSD misorientation maps of sample RG2. a) Kfs shown with IPF color coding, quartz

distinguished as phase (red) and plagioclase shown in band contrast mode (grey tones). Boudinaged Kfs recrystallizes in the neck to? new grains, showing crystallographic mismatches between 10 and > 90°; b) anastomosing recrystallized quartz bands (IPF color coded) show a large preference for reddish colors, proving that the c-axis distribution of recrystallized grains is oriented towards the observer. Maximum of crystallographic mismatches along grain boundaries is > 30°.

Fig. 13 - Pole figures, equal area projections, lower hemisphere. a) Multiples of Mean Unit Densities expressed by color bars. Bar with maxima just above 7 belong to pole figures B,C,E, while bar with maximum at 14 belongs to p~~l~~ole figure D; b) pole figure of recrystallized grains in the neck of the boudinaged Kfs (Fig.12a). White line is foliation trace. Maximum of {110} poles is consistent with this plane being the slip plain, while dispersion of slip direction <1-12> in a girdle centered on that maximum suggests that dislocation creep was accommodated by grain boundary sliding; c) quartz c-axis of grains within anastomosing bands within sample RG2 form a maximum in the central part of the pole figure, while poles to a- and m-prisms are dispersed along the equatorial section of the pole figure; d) distribution of c-axes within recrystallized quartz-rich band in sample RG2 forms a pronounced maximum in the center, while poles to a- and m-prisms form single maxima consistent with dextral shear and activation of prism <a> slip; e) quartz c-axis distribution within recrystallized grains of mylonitic monzogranite (sample RG4, Fig 15 b) forms a typical type-I crossed girdle, suggesting activation of basal <a>, rhomb <a> and prism <a> slip, and hence lower deformation T than for sample RG2.

Fig.14 - Micrograph of a discrete shear zone localized within Sample RG4 and EBSD analysis areas. a) The foliated monzogranite (top) evolves to a mylonite in correspondence of EBSD analysis area Fig. 15a. There is a deformation gradient from mylonite towards ultramylonite next to the quartz-rich vein with green tourmaline clasts; b) sketch of Fig. 14a highlighting the different deformational domains: G = foliated monzogranite; M = mylonite ; UM = ultramylonite. Location of Figures 15a and b is also shown.

Fig. 15. EBSD misorientation maps (IPF color coding) of the areas indicated in Fig.14. a) Monzogranite next to the mylonite boundary. Large quartz crystals are still preserved showing evidence of sub-grain boundaries. Recrystallization starts at sites of stress localization. Crystallographic mismatches are between 10 and >60°. The small grain size and evidence for mixing up of recrystallized grains belonging to different quartz domains suggest grain-boundary sliding to have been important; b) monzogranite deformed at higher strain. Quartz is fully recrystallized and forms a mylonitic foliation trending ENE-WSW in the map. Some remnants of larger grains with core and mantle structure are still preserved. Maximum misorientation of adjacent grains is larger than 30°. Variegated colors of grains testify that slip occurred along multiple systems during mylonitic deformation. Non indexed grey phases are Kfs and plagioclase.

Fig. 16 – Selected BSE-CL images of zircon grains from samples RG14 (a-d) and RG12 (e-h) representative of different CL features. Location of LA-ICP-MS spots is shown within zircon grains. Scale bar is 50 μm.

Fig. 17 – ^{206}Pb/^{238}U zircon data for RG14 and RG12 samples. U-Pb data are ordered and shown as vertical bars comprising 2sigma errors. Grey filled bars were considered for calculation of weighted average ages (horizontal hemi transparent boxes). Available U-Pb zircon data, with relative errors, for the Porto Azzurro pluton are also shown: white circles refer to ages from three zircon grains in Gagnevin et al. (2011); black square is the age of a zircon rim from the contact aureole (Musumeci et al., 2011).

Fig. 18 - T-t diagram showing the cooling history of the Porto Azzurro monzogranitic pluton

reproduced for 1 Myr after the magma emplacement fixed at 6.4 Ma at a depth of c. 6.5 km. The two cooling curves result from a static model (black line) and a dynamic model (grey line), incorporating the effect of unroofing simulated by an exponential law (see Table 1). Details of the modelling are provided in Caggianelli et al. (2018) and Caggianelli et al. (2014). Age obtained from zircon is plotted at the corresponding saturation temperature. Age of biotite (Maineri et al., 2003) is plotted at the corresponding closure temperature, calculated by Dobson (1973) formula.

Fig. 19 - Depth - strength diagrams at 300 kyr, 600 kyr and 1 Myr. They show depth of the brittle/ductile transitions and the contact between the Porto Azzurro monzogranite with the roof rocks. (a) Static model; (b) dynamic model incorporating the effect of unroofing.

Fig. 20 - z-t diagram showing the depth change of the rheological boundaries (black and blue lines) and of the monzogranite - roof rocks lithological boundary (grey line), as reproduced by a static (a) and dynamic model (b). The simulation lasts 1 Myr starting from 6.4 Ma and from a magma emplacement depth of 6.5 km. The shallower brittle-ductile (B/D) transition (black line), within the roof rocks, culminates to a minimum depth in correspondence of point x and disappears after point y. The deeper B/D transition (blue line), within the monzogranite, starts from point z and progressively sinks. The co-existence of B/D transitions is confined to the time interval between z and y points. The rheological evolution of the top of the pluton, relevant for the Capo Bianco outcrop, can be followed along the grey line. Thus, the points z and y mark the transition to the brittle domain for the monzogranite and the roof rocks, respectively.

Table 1 - Physical parameters used in the thermal model.

Table 2 - Flow law constants adopted for the dislocation creep exponential equation.

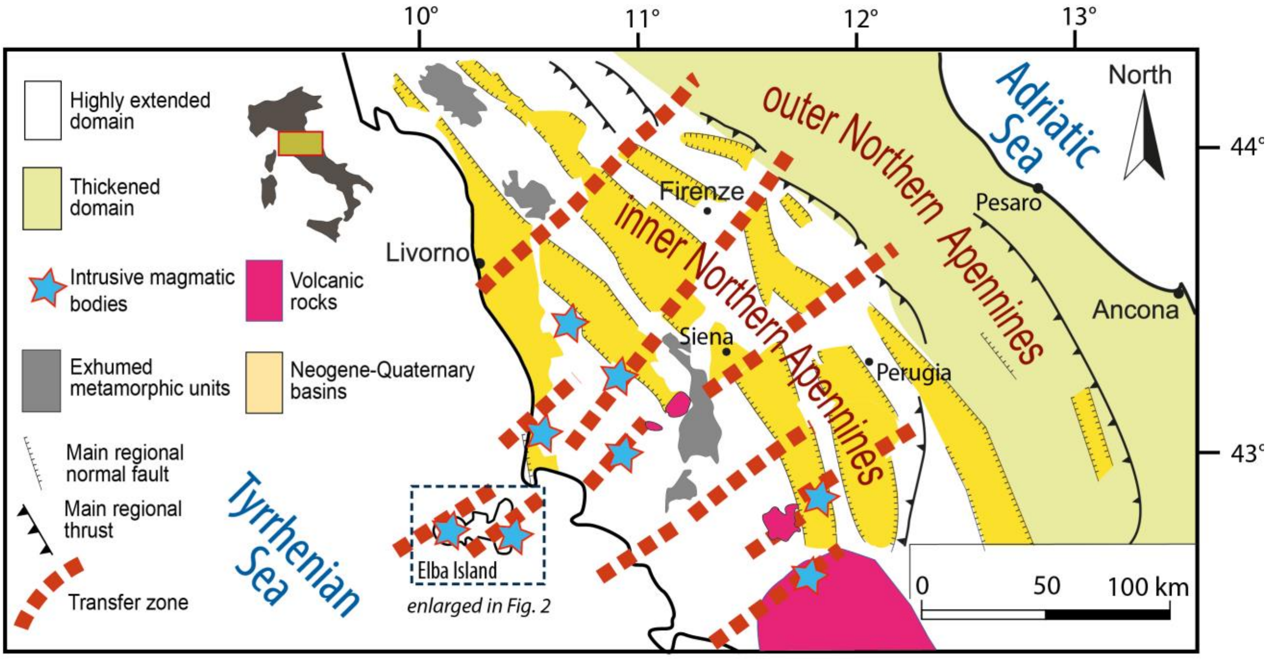

10°
11°
12°
13°
44°
43°
Highly extended domain
Thickened domain
Intrusive magmatic bodies
Volcanic rocks
Exhumed metamorphic units
Neogene-Quaternary basins
Main regional normal fault
Main regional thrust
Transfer zone
Tyrrhenian Sea
Adriatic Sea
North
outer Northern Apennines
inner Northern Apennines
Firenze
Livorno
Siena
Perugia
Pesaro
Ancona
Elba Island
enlarged in Fig. 2
0
50
100 km

Fig.1

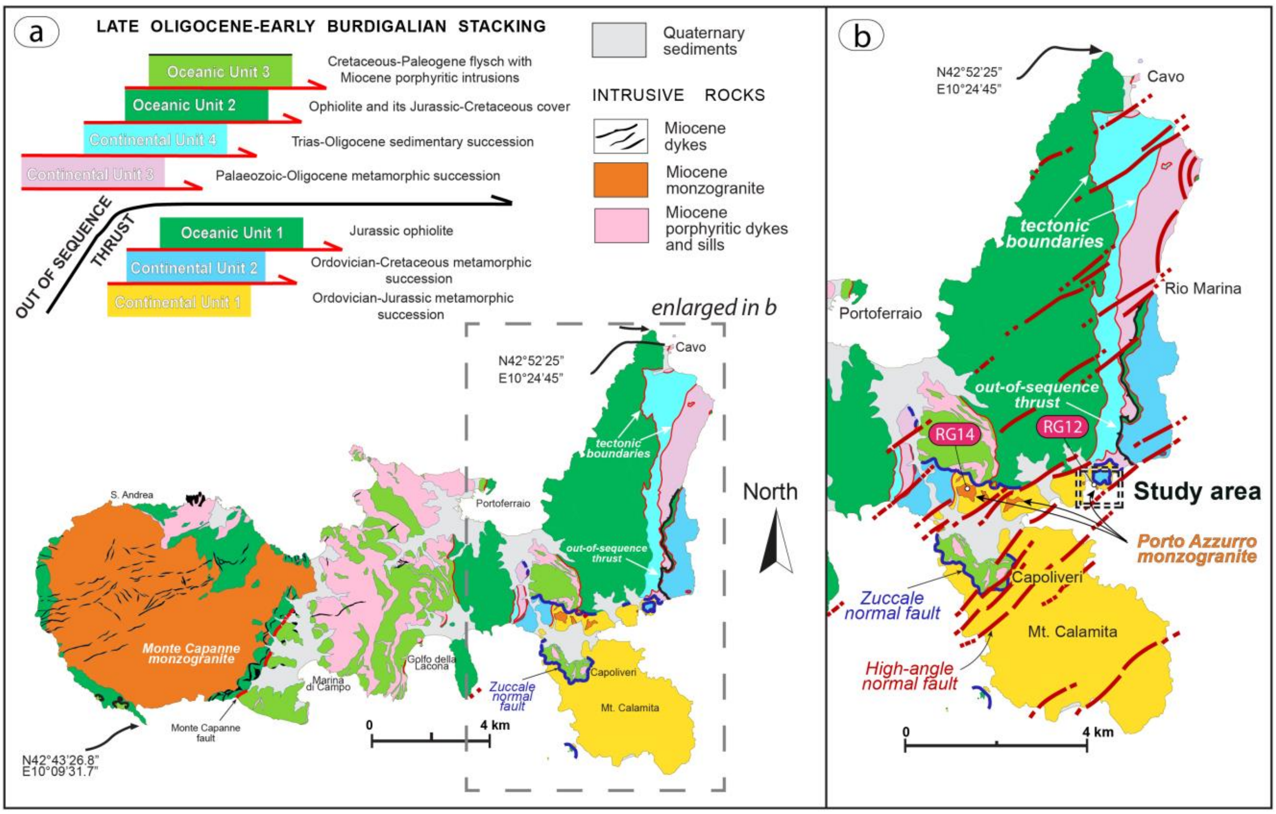
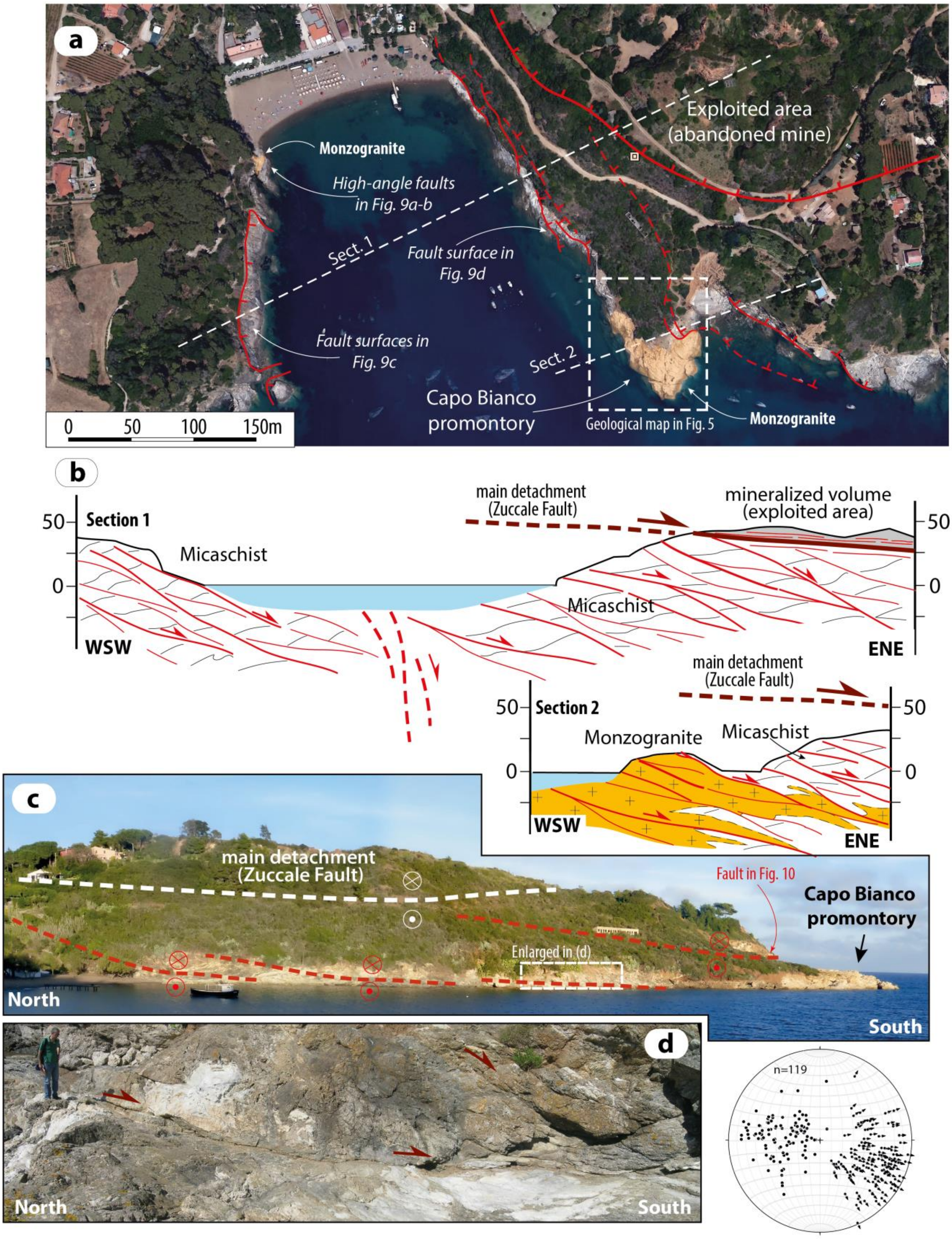

a
LATE OLIGOCENE-EARLY BURDIGALIAN STACKING
Oceanic Unit 3
Cretaceous-Paleogene flysch with Miocene porphyritic intrusions
Oceanic Unit 2
Ophiolite and its Jurassic-Cretaceous cover
Continental Unit 4
Trias-Oligocene sedimentary succession
Continental Unit 3
Palaeozoic-Oligocene metamorphic succession
OUT OF SEQUENCE THRUST
Oceanic Unit 1
Jurassic ophiolite
Continental Unit 2
Ordovician-Cretaceous metamorphic succession
Continental Unit 1
Ordovician-Jurassic metamorphic succession
Quaternary sediments
INTRUSIVE ROCKS
Miocene dykes
Miocene monzogranite
Miocene porphyritic dykes and sills
enlarged in b
N42°52'25"
E10°24'45"
Cavo
tectonic boundaries
Portoferraio
S. Andrea
out-of-sequence thrust
North
Monte Capanne monzogranite
Marina di Campo
Golfo della Lacona
Zuccale normal fault
Capoliveri
Mt. Calamita
0
4 km
Monte Capanne fault
N42°43'26.8"
E10°09'31.7"
b
N42°52'25"
E10°24'45"
Cavo
tectonic boundaries
Rio Marina
Portoferraio
out-of-sequence thrust
RG14
RG12
Study area
Porto Azzurro monzogranite
Capoliveri
Zuccale normal fault
Mt. Calamita
High-angle normal fault
0
4 km

Fig.2



a
Exploited area
(abandoned mine)
Monzogranite
High-angle faults
in Fig. 9a-b
Sect. 1
Fault surface in
Fig. 9d
Fault surfaces in
Fig. 9c
Sect. 2
Capo Bianco
promontory
Geological map in Fig. 5
Monzogranite
0 50 100 150m
b
main detachment
(Zuccale Fault)
mineralized volume
(exploited area)
Section 1
Micaschist
Micaschist
WSW
ENE
main detachment
(Zuccale Fault)
Section 2
Monzogranite
Micaschist
WSW
ENE
c
main detachment
(Zuccale Fault)
Fault in Fig. 10
Capo Bianco
promontory
Enlarged in (d)
North
South
d
North
South
n=119

Fig. 3

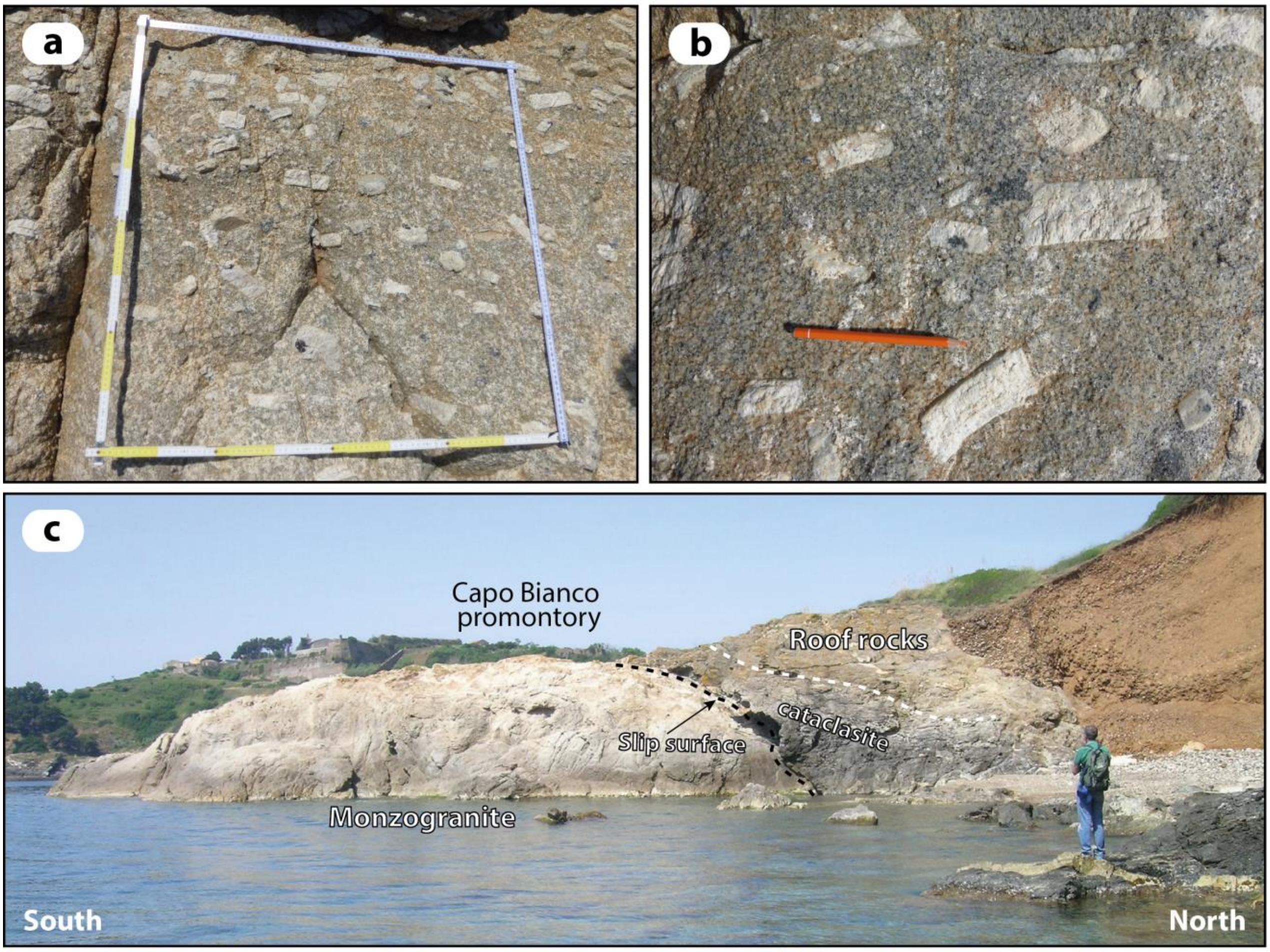
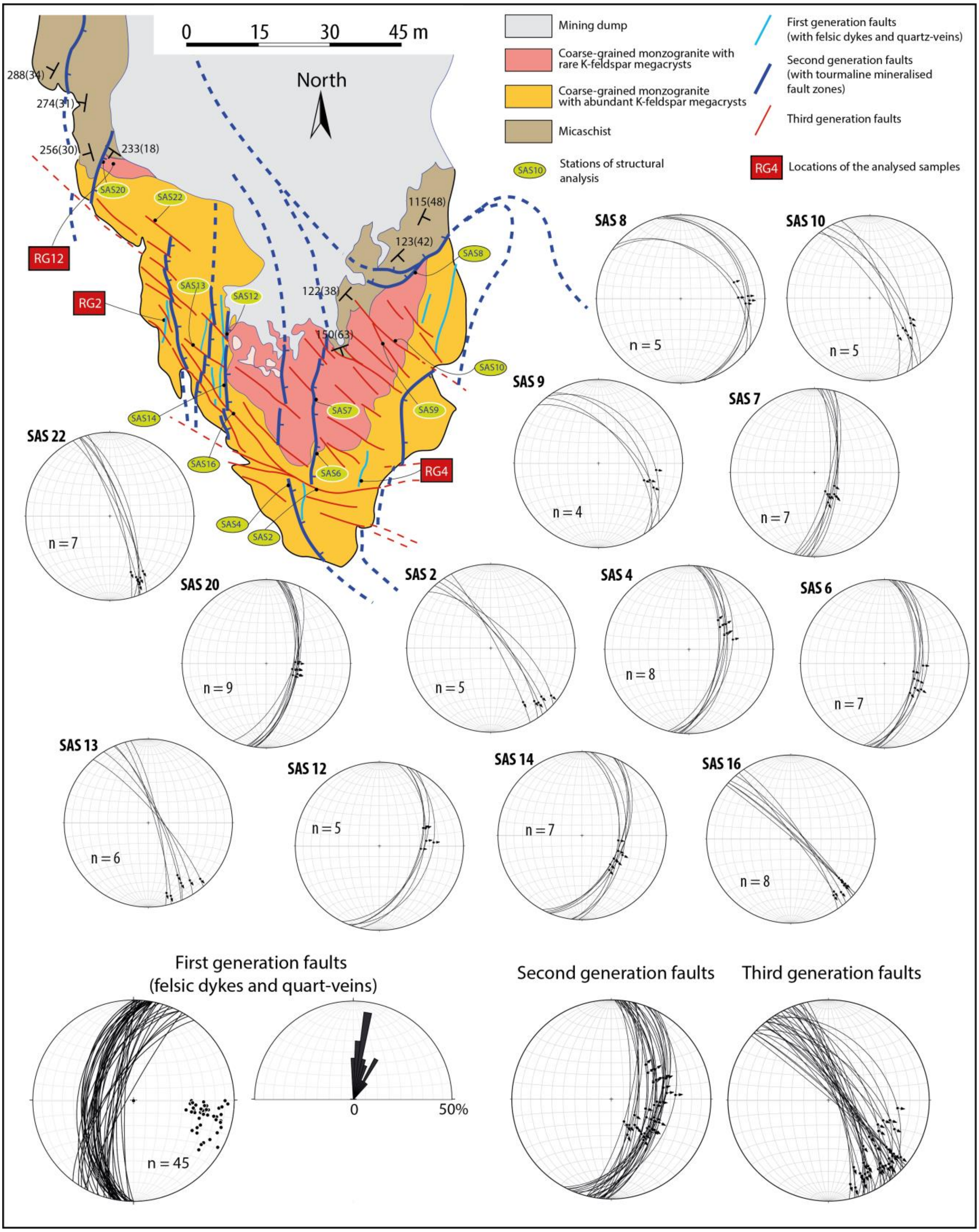

a
b
c
Capo Bianco
promontory
Roof rocks
cataclasite
Slip surface
Monzogranite
South
North

Fig. 4



0 15 30 45 m
North
Mining dump
Coarse-grained monzogranite with rare K-feldspar megacrysts
Coarse-grained monzogranite with abundant K-feldspar megacrysts
Micaschist
Stations of structural analysis
First generation faults (with felsic dykes and quartz-veins)
Second generation faults (with tourmaline mineralised fault zones)
Third generation faults
Locations of the analysed samples
288(34)
274(31)
256(30)
233(18)
115(48)
123(42)
122(38)
150(63)
RG12
RG2
RG4
SAS 8
n = 5
SAS 10
n = 5
SAS 9
n = 4
SAS 7
n = 7
SAS 22
n = 7
SAS 20
n = 9
SAS 2
n = 5
SAS 4
n = 8
SAS 6
n = 7
SAS 13
n = 6
SAS 12
n = 5
SAS 14
n = 7
SAS 16
n = 8
First generation faults (felsic dykes and quart-veins)
n = 45
0
50%
Second generation faults
Third generation faults

Fig.5

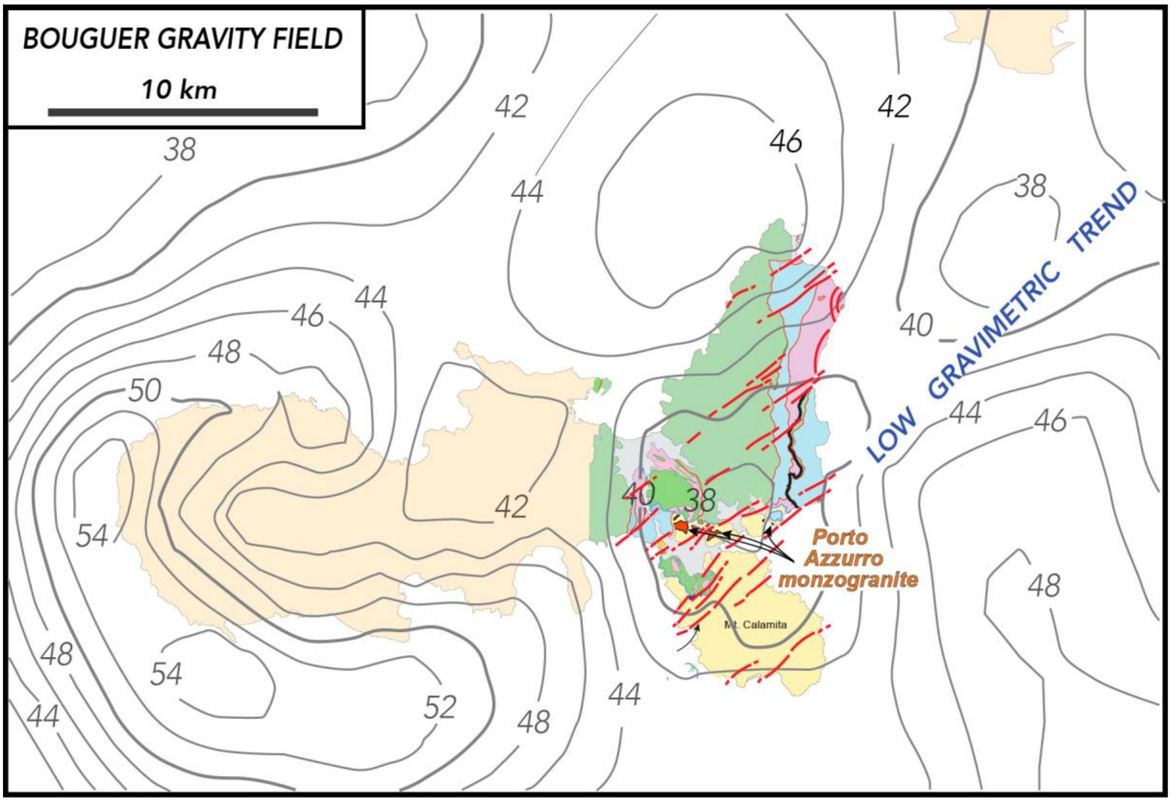

BOUGUER GRAVITY FIELD
10 km
38
42
44
46
48
50
54
54
48
44
52
48
44
42
46
44
42
40
38
38
40
44
46
48
LOW GRAVIMETRIC TREND
Porto
Azzurro
monzogranite
Mt. Calamita

Fig. 6

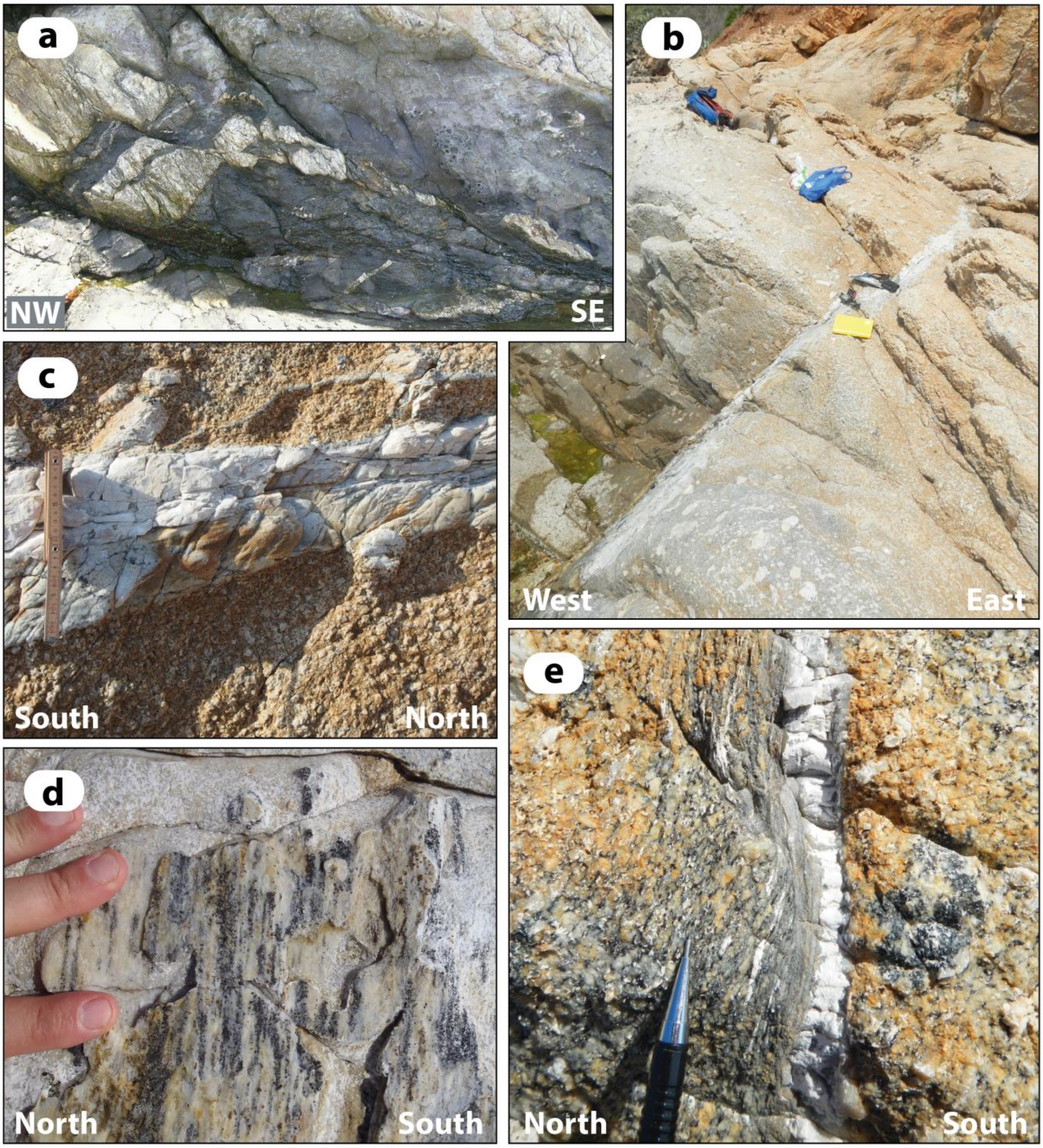


Fig. 7

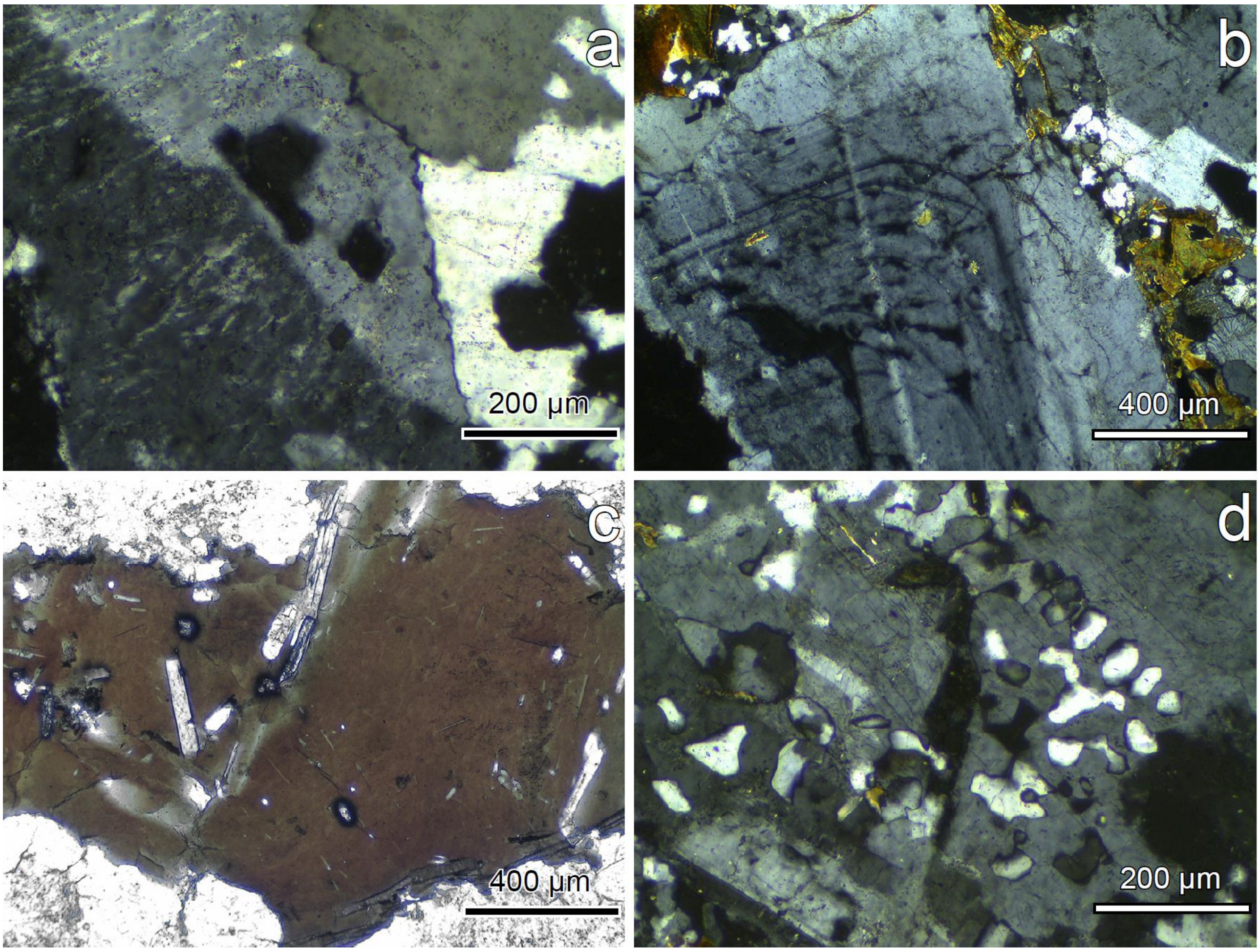


Fig. 8

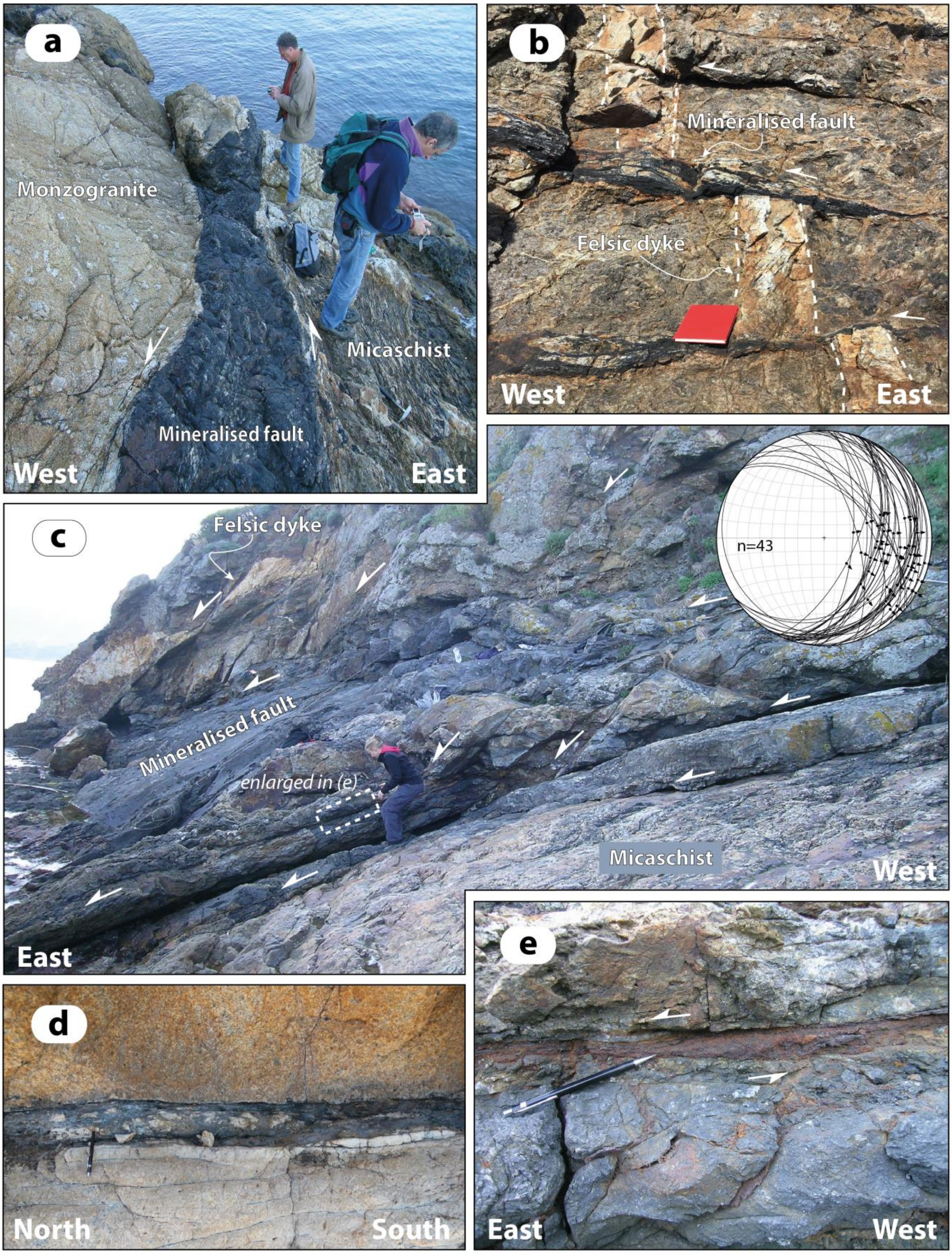

a
Monzogranite
Micaschist
Mineralised fault
West
East
b
Mineralised fault
Felsic dyke
West
East
c
Felsic dyke
n=43
Mineralised fault
enlarged in (e)
Micaschist
West
East
d
North
South
e
East
West

Fig. 9

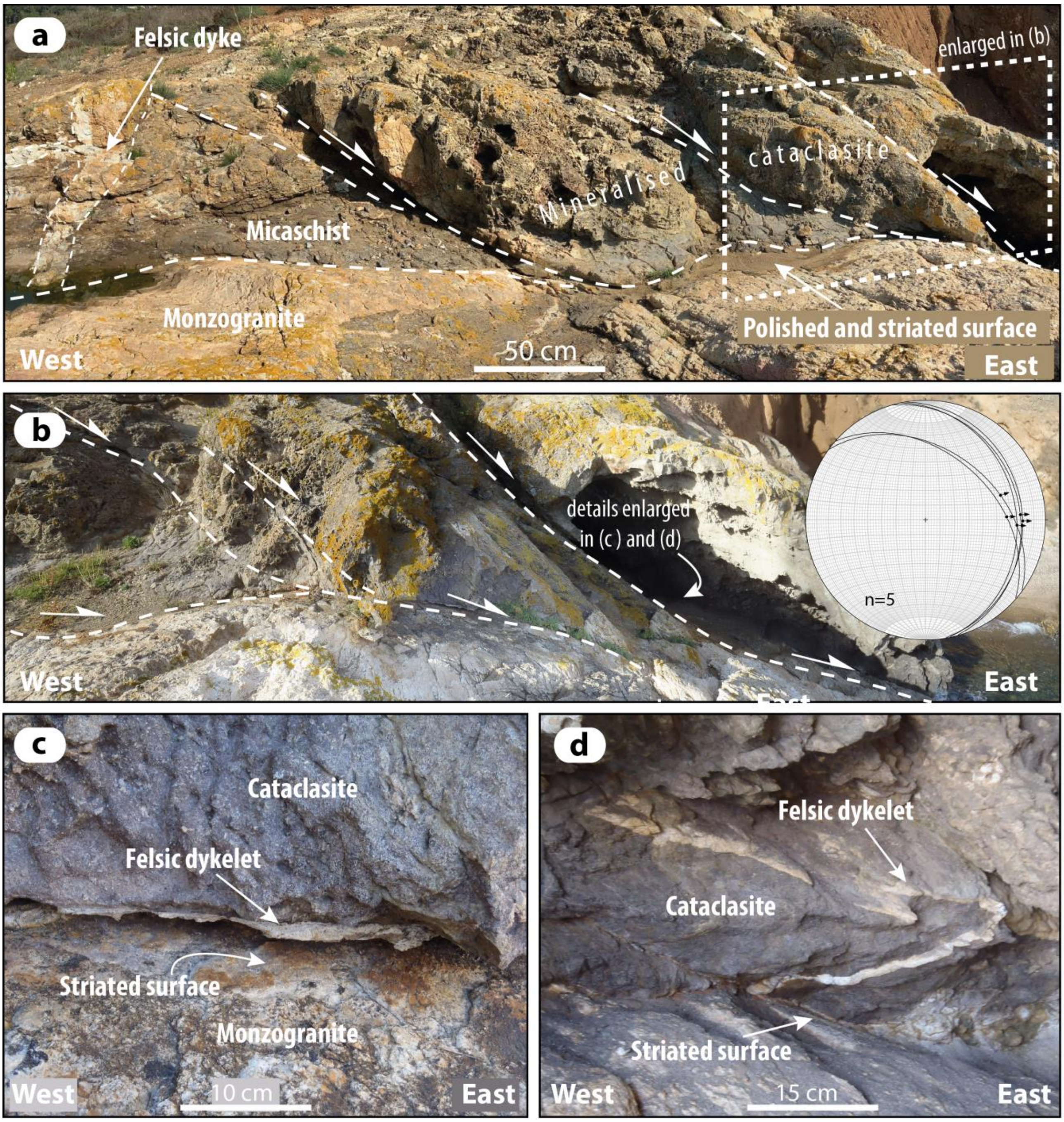
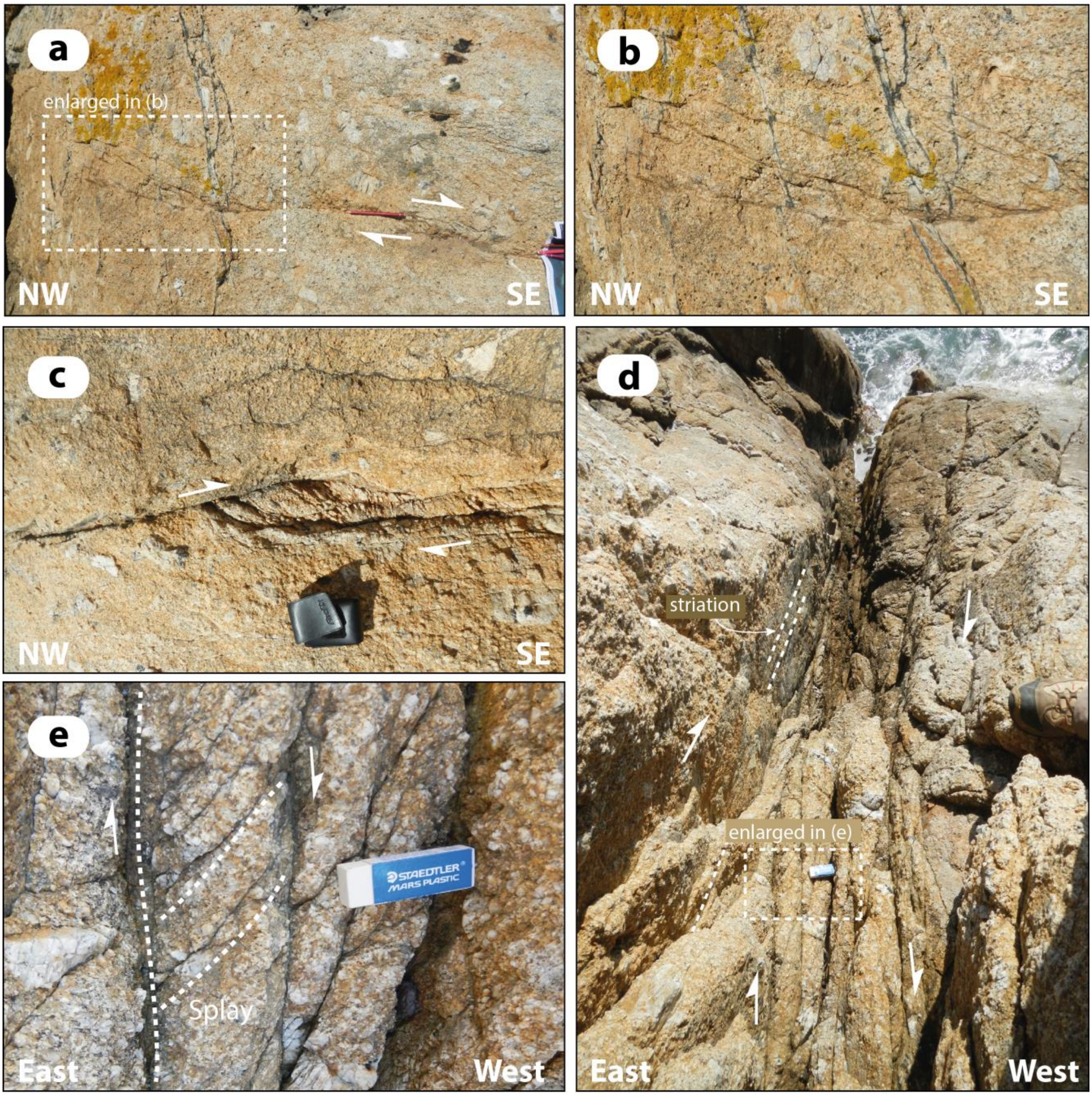

a
Felsic dyke
enlarged in (b)
cataclasite
Mineralised
Micaschist
Monzogranite
Polished and striated surface
50 cm
West
East
b
details enlarged
in (c) and (d)
n=5
West
East
c
Cataclasite
Felsic dykelet
Striated surface
Monzogranite
10 cm
West
East
d
Felsic dykelet
Cataclasite
Striated surface
15 cm
West
East

Fig. 10



a
enlarged in (b)
NW
SE
b
NW
SE
c
NW
SE
d
striation
enlarged in (e)
East
West
e
STAEDTLER
MARS PLASTIC
Splay
East
West

Fig. 11

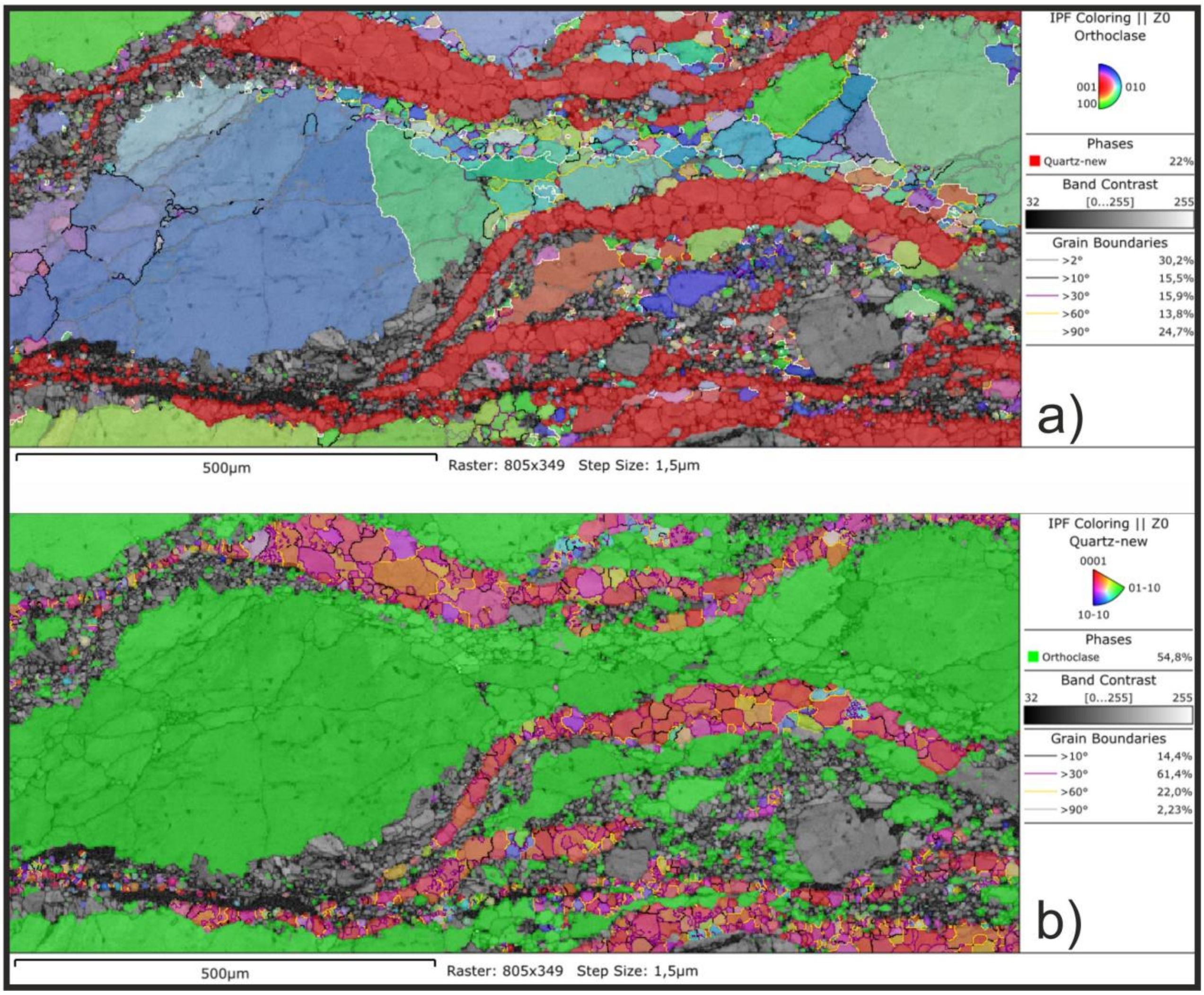
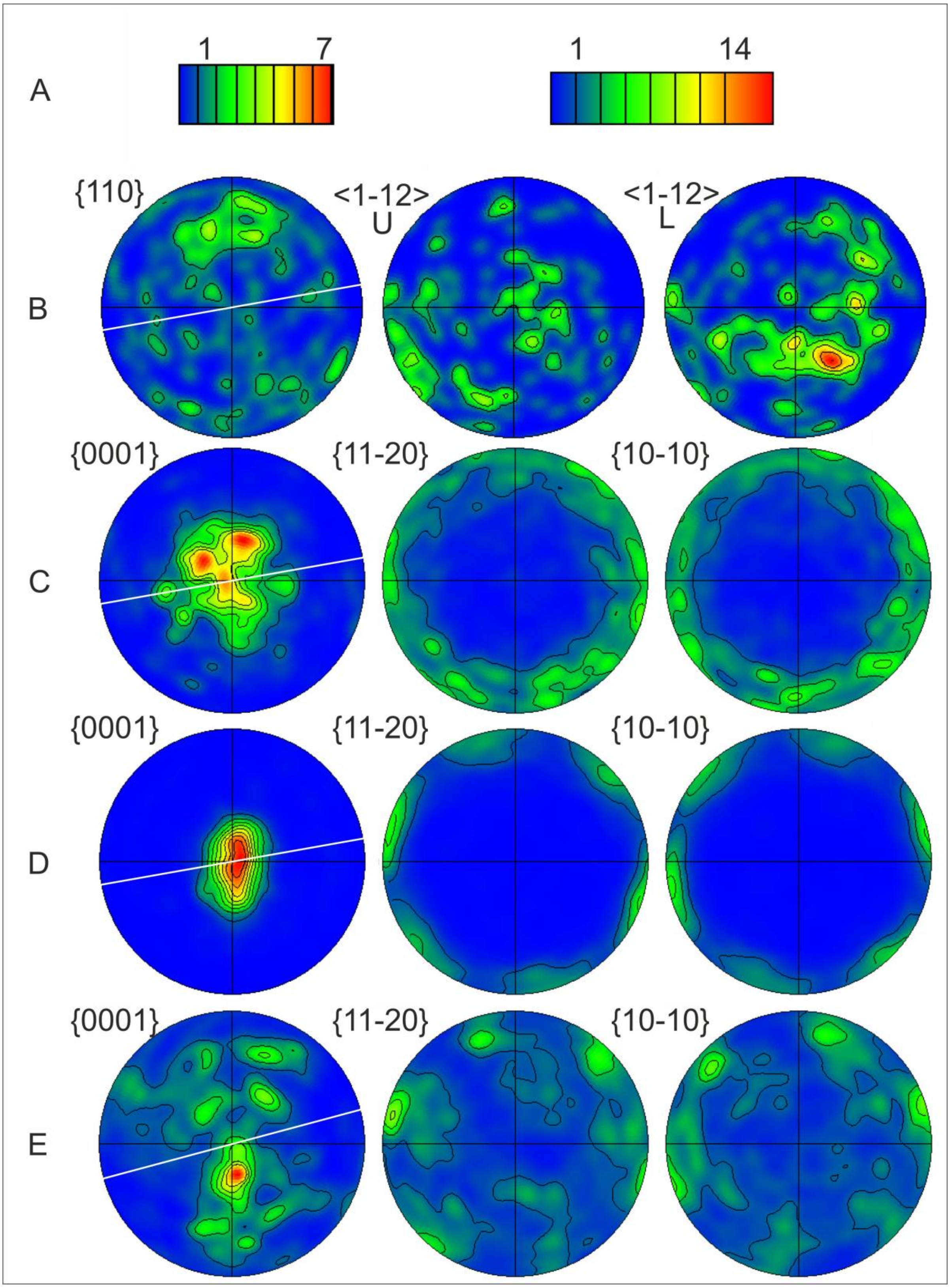

IPF Coloring || Z0
Orthoclase
001
100
010
Phases
Quartz-new 22%
Band Contrast
32 [0...255] 255
Grain Boundaries
>2° 30,2%
>10° 15,5%
>30° 15,9%
>60° 13,8%
>90° 24,7%
a)
500µm
Raster: 805x349 Step Size: 1,5µm
IPF Coloring || Z0
Quartz-new
0001
01-10
10-10
Phases
Orthoclase 54,8%
Band Contrast
32 [0...255] 255
Grain Boundaries
>10° 14,4%
>30° 61,4%
>60° 22,0%
>90° 2,23%
b)
500µm
Raster: 805x349 Step Size: 1,5µm

Fig. 12



A
1
7
1
14
B
{110}
<1-12>
U
<1-12>
L
C
{0001}
{11-20}
{10-10}
D
{0001}
{11-20}
{10-10}
E
{0001}
{11-20}
{10-10}

Fig. 13

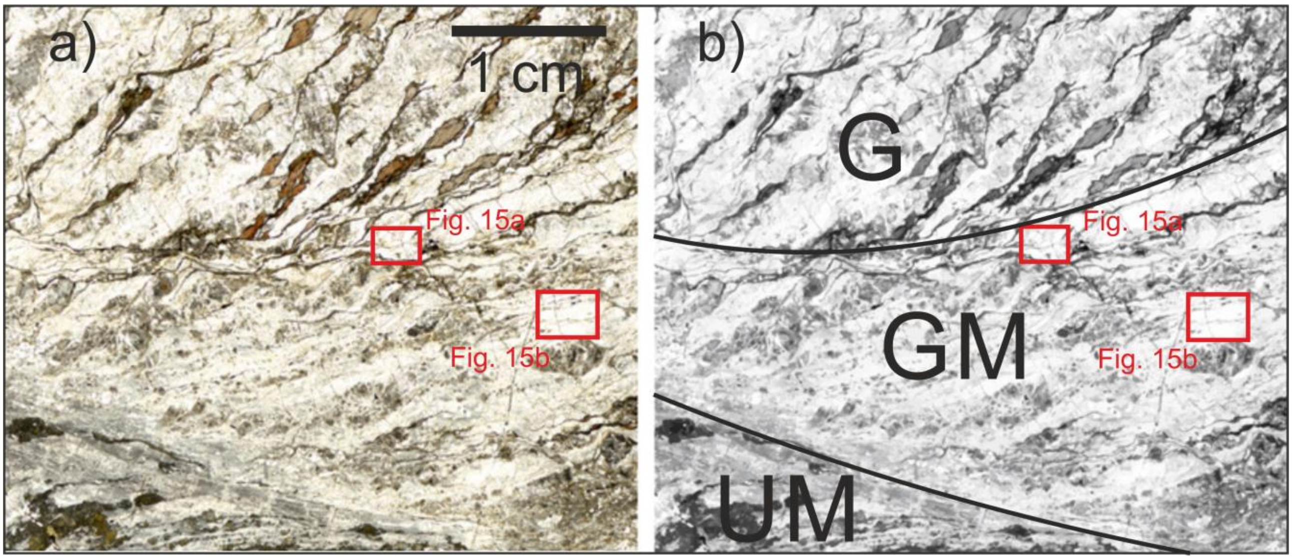


Fig. 14

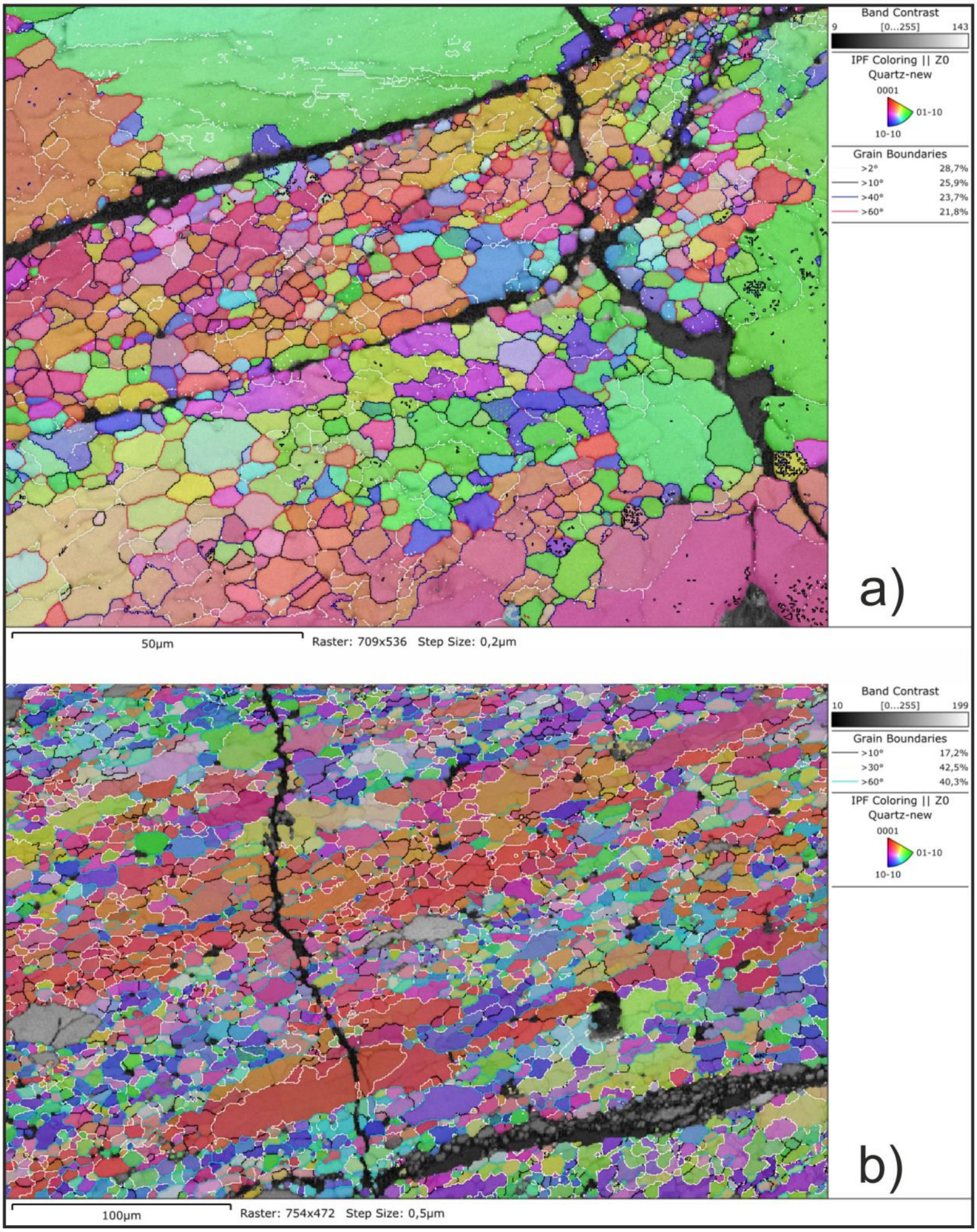

Band Contrast
9 [0...255] 143
IPF Coloring || Z0
Quartz-new
0001
01-10
10-10
Grain Boundaries
>2° 28,7%
>10° 25,9%
>40° 23,7%
>60° 21,8%
a)
50µm
Raster: 709x536 Step Size: 0,2µm
Band Contrast
10 [0...255] 199
Grain Boundaries
>10° 17,2%
>30° 42,5%
>60° 40,3%
IPF Coloring || Z0
Quartz-new
0001
01-10
10-10
b)
100µm
Raster: 754x472 Step Size: 0,5µm

Fig. 15

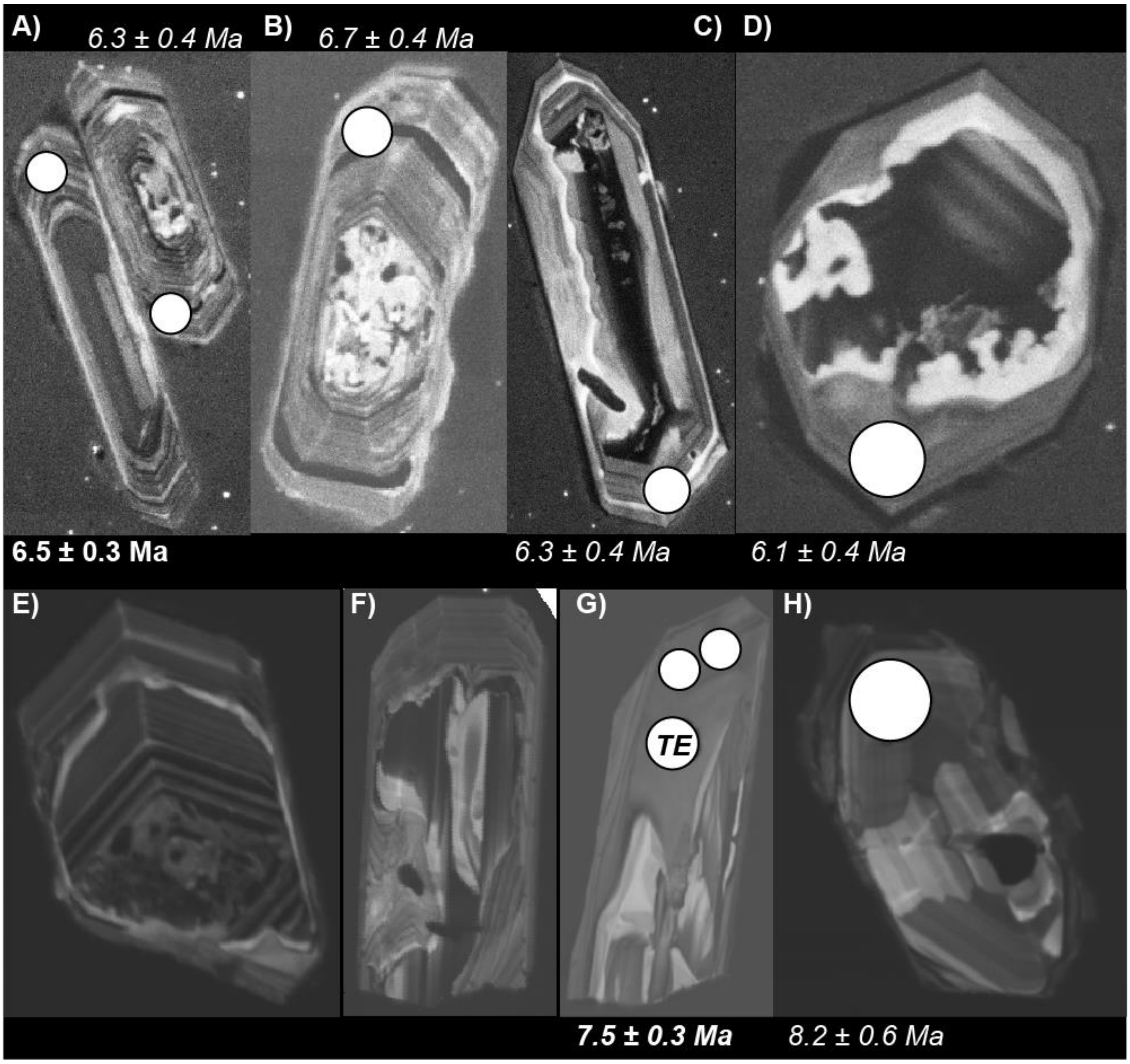

A)
6.3 ± 0.4 Ma
B)
6.7 ± 0.4 Ma
C)
D)
6.5 ± 0.3 Ma
6.3 ± 0.4 Ma
6.1 ± 0.4 Ma
E)
F)
G)
H)
TE
7.5 ± 0.3 Ma
8.2 ± 0.6 Ma

Fig. 16

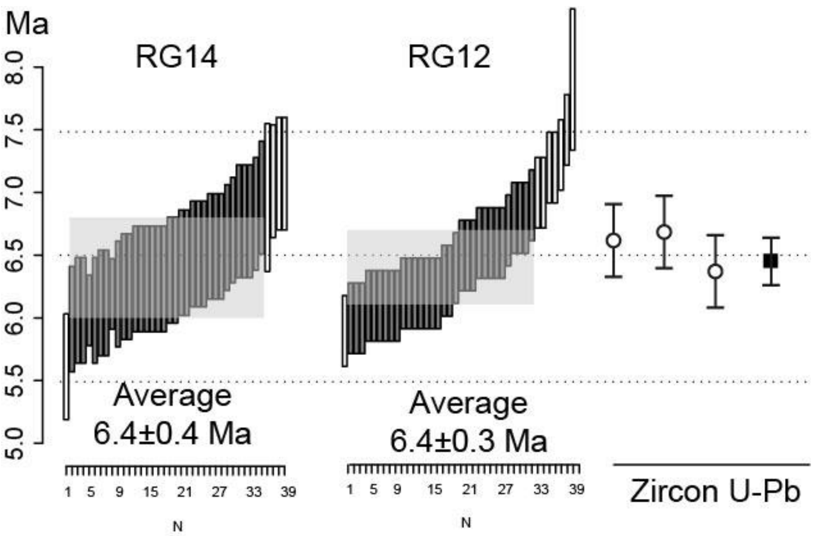
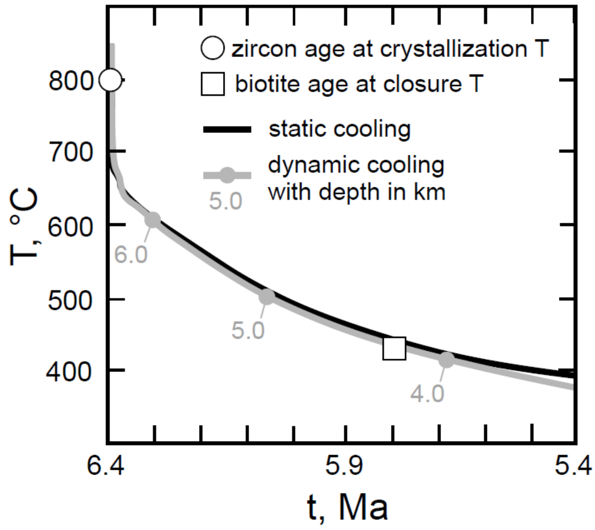

Ma
8.0
7.5
7.0
6.5
6.0
5.5
5.0
RG14
RG12
Average
6.4±0.4 Ma
Average
6.4±0.3 Ma
1 5 9 15 21 27 33 39
N
1 5 9 15 21 27 33 39
N
Zircon U-Pb

Fig. 17



zircon age at crystallization T
biotite age at closure T
static cooling
dynamic cooling
with depth in km
5.0
6.0
5.0
4.0
800
700
600
500
400
T, °C
6.4
5.9
5.4
t, Ma

Fig. 18

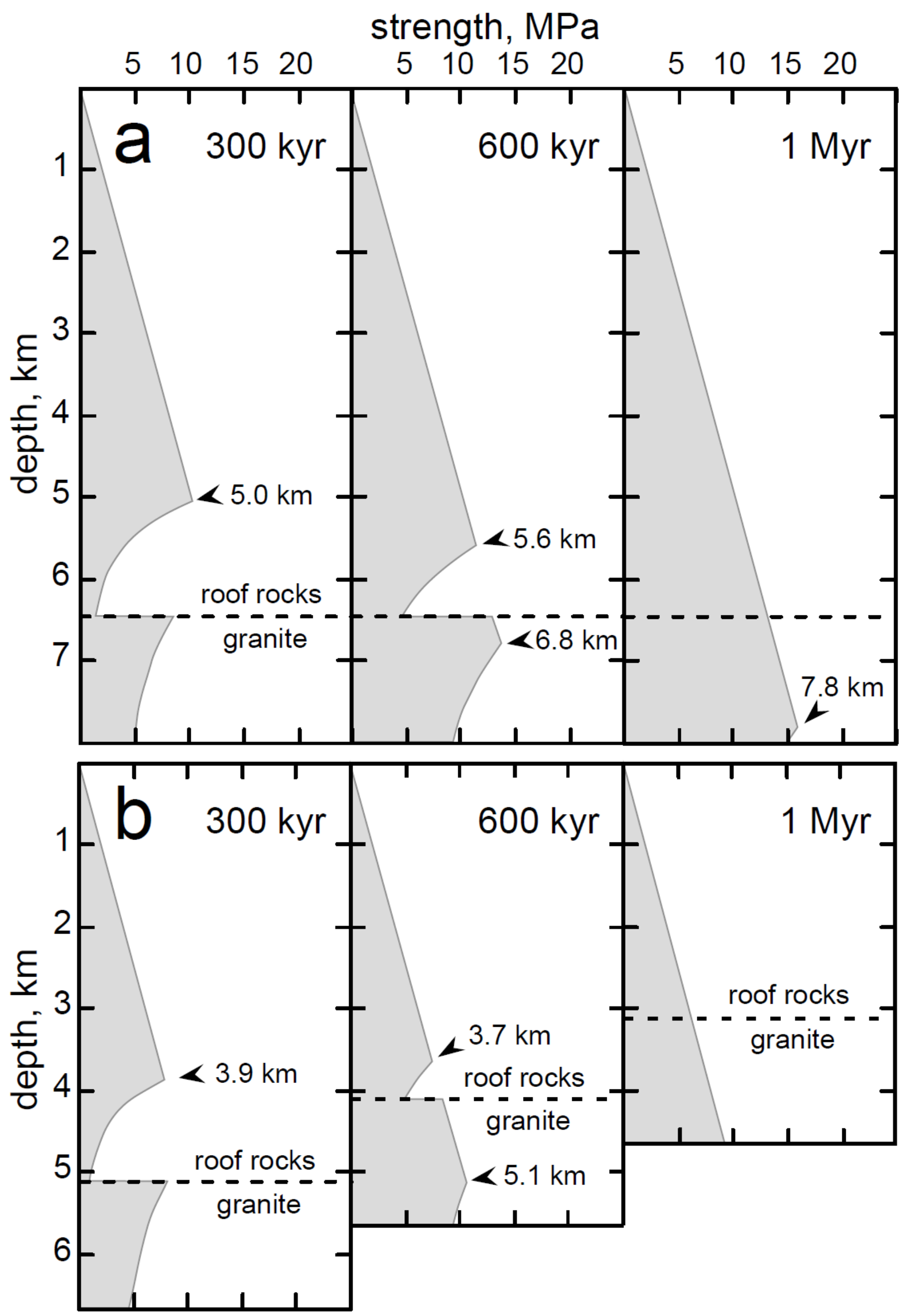

strength, MPa
5
10
15
20
depth, km
1
2
3
4
5
6
7
a
300 kyr
600 kyr
1 Myr
5.0 km
5.6 km
6.8 km
7.8 km
roof rocks
granite
b
300 kyr
600 kyr
1 Myr
3.9 km
3.7 km
5.1 km
roof rocks
granite
roof rocks
granite
roof rocks
granite

Fig. 19

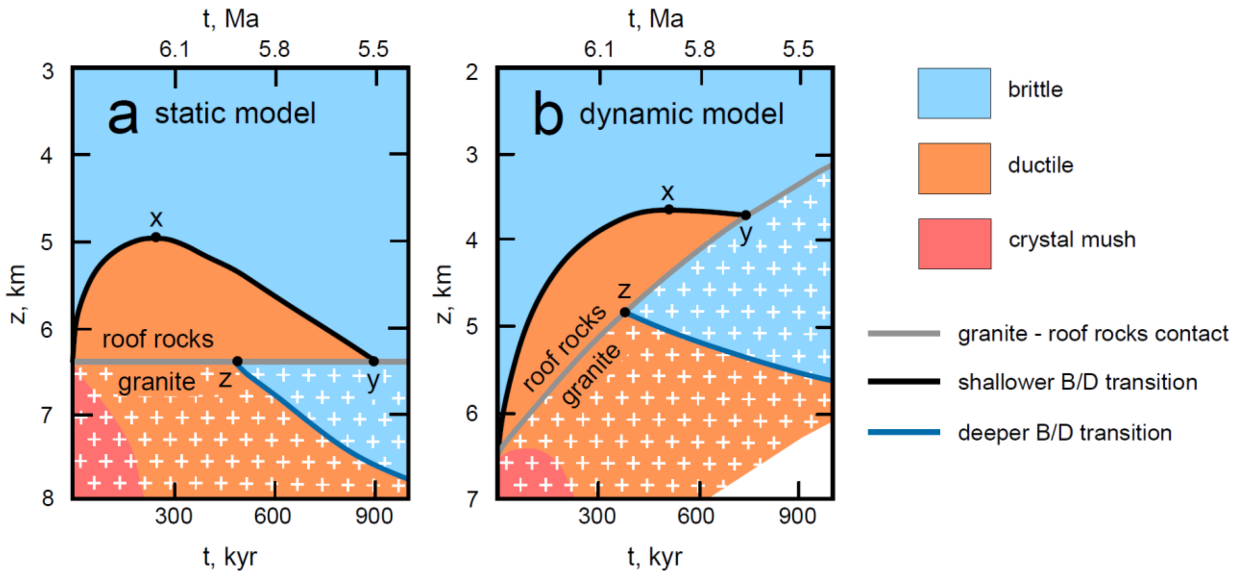


Fig. 20

Parameter	Symbols and equations	Values
Thermal conductivity [W m^{-1} K^{-1}]	K	1.85 (crust) 3.35 (mantle)
Density [kg m^{-3}]	ρ	2750 (crust) 3300 (mantle)
Specific heat [J kg^{-1} K^{-1}]	C_P	1000 (crust) 1100 (mantle)
Heat generation rate [µW m^{-3}]	$A = A_0\, e^{(-z/D)}$	A_0 = 2 D = 12000 m
Latent heat of crystallization [kJ kg^{-1}]	ΔH	300 (for a T range of 850-650 °C)
Unroofing rate [m yr^{-1}]	$v_z = dz/dt = -c\, z$	$v_{z\,(t=0)} = 5 \times 10^{-3}$ m yr^{-1} $c = 8.93 \times 10^{-7}$ yr^{-1}

Tab. 1

	Lithology	Rheological analog	Creep parameters		
			A_c [$MPa^{-n}s^{-1}$]	n	E [J mol^{-1}]
Roof rock	micaschist and quartzite	wet quartzite	3.2×10^{-4}	2.3	1.54×10^{5}
Pluton	monzogranite	granite	1.80×10^{-9}	3.2	1.23×10^{5}

Tab. 2