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The role of trapped fluids during the development and deformation of carbonate/shale intra-wedge tectonic mélange

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Abstract: In contrast to the numerous studies on exhumed tectonic mélanges along subduction channels, in accretionary wedge interiors, deformation mechanisms and related fluid circulation in tectonic mélanges are still underexplored. Here we combine structural and microstructural observations with geochemical (stable and clumped isotopes and isotope composition of noble gases in fluid inclusions of calcite veins) and geochronological data (U-Pb dating) to define deformation mechanisms and syn-tectonic fluid circulation within the Mt. Massico intra-wedge tectonic mélange, located in the inner part of the central-southern Apennines accretionary wedge, Italy. This mélange developed at the base of a clastic succession, and shear deformation was characterized by disruption of the primary bedding, mixing, and deformation of relicts of competent olistoliths and strata within a weak matrix of deformed clayey and marly interbeds. Recurrent cycles of mutually overprinting fracturing/veining and pressure-solution processes generated a block-in-matrix texture. The geochemical signatures of syn-tectonic calcite veins suggest calcite precipitation in a closed system from warm (108-147 °C) paleofluids, with $\delta^{18}O$ composition between +9‰ and 14‰, such as trapped pore water from diagenesis after extensive ^{18}O exchange with the local limestone host rock ($\delta^{18}O$ values between +26‰ and +30‰) and/or derived by clay dehydration processes (at T > 120 °C). The $^{3}He/^{4}He$ ratios in fluid inclusions trapped in calcite veins are lower than 0.1 Ra, hence He was exclusively sourced from the crust, excluding mantle-derived fluids. We conclude that: (1) intraformational rheological contrasts, inherited trapped fluids, and low-permeability barriers such as clay-rich stylolites, can favour the development of fluid overpressure and the generation of intra-wedge tectonic mélanges; (2) clay-rich intra-wedge tectonic mélanges may generate efficient barriers within accretionary wedges for vertical and lateral redistribution of fluids from reservoirs outside the mélange. We highlight that the integration of geochemical and geochronological methods can be a powerful approach to better constrain, in the future, the burial-thermal evolution of fold and thrust belts and sedimentary basins.

Keywords: tectonic mélange; fold and thrust belt; fault-fluid interaction; stable and clumped isotopes; noble gases; U-Pb dating

1. Introduction

Décollements and thrust faults are major structures that control the internal architecture of fold-and-thrust belts (e.g. Morley et al., 2017). In particular, décollements commonly generate along evaporite- or clay-rich formations, control the style and timing of folding and tectonic imbrication, and may act as barriers or conduits for crustal flow of fluids, including hydrocarbons. Décollements can often act as channelized paths for fluid redistribution within the crust (e.g. Vrolijk et al., 1988; Vannucchi et al., 2008).

The shearing of clay-rich and initially layered formations along décollements can generate tectonic mélanges through primary bedding disruption, pressure-solution, veining, and offscraping/mixing of blocks from competent formations located above and below the décollement horizon (see Festa et al., 2012 for review). Such deformations can lead to block-in-matrix fabrics, consisting of competent blocks scattered in a weak matrix, which are typically characterized by heterogeneous mechanical and permeability properties (e.g. Fagereng, 2011). Outcrop, microstructural, and geochemical characterizations of tectonic mélanges can contribute to constrain their spatio-temporal evolution as well as the mechanical properties and hydrogeological characteristics (paleofluid flow).

In this context, several studies focused on tectonic mélanges exhumed from basal décollements at the subduction interface along the toe of accretionary wedges and from subduction channels (e.g. Meneghini et al., 2009; Vannucchi et al., 2008; see Festa et al., 2012 for review), where deformation can reach metamorphic conditions (e.g. Fagereng and Cooper, 2010). Only few studies focused on intra-wedge tectonic mélanges developed at diagenetic P-T conditions along décollements at the base of thrust sheets (e.g. Vannucchi and Bettelli, 2002; Codegone et al., 2012; Dewever et al., 2013; Ogata et al., 2012; Smeraglia et al., 2019). However, intra-wedge tectonic mélanges may strongly affect the mechanical behavior of inner décollements, thus influencing accretionary wedge geometry and kinematics, and can modulate the transport of fluids (i.e. water, hydrocarbon) across fold-and-thrust belts. Therefore, the understanding of mechanical and hydrological properties of exhumed mélanges is fundamental to unravel fluid (paleo)pathways and asses potential reservoirs for geofluid (i.e. hydrocarbon) accumulation in fold-and-thrust belts.

For these reasons, here we combine outcrop and microstructural observations with geochemical analyses (stable and clumped isotopes and isotope composition of noble gases in fluid inclusions of calcite veins) to unravel the deformation mechanisms, paleohydrology (i.e., fluid conduit/barrier behavior), and the temperature conditions during fluid flow within a 150 m-thick intra-wedge tectonic mélange in the Mt. Massico Ridge (e.g. Billi et al., 1997; Vitale et al., 2018; Smeraglia et al., 2019), which is located in the inner sector (Tyrrhenian side) of the central-southern Apennines fold-and-thrust belt. In this area, outcrop conditions allowed the reconstruction of syn-tectonic paleofluid circulation and the unraveling of progressive intra-wedge mélange deformation, from the undeformed host rock to the finite fabric development by tectonic deformation.

2. Geological background

2.1 Central-southern Apennines

The central-southern Apennines are a late Oligocene to present fold-and-thrust belt generated by the eastward rollback of the west-dipping subduction of the Adriatic plate below the European continental margin (Fig. 1a; e.g., Carminati et al., 2012). Orogenic accretion was accomplished through and accommodated by a set of NE-verging thrust systems, which scraped off pre- and syn-orogenic deposits of the Adriatic plate (Fig. 1b). The pre-orogenic deposits consist of ~4,000-5,000 m-thick Upper Triassic-Middle Miocene carbonate platform deposits (i.e. Apennine platform; e.g. Vezzani et al., 2010; Vitale and Ciarcia, 2018) and of ~100 m-thick middle Miocene hemipelagic marls deposited in a transitional foreland-to-foredeep environment (e.g., Vezzani et al., 2010). The syn-orogenic deposits consist of up to ~3,100 m-thick upper Miocene siliciclastic sandstones, marls, and claystones deposited in a foredeep environment (e.g., Vitale and Ciarcia, 2018). Thrust faults developed during wedge accretion and juxtaposed pre-orogenic deposits onto syn-orogenic sediments, generating stacks of imbricate thrust sheets from the surface down to depths of ~10 km (Fig. 1a,b; e.g. Mostardini and Merlini, 1986; Vezzani et al., 2010). Since early Pliocene time, the internal and axial part of the central-southern Apennines belt underwent tectonic uplift and regional extension, related to the opening of the Tyrrhenian backarc basin and accommodated by the development of NW-SE-oriented extensional faults (Fig. 1a,b; Malinverno and Ryan, 1986).

2.2 The Mt. Massico ridge

Mt. Massico is a NNE-SSW-trending ridge located in the innermost part of the central-southern Apennines (Fig. 1a; e.g. Billi et al., 1997; Luiso et al., 2018; Vitale et al., 2018). It is bounded by NE-striking active extensional faults and surrounded by the Quaternary Garigliano and Volturno river plains, the Roccamonfina volcano (Quaternary), and the Tyrrhenian Sea basin (Figs. 1a and 2a).

In the central and northeastern parts of the Mt. Massico ridge, Upper Triassic to Upper Cretaceous carbonates are exposed (Fig. 1a; Vitale et al., 2018) and arranged to form a ENE-verging asymmetric anticline with an upright forelimb and a WSW-dipping (40°-50°) backlimb (Fig. 1a). In the forelimb, a thrust juxtaposes Upper Cretaceous limestones onto Tortonian syn-orogenic sandstones (see Figure 2 in Sgrosso, 1974). The WSW-dipping backlimb continues in the southwestern part of the ridge, where middle-upper Miocene sediments are exposed (Fig. 2a,b).

This succession lies paraconformably atop Upper Cretaceous limestones and begins from the bottom with ~50 m-thick bryozoan-rich limestones of the Cusano Fm., followed upward by hemipelagic marls of the Longano Fm., with a preserved thickness of less than 1-3 meters (Sgrosso, 1974; Vitale et al., 2018) (Fig. 2a). Both these formations were deposited in a foreland environment and are Serravallian in age (Sgrosso, 1974). Above the hemipelagic marls, the sedimentary succession evolves into the Tortonian-lower Messinian siliciclastic/carbonaticlastic deposits of the

Caiazzo Fm. (hereafter called clastic deposits, Fig. 2a,b), which are interpreted as wedge-top basin deposits (Vitale et al., 2018). The clastic deposits are made up of: (1) massive-to-bedded siliciclastic sandstones with clayey, marly, and calcarenite interbeds and (2) polymictic breccias/conglomerates characterized by Mesozoic-Cenozoic carbonate blocks and siliciclastic clasts (up to a few dm^3 in volume), including also clasts of volcanic, intrusive, and metamorphic rocks (Fig. 2a) (Sgrosso, 1974). The clastic deposits are unconformably overlain by well-bedded lower Messinian calcarenites (Fig. 2a). Intercalated within the clastic deposits are hundreds of olistoliths, from a few dm^3 to several thousands of m^3 in volume. Olistoliths consist of Mesozoic-Cenozoic limestones and exotic rocks not exposed in the local sedimentary succession (hereafter named exotic olistoliths), such as marbles and recrystallized carbonates as well as volcanic, intrusive, and metamorphic rocks, and deep basinal Paleocene limestone blocks (Sgrosso, 1974; Di Girolamo et al., 2000; Vitale et al., 2018). The total thickness of the upper Tortonian-lower Messinian succession is ~800 m (Smeraglia et al., 2019).

2.3 The Mt. Massico mélange

The about ~150 m thick basal portion of the clastic deposits and the underlying hemipelagic marls are strongly deformed, showing a block-in-matrix fabric typical of tectonic mélanges (Fig. 2a,b; Vitale et al., 2018; Smeraglia et al., 2019). Based on new geological mapping (see also Vitale et al., 2018 for new stratigraphic constraints), structural analysis, U-Pb dating (syn-tectonic calcite ages of 10.5 ± 2.5 My, 7.0 ± 1.6 My, and 5.1 ± 3.7), and thermal modelling constrained by mixed layers illite-smectite, Smeraglia et al. 2019 proposed that the Mt. Massico tectonic mélange developed, since late Messinian times, by out-of sequence thrusting of multiple thrust sheets above the clastic deposits located in the backlimb of the Mt. Massico anticline (Fig. 9a-c) and by localization of deformation at the base of clastic deposits. In this context, the original stratigraphic boundary between the clastic deposits/hemipelagic marls deposits and the underlying limestones was sheared as a décollement horizon and an intra-wedge tectonic mélange developed due to the strong rheological contrast between the weak clastic deposits and the competent limestones. This mélange is characterized by a pervasive S-C and S-CC’ fabrics showing a coherent transport direction towards ESE, accommodated by partitioned reverse and dextral transpressive tectonics (Fig. 2a; Smeraglia et al., 2019).

3. Methods

Aiming at characterize the structurally-controlled fluid circulation and the geochemical signature of the syn-tectonic fluids in the transition from the undeformed host rock to the tectonic mélange formation, we focused our structural and geochemical studies within the San Sebastiano and Mt. Cicoli areas (Fig. 2). In these areas, it is in fact possible to observe the progressive transition from poorly deformed clastic deposits to moderately and strongly deformed parts of the tectonic mélange along the boundary with the underlying bryozoan-rich limestones of the Cusano Fm. (Fig. 2a). To achieve this purpose we integrated: (1) geological field mapping at 1:10,000 scale, and available maps (Billi et al., 1997; Vitale et al., 2018; Smeraglia et al., 2019), to unravel the structural architecture and the deformation history of the intra-wedge tectonic mélange; (2) microstructural (optical microscope) analyses to unravel the deformation mechanisms acting during mélange generation; (3) Stable carbon and oxygen and carbonate clumped isotopes to reconstruct the origin and precipitation temperatures of the paleofluid circulating during mélange generation; (4) H_2O+CO_2, N_2, and minor gaseous species (He, Ne, and Ar concentrations and isotope ratios) in fluid inclusions on previously U-Pb dated syn-tectonic carbonate samples/veins (ages of 10.5 ± 2.5 My, 7.0 ± 1.6 My, and 5.1 ± 3.7; Smeraglia et al., 2019) . Sample preparation and analytical details are described in the Supplementary Material.

4. Results

4.1 Outcrop observations

A progressive transition in shear strain localization is observed when moving across strike along the studied area. Within the poorly deformed clastic deposits, bedding is still observable (Fig. 3a,b). The competent sandstones beds and the marly-shaly interbeds are gently folded and/or cut by low-displacement reverse faults displaying a ramp-and-flat geometry, with the flat segment located within marly and shaly interbeds (Fig. 3b). The marly and shaly interbeds are characterized by a weak and bedding-parallel scaly foliation (Fig. 3a,b); however, in places, an incipient S-C fabric characterized by bedding-parallel S planes and bedding-oblique C planes occurs dismembering the sandstone interbeds (Fig. 3a).

We recognized a up ~100 m-thick zone, referred as moderately deformed zone of the tectonic mélange, exposed within a diffuse boundary between the poorly deformed clastic deposits and the strongly deformed part of the tectonic mélange (Fig. 2). This zone is characterized by still detectable primary structures such as bedding, but competent strata (i.e. sandstones and calcarenites interbeds) show pinch-and-swell and boudin-like geometries. Pervasive and anastomosing bedding-parallel scaly foliation occurs within marly and shaly interbeds (Fig. 3c,d) and, in places, S-C fabrics can be observed (Fig. 3e). With increasing deformation, primary foliation is dismembered and relicts of bedding are scattered within the scaly foliation generating a block-in-matrix fabric (Fig. 3d-e). In particular, such relicts show bedding-perpendicular calcite veins (Fig. 3d,e).

The strongly deformed part of the tectonic mélange, up to ~50 m-thick, shows well-developed S-C and S-CC' tectonites and sedimentary structures are completely obliterated by tectonic fabric (Figs. 3f-h and 4a,b). In particular, the WNW- to WSW-dipping S- and C-surfaces show dip angles ranging ~60°-70° and ~10°-30°, respectively, with kinematic indicators showing mainly reverse dip-slip to right-lateral transpressional movements (Fig. 2a). Marls and shales are affected by S-C foliations and clasts of competent rocks (i.e. sandstones and calcarenites) are deformed in sigmoidal-lozenge-shaped structures consistent with the overall sense of shear (Figs. 3f-g and 4a-c). Stylolites and slickenfibers are aligned parallel respect to S and C planes, respectively, whereas sigmoidal clasts of competent rocks are permeated by calcite veins (Figs. 3g,h and 4d).

With increasing deformation, the sigmoidal relicts of sedimentary structures are progressively thinned and completely obliterated by S-C foliation so that it is difficult to distinguish them from the pervasive scaly foliation at the naked eye (Fig. 4c). In places, foliated marls and shales wrap olistoliths inherited from clastic deposits. The olistoliths preserve their original (from irregular to rounded) shapes, generating a block-in-matrix fabric (Fig. 4e-g). Convolute/contorted foliations also occur (Fig. 4h). Their origin is not clear and it cannot be excluded that they may be relicts of soft sediment (fluid escape) structures. When observable, the boundary between the tectonic mélange and the underlying bryozoan-rich limestones (not affected by deformation and with sedimentary structures and fossils still recognizable) is sharp.

4.2 Microstructural observations

We focused our microstructural observations on the strongly deformed parts of the tectonic mélange, with special focus on the sigmoidal-lozenge clasts of competent rocks (i.e. sandstones and calcarenites), since they are cohesive and can be easily cut for thin section preparation.

The sigmoidal-lozenge clasts are characterized by pervasive stylolites, parallel to S, C, and C' planes (Fig. 5a-c). Stylolites evolve from rough, teeth-shaped, and indented morphology (Figs. 5a-c and 6a-b) to smooth, continuous, and thick (up to 1 mm) dissolution seams (Figs. 5a-c and 6c) filled by insoluble material (i.e. clay minerals and oxides), small clasts of host rock limestones, and fragments of calcite veins (Fig. 6c). Stylolites and dissolution seams bound sigmoids defining a fabric sub-parallel to S, C, and C' planes (Figs. 5a-c and 6a,b). Sigmoids consist of host rock commonly with deformed microfossils, and/or reworked calcite veins (Fig. 6a,b).

We identify fibrous and blocky calcite veins. Fibrous veins are oriented perpendicular to stylolites ad are characterized by fibrous crystals perpendicular to vein margins (Fig. 6 d-f). Crystals are deformed by stylolites (Figs. 6d-f and 7a). The slickenfibers observed at the outcrop scale along S, C, and C' planes are characterized, at the microscale, by fibrous veins located along stylolite jogs, characterized by fibrous crystals parallel to stylolites (Fig. 6b,f). In places, fibrous veins occur within thick dissolution seams (Fig. 7a,b). In this case, vein margins and fibrous crystals are roughly parallel and perpendicular to dissolution seam margins, respectively (Fig. 7a,b). Blocky veins are filled by blocky to elongated-blocky calcite crystals and are oriented parallel, perpendicular, or oblique to stylolites (Fig. 7c,d).

Concerning the relative timing of structures, we observed mutual crosscutting relationships between veins and stylolites (Figs. 5c and 7f). In particular, blocky veins perpendicular to stylolites cut or are deformed by stylolites and/or by blocky/fibrous veins oriented oblique to stylolites (Figs. 5 and 7c-f). Fibrous veins parallel and/or perpendicular to stylolites are deformed by stylolites and/or by blocky veins oriented oblique to stylolites (Figs. 5, 7a,b, and 6d-f)

4.3 Stable and clumped isotopes

Results from stable and clumped isotopes analyses are shown in Fig. 8 and listed in Table S1 and S2. Results are reported in the conventional δ notation with respect to the Vienna Pee Dee Belemnite (VPDB) for $\delta^{13}C$ and Vienna Standard Mean Ocean Water (VSMOW) for $\delta^{18}O$. Clumped isotopes are reported in the Carbon Dioxide Equilibrium Scale (Dennis et al. 2011).

The host rock $\delta^{13}C$ and $\delta^{18}O$ values, measured on limestones interbeds within clastic deposits, range between -0.5‰ and +1.5‰, and between +26‰ and +29.5‰, respectively. Such values are typical of Miocene marine carbonates in the Apennines, Italy (e.g., Hilgen et al., 2005).

Blocky veins $\delta^{13}C$ and $\delta^{18}O$ values range between 0‰ and +0.8‰ and between +22.3‰ and +27‰, respectively, with one blocky vein showing $\delta^{13}C$ and $\delta^{18}O$ values overlapping those of the host rock. Fibrous veins $\delta^{13}C$ and $\delta^{18}O$ values range between -0.4‰ and +1.6‰ and between +22.4‰ and +27.8‰, respectively. Three fibrous veins show $\delta^{13}C$ and $\delta^{18}O$ values overlapping those of the host rock. Overall, both blocky and fibrous veins show $\delta^{13}C$ values overlapping with those of the host rock and an average $\delta^{18}O$ depletion of ~4‰ respect to the host rock.

Clumped-isotope data from blocky veins (four samples) and fibrous veins (five samples) yields $\Delta47$ values between 0.424 and 0.482 (Table S2). These values correspond to temperatures between 108 ± 13 °C and 147 ± 20 °C (Fig. 8b and Table S1), using the equation of Bernasconi et al. (2018). The calculated $\delta^{18}O$ paleofluid compositions, using the O'Neil et al. (1969) equation developed for calcite precipitation temperature in the 0-500 °C range, range between 9.1‰ and 13.7‰ (Fig. 8b and Table S1).

4.4 Noble gas analysis

The concentrations of CO_2+H_2O, N_2, light noble gases (He, Ne, Ar), and $^3He/^4He$, $^4He/^{20}Ne$, $^{40}Ar/^{36}Ar$ isotopic ratios in fluid inclusions hosted in syntectonic calcite veins (samples 239, 257 and 260) are reported in Table S3.

CO_2+H_2O and N_2 show concentrations ranging between $1.1x10^{-6}$ and $8.3x10^{-6}$ mol/g and between $3.1x10^{-7}$ and $6.7x10^{-7}$ mol/g, respectively. The N_2/Ar ratios range between 3092 and 3633. These values are much more higher than the N_2/Ar ratio in the atmosphere (84.1) and the N_2/Ar ratio in air-saturated water at standard temperature and pressure (38). The $^{40}Ar/^{36}Ar$ ratios vary from 367.3 to 389.9, slightly higher than the $^{40}Ar/^{36}Ar$ ratio in atmosphere (298.6).

He/Ar and He/N_2 ratios range between 0.06-0.14 and $2.1\text{-}3.8x10^{-5}$, respectively, and are 1-2 orders of magnitude higher than the theoretical values in the atmosphere and in the air-saturated water. The $^4He/^{20}Ne$ ratios range between 51.7 and 96.5 and are more than two orders of magnitude higher than the typical $^4He/^{20}Ne$ ratio in air saturated water (0.268), consistently with fluids trapped at great depth. All these ratios testify an excess of He and N_2 respect to the typical concentrations in atmosphere-derived fluids. The $^3He/^4He$ ratios, corrected for the air contamination (R/Ra ratio) range between 0.05 and 0.09, consistently with those of crustal fluids, thus excluding a contribution of He derived from atmospheric air and ^3He from the mantle (Fig. 8d).

The $^3He/^4He$ and $^{40}Ar/^{36}Ar$ ratios are useful tracers to ascertain the origin of the fluids producing the mineralization. However, one condition is that these materials must contain volatiles trapped during precipitation processes and that their isotopic compositions have not been modified over time. Significant post precipitation processes that may affect the noble gases content in the fluid inclusions regard the addition of (1) ^{4}He, ^{40}Ar produced from the radiogenic decay of U, Th and K, and (2) cosmogenic ^{3}He derived from the exposure to cosmic ray.

In order to estimate the contribution of radiogenic ^{40}Ar into the fluid inclusions, we computed the ^{40}Ar*, which is the amount of ^{40}Ar corrected for the atmospheric contributions (^{40}Ar* = ^{40}Ar/^{36}Ar$_{Measured}$ - ^{40}Ar/^{36}Ar$_{Atmosphere}$ x ^{36}Ar$_{Measured}$). The amounts of ^{40}Ar* range between 8.99 x 10^{-11} to 1.38 x 10^{-10} mol/g (Table S3). The amounts of the measured ^{4}He and measured ^{40}Ar* do not show any correlations with the ages of the veins (Fig. S2a,b), thus excluding the accumulation of radiogenic He and Ar produced in the veins by the U, Th and K decay during time.

All calcite veins have been collected in outcrops whose time of exposure to the cosmic ray at the surface is not evaluable; hence, the potential contribution of cosmogenic ^{3}He trapped in the fluid inclusions is difficult to be assessed. However, there is no correlation between the amount of ^{3}He and the ages of calcite veins (Fig. S2c), which can supports a progressive accumulation over time of cosmogenic ^{3}He and subsequent migration into the fluid inclusions.

On the basis of the average U and Th amounts in the veins of calcite (1.58 and 2.28 ppm for sample 257; 0.35 and 0.01 ppm for sample 239; 0.17 and 0.02 ppm for sample 260; Table S4) is possible to compute the potential ^{4}He/^{40}Ar* ratio produced by the decay of U and Th. The computed ^{4}He/^{40}Ar* ratios range between 10 and 100. These values are much higher than the same ratios in the fluid inclusions (0.61, 0.35, and 0.40 for samples 257, 239, and 260, respectively, Table S3). This suggests that the fluid inclusions are not modified by the contributions of both of radiogenic ^{4}He and ^{40}Ar produced within the rocks. We conclude that the isotopic ratios of He and Ar in the fluid inclusions are representative of the pristine isotopic signatures of the trapped fluids.

5. Discussion

5.1 Mélange evolution and deformation mechanisms

Since late Messinian times, the Mt. Massico tectonic mélange developed by out-of sequence thrusting of multiple thrust sheets above the clastic deposits and by localization of deformation at the base of clastic deposits (Fig. 9a-c; Smeraglia et al., 2019). Therefore, when out-of sequence deformation initiated, WSW-ENE shortening affected W-SW-dipping (30°-55°) Meso-Cenozoic carbonates and middle-upper Miocene clastic deposits, tilted by previous thrust-related folding (Fig. 2a,b and 9b; Smeraglia et al., 2019). In this context, both layer-perpendicular/oblique shortening and layer-parallel extension occurred in the early phases of deformation (Fig. 10a,b). Layer-perpendicular/oblique shortening was driven by sub-horizontal tectonic compression and produced bedding-parallel foliation by pressure-solution and stylolites generation in the marly and shaly interbeds (Figs. 3a,b). Layer-parallel extension produced dismembering, boudinage, and fracturing of competent bedding (i.e. sandstones and calcarenites interbeds; Fig. 3c).

With increasing deformation, the hemipelagic marls between the clastic deposits and bryozoan-rich limestones were scraped off and mixed within the base of clastic deposits, the competent beds were almost completely dismembered, although some relicts of bedding were scattered within the shaly/marly matrix (Figs. 3d, 4e-g, and 10c,d). In this framework, S-C tectonites and block-in-matrix fabric developed within the strongly deformed parts of the tectonic mélange by brittle-ductile processes such as pressure-solution, fracturing/veining, and frictional sliding, with absence of evident cataclasis of the host rock (Fig. 10c,d). The absence of cataclasis is related to the occurrence of widespread pressure solution processes, triggered and enhanced by clay-rich host rocks (marls and shales; Renard et al., 1997), which promoted the dissipation of tectonic stress through host rock dissolution, at slow strain rates rather, than frictional processes (i.e. fracturing, grain rotation, abrasion) commonly occurring during cataclasis in competent rocks, such as pure carbonates (e.g. Billi, 2010). In particular, pressure-solution was more pervasive in weak marly and shaly interbeds than in calcarenite interbeds, producing a network of stylolites aligned along S, C, and C' planes. Stylolites developed through host rock dissolution and insoluble materials (i.e. mainly phyllosilicates) concentrated within dissolution seams (Figs. 3f-h and 4a-c) (e.g. Tesei et al., 2013). Fracturing/veining processes, coupled with pressure-solution, affected the relicts of competent rocks (Figs. 3f-h, 4a-d, and 10c,d). These types of deformation mechanisms commonly occur during tectonic mélange generation (e.g., Festa et al., 2012 for review).

5.2 Vein development within the mélange

The occurrence of veins perpendicular to stylolites suggests that both structures formed under the same stress regime, characterized by compression sub-perpendicular to the stylolites, thus generating mode I vein opening perpendicular to stylolites to regulate their evolution through time (Figs. 5a-c, 6d, 7c-f, and 11; e.g., Gratier et al., 2013). Veins parallel to stylolites (Figs. 6f and 7a,b) indicate opening direction parallel to the maximum compression direction and suggests vein opening under an unfavorable stress regime (i.e. opening direction should be perpendicular with respect to the maximum principal stress). In particular, development of fibrous veins (shear veins/slickenfibers) characterized by fibrous crystals perpendicular to the maximum principal stress have been observed in tectonic mélange and have been related to shear along pre-existing weak planes such as clay-rich stylolites, assisted by fluid overpressure (Fagereng et al., 2010). Alternatively, the unfavorable orientations between veins and tectonic stresses can be explained by episodic stress rotation within the mélange, vein reworking/rotation during deformation, and/or overpressured fluids that may have opened pre-existing discontinuities, such as stylolites. Although stylolites are commonly considered efficient barriers to fluid flow (Toussaint et al., 2018 for review), veins developed along stylolites (Fig. 7a,b) suggest that they can be preferential pathways for fluid redistribution, consistently with recent experimental and field evidence (e.g., Heap et al., 2014; Bruna et al., 2019). In particular, the low-permeability network generated by stylolites can favor fluid overpressure rise.

Stylolite steps filled by calcite crystals (Fig. 6f) suggest opening of voids (i.e. extensional pull-apart) due to slip along undulated stylolites, which promoted calcite precipitation into newly created space, a mechanism commonly occurring during slickenfiber generation (e.g. Fagereng and Byrnes, 2015). This mechanism indicates that frictional sliding occurred along S-, C-, and C'-planes by smoothing the tooth-shaped margins of stylolites and slip along thick clay-rich dissolution seams (Fig. 10d; Tesei et al., 2013).

Fibrous crystals (Figs. 6d-f and 7a,b) generated through multiple and micrometer-thick opening increments (e.g., Bons et al., 2012), is incompatible with coeval growth of blocky and/or elongated-blocky crystals that typically develop in fluid-filled spaces generated by a single opening increment larger than that of fibrous veins (e.g. Bons et al., 2012). These different deformation mechanisms suggest various slip rates and deformation behaviors within the mélange. Continuous deformation may have occurred within the ductile clay-rich matrix, while discontinuous slip may have occurred during fibrous and blocky vein generation in more competent carbonate-rich blocks (e.g. Fagereng, 2011). In particular, fibrous veins suggest discontinuous creep at very slow slip rates. On the contrary, impulsive deformation at high and fast slip rates may have occurred during blocky vein generation generated by impulsive crackle-like brecciation (e.g. Woodcock et al., 2014; Fagereng and Byrnes, 2015). In both cases, fracturing/veining were probably assisted by fluid overpressure, consistently with crackle-like brecciation (Fig. 5) and the dense network of clay-rich stylolites, which may have created impermeable barriers and fluid pressure rises over time (Moore and Vrolijk, 1992).

Crosscutting relationships between structures show recurrent cycles of mutually overprinting brittle (fracturing/veining and frictional sliding) and ductile (pressure-solution) processes indicate a multiphase deformation history (Figs. 5, 6, 7, and 11). Stylolite formation through fibrous veins (Fig. 6d,e) indicate the interruption of the fracturing event and the onset of a new pressure-solution phase, indicating alternating phases of frictional sliding and dissolution processes (Fig. S1; e.g., Tesei et al., 2013; Giorgetti et al., 2016). Reworking of inherited veins also occurred (Figs. 10b and 11). This inference is consistent with previous U/Pb dating on three calcite-filled veins within the tectonic mélange showing ages of 10.5 ± 2.5 My, 7.0 ± 1.6 My, and 5.1 ± 3.7 My, respectively indicating multiple events of calcite precipitation during deformation (Smeraglia et al., 2019).

5.3 Fluid source and syn-tectonic fluid circulation

Geochemical data indicate calcite precipitation at temperatures between ~100 and ~150 °C from retained pore water, such as marine water retained in the clastic sediments during diagenesis, modified or completely buffered by isotope exchange with the host rock. We base this interpretation on the following evidence:

(1) The calculated $\delta^{18}O$ paleofluid composition, which was in equilibrium with the calcite at the time of mineral growth, ranges between +9.1‰ and +13.7‰ (Fig. 8b). These values indicate ^{18}O enrichment suggesting extensive oxygen exchange between the fluids and the limestone interbeds of clastic deposits. In particular, most of the calculated $\delta^{18}O$ paleofluid compositions ranging between +9‰ and +11‰ (Fig. 8b), indicate high water/rock ratios and paleofluids with nearly constant $\delta^{18}O$ composition, which were responsible for calcite precipitation at progressive increasing temperatures from 108 to 146 °C (Fig. 8b).

(2) Data from fluid inclusions in veins suggest an excess of radiogenic He and Ar in fluids circulating during the precipitation of calcite, indicating a source of both He and Ar other than the atmosphere and the mantle. In particular, the $^{3}He/^{4}He$ ratios are lower than 0.1 R/Ra (Fig. 8d and Table S3) indicating a crustal fluid. In addition, these values are similar to those of crustal-derived fluids (with limited or negligible mantle contribution) enriched in ^{4}He recorded in natural gaseous emissions of the central-northern Apennines (e.g., Buttitta et al., 2020).

(3) The R/Ra ratios of calcite veins are lower than those calculated from actual springs along the Mt. Massico ridge and related to the Roccamonfina volcano (0.39 Ra to 1.99 Ra, typical of mantle-derived fluids; Cuoco et al., 2017), located ~20 towards the NE respect to the study area (Fig. 1), thus excluding the contribution of mantle-derived fluids circulating within the tectonic mélange. In addition, the volcanic activity at Roccamonfina occurred between ~0.6 and ~0.1 My, ~5 My after the latest tectonic activity documented at Mt. Massico (U-Pb vein dated at 5.1 ± 3.7 My, Smeraglia et al., 2019), further excluding the mixing of mantle-derived fluids from Roccamonfina with crustal fluids circulated within the tectonic mélange.

However, we cannot completely exclude that also meteoric water partly infiltrated within the tectonic mélange during deformation and/or exhumation, and was modified by ^{18}O isotope exchange with the host rock, acquiring the calculated $\delta^{18}O$ paleofluid composition.

Based on geochemical and geological data, we suggest that syn-tectonic fluid circulation within the Mt. Massico mélange occurred in a closed system, without a strong connection with external reservoirs (i.e. meteoric or mantle-derived fluids). We therefore propose that during sedimentary and tectonic burial, marine-derived fluids trapped within sandstones and shale interlayers were progressively heated up to 147 °C. During deformation, regional tectonic shortening caused sandstones and shale compaction, porosity reduction, and, eventually, expulsion of previously stored fluids and calcite precipitation into newly-created fractures. In addition, the progressive smectite to illite conversion through mixed layers illite-smectite occurring from temperatures of 60-70 °C to 210 °C (Aldega et al., 2017), may have triggered additional source for local fluids due to water expulsion during clay dehydration (e.g. Moore and Vrolijk, 1992).

Syn-tectonic fluid circulation in closed systems has been already observed in tectonic mélange exposed in various fold-and-thrust belts (i.e. Apennines, Pyrenees) developed within sedimentary succession with alternating of sandstones, marls, and shale (e.g., Vannucchi et al, 2010; Gabellone et al., 2013; Lacroix et al., 2014). The Mt. Massico mélange did not acted as a conduit for external fluids as observed within intra-wedge tectonic mélange in the Apennines (Meneghini et al., 2012), Sicilian (Dewever et al., 2013), Thailand (Hansberry et al., 2015), and Japan (Raimbourg et al., 2015) fold-and-thrust belts. We explain this difference suggesting that meso- and microstructures of the Mt. Massico mélange, such as clay-rich stylolites and marls-shale interlayers (Figs. 3, 4, 6, and 7), created efficient low-permeability barriers, which prevented the ingress of fluids from external reservoirs. Otherwise, the occurrence of regional thrusts, both below and above the Mt. Massico tectonic mélange, may have acted as regional conduits for the drainage of external fluids within the accretionary wedge and away from the tectonic mélange.

We observe an inverse correlation between fluid temperature and ages of calcite veins, showing an increase of calcite precipitation temperature with decreasing ages (Fig. 8c; U-Pb data from Smeraglia et al., 2019). The vein formed at 10.5 ± 2.5 My at temperature of 108 °C suggests an early phase of deformation during wedge accretion (e.g. Tavani et al., 2015). This is not consistent with the burial history proposed by Smeraglia et al. (2019), which shows much lower temperature for the base of the clastic deposits than that recorded by clumped isotopes before the onset of thrusting at 10.5 ± 2.5 My (i.e. Tortonian time; see Fig. 13a in Smeraglia et al., 2019). This can be explained by the lateral and upward migration of deep-seated warm fluids, previously stored in areas already affected by tectonic burial (e.g., Minshull et al., 1989), which circulated within the clastic deposits at shallow depths during the Tortonian. The occurrence pre-Tortonian tectonic burial is highlighted by the borehole stratigraphy of Mara 01 well, located towards the SW respect the Mt. Massico, showing the thrusting of Triassic deposits above Paleocene-Eocene deposits (See Fig. 2 in Smeraglia et al., 2019)

Veins formed at 7.0 ± 1.6 My and 5.1 ± 3.7 My at temperatures of 121 and 147 °C, respectively, are consistent with the progressive burial due to thrust sheet emplacement above the mélange (Smeraglia et al., 2019). However, we point out that most of precipitation temperatures are fully consistent with the maximum burial temperature of 140 °C at depths of ~4 km experienced by the mélange, calculated by 1D thermal modelling constrained by mixed layers illite-smectite (Smeraglia et al., 2019). This indicates that fluid expulsion occurred under maximum burial condition, suggesting that tectonic overburden promoted and/or triggered fluid overpressure and hydrofracturing.

Conclusions

Structural and microstructural observations combined with geochemical data (stable and clumped isotopes and isotope composition of noble gases in fluid inclusions of calcite veins) along the carbonate/shale-bearing tectonic mélange within the central Apennines accretionary wedge (Mt. Massico mélange) show that:

(1) Intra-wedge tectonic mélange generates along pre-existent intra- and interformational rheological contrast occurring along carbonate/shale sedimentary successions. In particular, deformation is localized at the boundary between weak (Tortonian clastic deposits and hemipelagic marls) and competent (Meso-Cenozoic limestones) rocks. The development of tectonic mélange occurred by disruption of the primary bedding, mixing, and deformation of relicts of competent blocks (i.e., olistoliths and relicts of competent strata) within a weak matrix (i.e., deformed shaly and marly interbeds), through recurrent cycles of mutually overprinting brittle (fracturing/veining and frictional sliding) and ductile (pressure-solution) processes, indicating a polyphase deformation history characterized by fast to slow strain rates.

(2) The geochemical signatures of syn-tectonic calcite veins indicate calcite precipitation from warm (108-147 °C) and modified pore fluids, by isotope exchange with the local host rock. Fluid circulation occurred mostly in a closed system dominated by the expulsion and redistribution of pore fluids trapped during sedimentation/diagenesis and/or derived by clay dehydration processes (at $T > 120$ °C), without the interaction of externally derived fluids (i.e. meteoric and/or mantle-derived fluids). A minor contribution from lateral-upward migration of deep-seated warm fluids, previously stored in areas affected by tectonic burial located towards the SW from the Mt. Massico, is inferred in the early phase of the deformation of tectonic mélange. Low-permeability barriers, generated by clay-rich stylolites, promoted the generation of fluid overpressure, which were mostly expelled at maximum burial conditions (T 140 °C and depth of ~4 km) in the final stage of mélange deformation. Therefore, clay-rich intra-wedge tectonic mélanges, along décollement zones, may generate efficient barriers within accretionary wedges for vertical and lateral redistribution of fluids from reservoirs outside the mélange.

We highlight that the integration of geochemical (stable and clumped isotopes, analysis of gaseous species in fluid inclusions) and geochronological (U-Pb dating) methods can be a powerful approach to better constrain the burial-thermal evolution and the fluid storage capacity of fold and thrust belts and sedimentary basins, including the associated fault network.

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Figure Captions

Figure 1. (a) Simplified geological map of the central-southern Apennines (Italy) showing main thrusts and location of the study area. **(b)** Schematic geological cross-section through the central-southern Apennines (modified after Mostardini and Merlini, 1986). (c) Geological cross-section through the Mt. Massico ridge (i.e., the study area).

Figure 2. (a) Detailed geological map of the southwestern part of the Mt. Massico ridge with location of key outcrops where the tectonic mélange is exposed. Schmidt nets (lower hemisphere) showing the attitude of S-C planes with related slip vectors within the tectonic mélange. Modified after Smeraglia et al. (2019) **(b)** Geological cross-sections across the Mt. Massico ridge.

Figure 3. Outcrop-scale structural features of the Mt. Massico tectonic mélange. **(a,b)** The poorly deformed clastic deposits showing bedding and incipient foliation. Note folded strata and low-displacement faults showing ramp and flat geometry. **(c)** The moderately deformed zone showing competent strata (i.e. sandstones and calcarenites interbeds) characterized by pinch-

and-swell and boudine-like geometries. **(d)** Relict of competent strata. **(d)** Detail of competent strata (Fig. 3c) showing veins perpendicular to bedding and S-C foliation. **(f,g)** The strongly deformed part of the tectonic mélange showing S-C foliation within marls and shale and relict of competent strata (limestones and sandstones) showing sigmoidal shapes. **(h)** Relict of competent limestones showing stylolites along S- and C-planes. Inset shows the Schmidt nets (lower hemisphere) showing the attitude of S-C planes with related slip vectors.

Figure 4. Outcrop-scale structural features of the Mt. Massico tectonic mélange. **(a-c)** The strongly deformed part of the tectonic mélange characterized by S-C and S-CC' tectonites within marls and shale and relict of competent strata (limestones and sandstones) showing sigmoidal shapes. **(d)** Relict of competent limestones showing a pervasive network of calcite veins. **(e-h)** Inherited competent (limestones) olistoliths scattered within the weak (marls and shale) scaly foliation, generating the block-in-matrix fabric typical of tectonic mélange. **(h)** Contorted and convoluted foliation within the S-C tectonite.

Figure 5. High-resolution hand sample scan of relicts of competent limestones showing sigmoidal shapes. Stylolites are aligned along S-, C- and C'-planes. Calcite veins are oriented parallel, perpendicular, and/or oblique with respect to stylolites. No stylolites affecting calcite veins and calcite veins cutting through stylolites occur.

Figure 6. Microstructures from the tectonic mélange. **(a-c)** Stylolites affecting host rock relicts and calcite veins. Note the insoluble materials (clays and oxides) filling up to 1-mm thick dissolution seams. **(d,e)** Fibrous veins characterized by fibrous calcite crystals with long axis oriented roughly parallel to stylolites. Note stylolites affecting fibrous veins. **(f)** Fibrous crystals developed along stylolites jogs along C-plane. Fibrous crystals with long axis oriented roughly parallel to stylolites.

Figure 7. Microstructures from the tectonic mélange. **(a,b)** Fibrous veins characterized by fibrous calcite crystals with long axis oriented perpendicular to stylolites. Note that this type of fibrous veins develop within stylolites. **(c,d,e,f)** Blocky veins characterized by blocky calcite crystals. Blocky veins are oriented parallel, perpendicular, and/or oblique respect to stylolites. Note stylolites affecting blocky veins and mutual crosscutting relationships between blocky veins.

Figure 8. (a) $\delta^{13}C$ (‰ V-PDB) versus $\delta^{18}O$ (‰ V-SMOW) diagram from blocky and fibrous veins and host rocks in the Mt. Massico tectonic mélange. Notice the average $\delta^{18}O$ depletion of ~4‰ between mélange-related mineralizations and host rock. **(b)** Oxygen isotope fractionation during equilibrium precipitation: $\delta^{18}O$ of fault-related mineralizations and paleofluid compositions (curves) as a function of temperature. The $\delta^{18}O$ calculated paleofluid compositions (between 9‰ and 14‰) indicate strong ^{18}O exchange with the local limestone host rock. Notice that warmer fluids have similar $\delta^{18}O$ paleofluid composition suggesting that the same paleofluid was the source for calcite precipitation at increasing temperatures and depths. **(c)** Temperature (°C, clumped isotopes) versus age (My, U-Pb dating, data from Smeraglia et al., 2019) diagram for three dated veins. Notice the increase in paleofluid temperature with decreasing ages, that indicates progressive tectonic burial during discontinuous thrust sheet emplacement through time. **(d)** $^{4}He/^{20}Ne$ ratio versus $^{3}He/^{4}He$ (R/Ra) ratio diagram. Note that the calcite veins are in the range of R/Ra values typical of crustal fluids of the central Apennines.

Figure 9. (a) Large-scale tectonic setting of the Apennine subduction (modified after Scrocca et al., 2005) and location of the Mt. Massico structure within the accretionary prism. **(b-c)** Simplified sketch showing the main sedimentary and tectonic events in the Mt. Massico area

(Modified after Smeraglia et a., 2019): **(b)** During late Tortonian-early Messinian times, orogenic compression generated a fault propagation anticline located in the northeastern part of the Mt. Massico ridge, juxtaposing pre-orogenic limestones in the hangingwall with syn-orogenic Tortonian sandstones in the footwall. **(c)** During late Messinian-early Pliocene times, out-of-sequence thrusting occurred and a ~3,300 m-thick stack of imbricate thrust sheets thrust onto the Tortonian-lower Messinian clastic deposits and calcarenites. At this stage the tectonic mélange develop at the base of the clastic deposits. Notice that remnants of such thrust sheets are located in the Mt. Petrino area (Fig. 2).

Figure 10. Simplified evolution of the Mt. Massico tectonic mélange (See Fig. 10b,c for the structural location of the evolution scheme). **(a)** During the early deformation stage, veins perpendicular to bedding develop within the clastic deposits in response to far field tectonic stress. **(b)** Initial shortening affect the clastic deposits generating a weak scaly foliation parallel respect to bedding (see Fig. 3a,b). **(c)** Progressive shortening generated S-C fabric and boudinage of competent bedding (limestones and sandstones). **(d)** During the late stage of deformation, the strongly deformed tectonic mélange develop. In particular, pervasive S-C tectonites develop within the weak (marls and shale) matrix and relicts of competent strata are deformed generating scattered clasts with sigmoidal shape. Notice that hemipelagic marls are scraped off and mixed within the tectonic mélange, while the underlying Meso-Cenozoic limestones are not deformed. The inset shows the different type of veins within the mélange (blocky and fibrous) and their structural position respect to stylolites.

Figure S1. Schematic evolutionary model for mélange-related veins evolution through time. Notice alternate cycles of mutually overprinting brittle (fracturing/veining) and ductile (pressure-solutiona and stylolites generation) processes.

Figure 1

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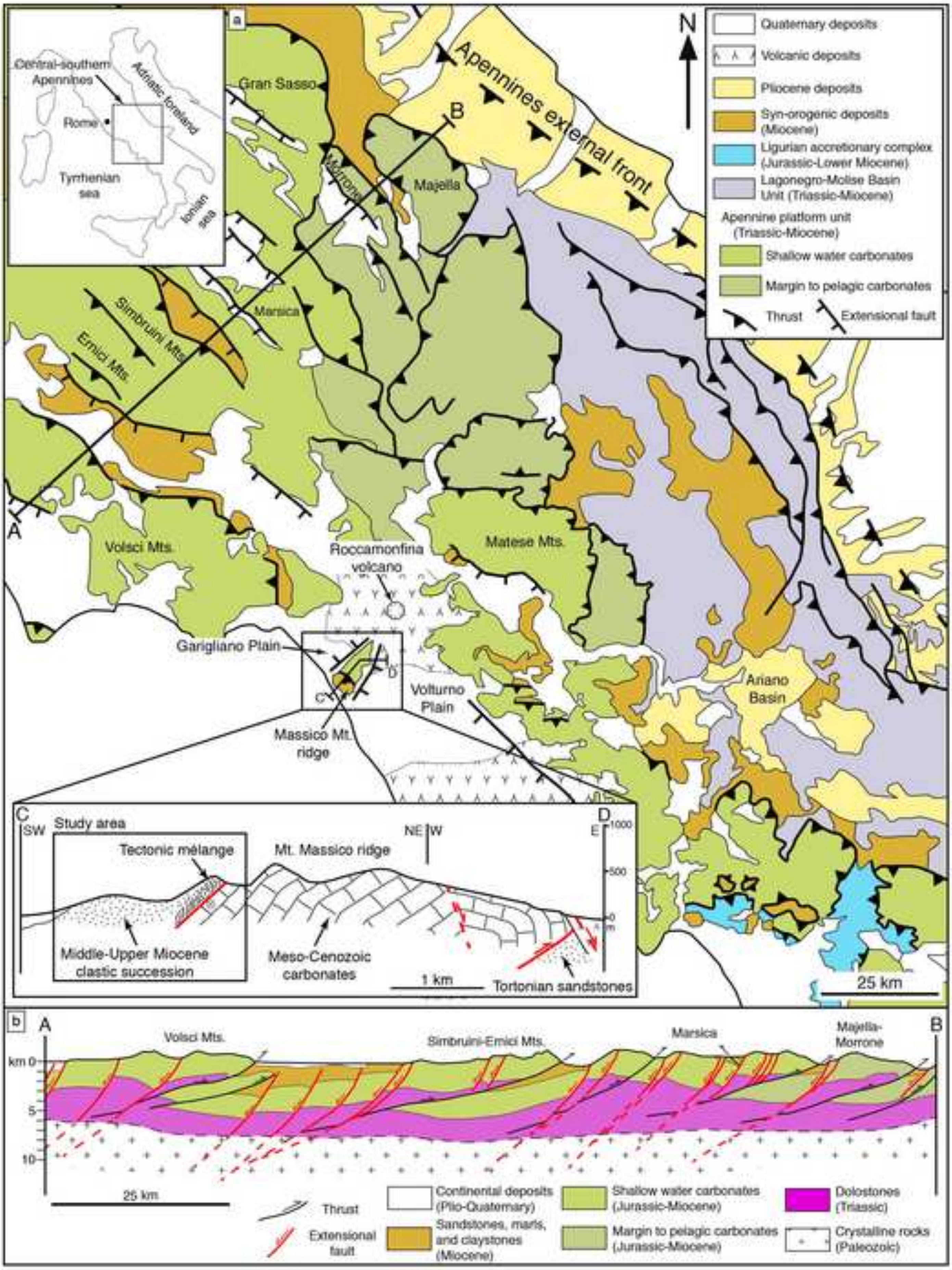

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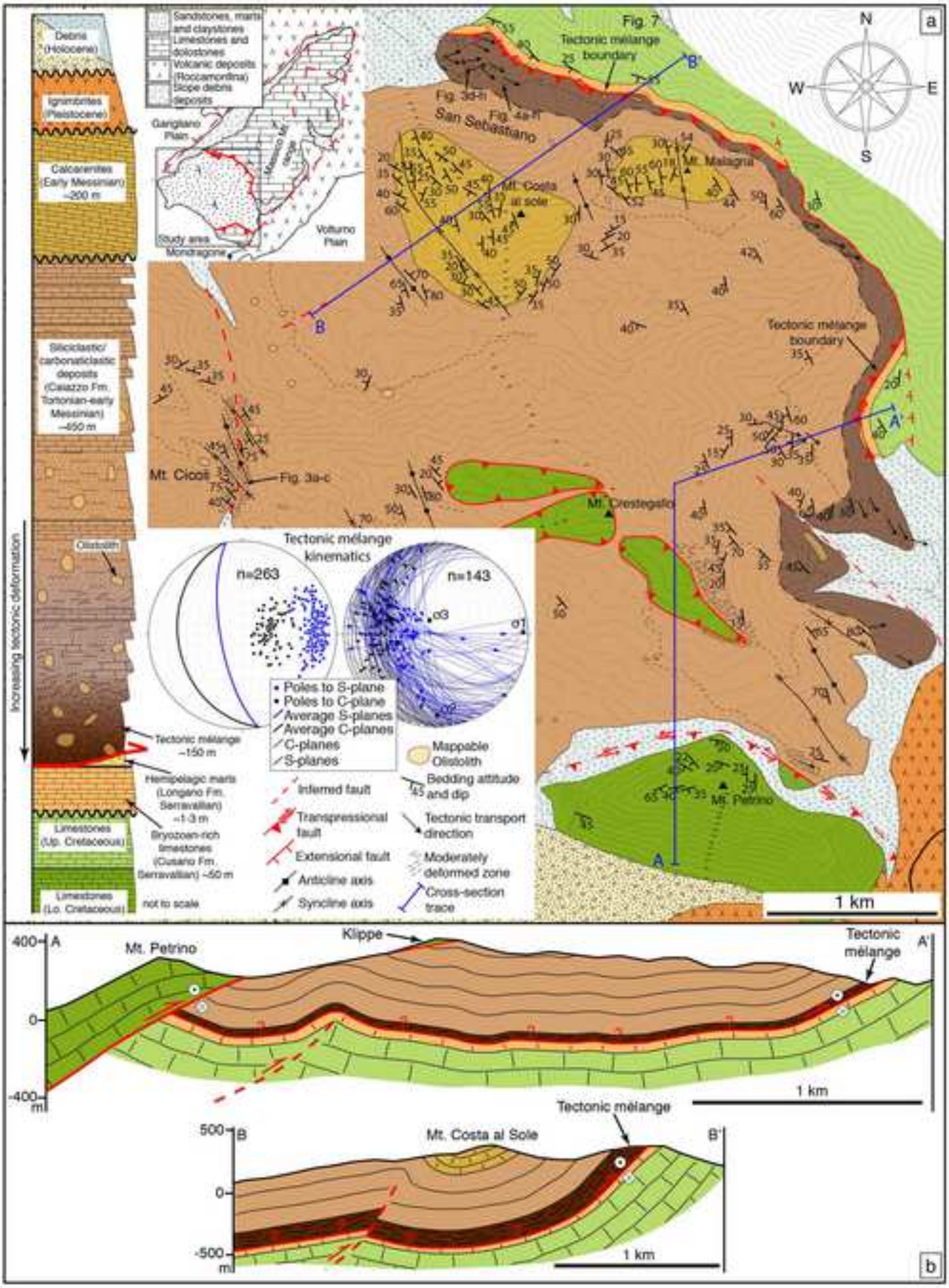

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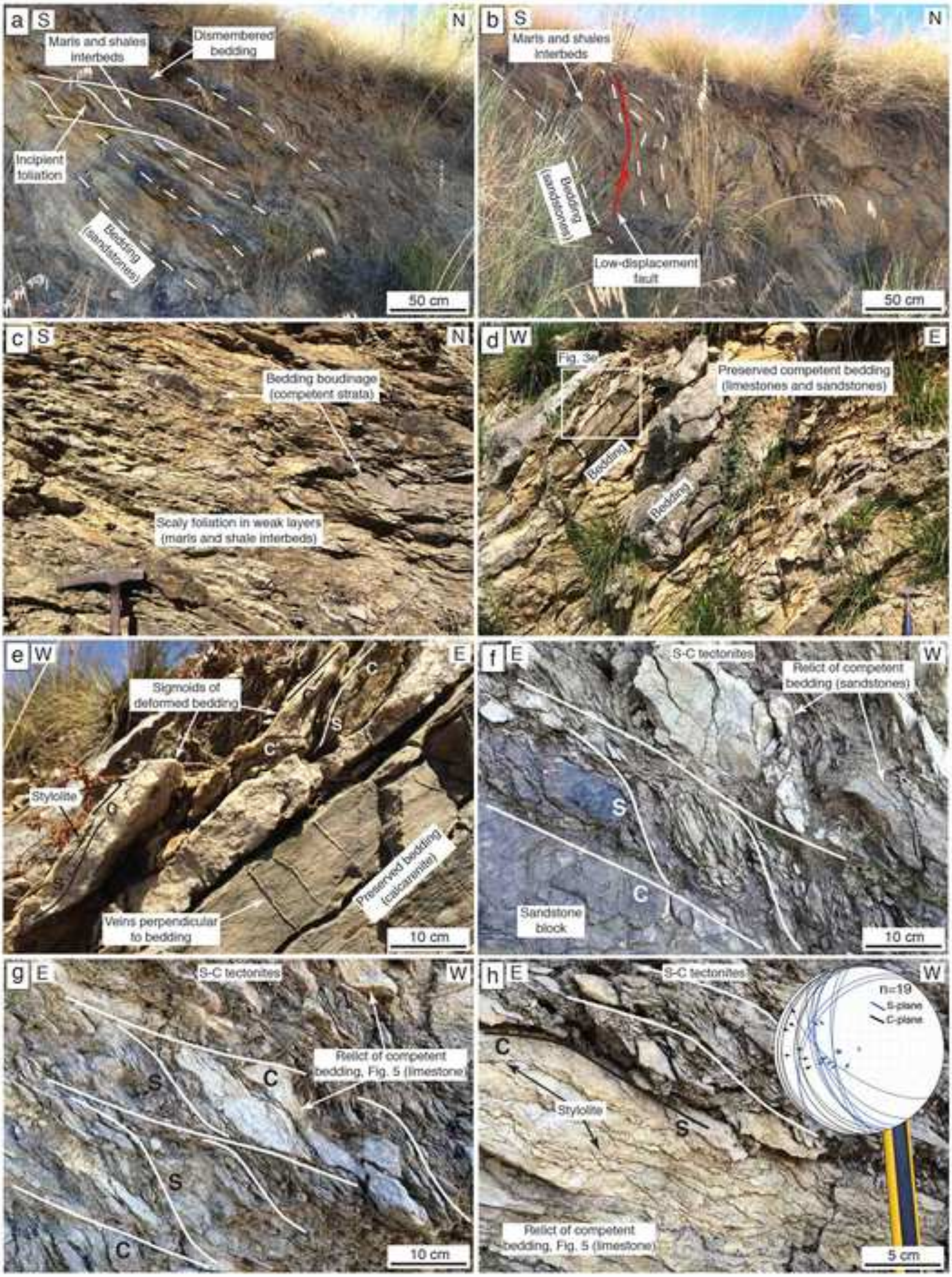

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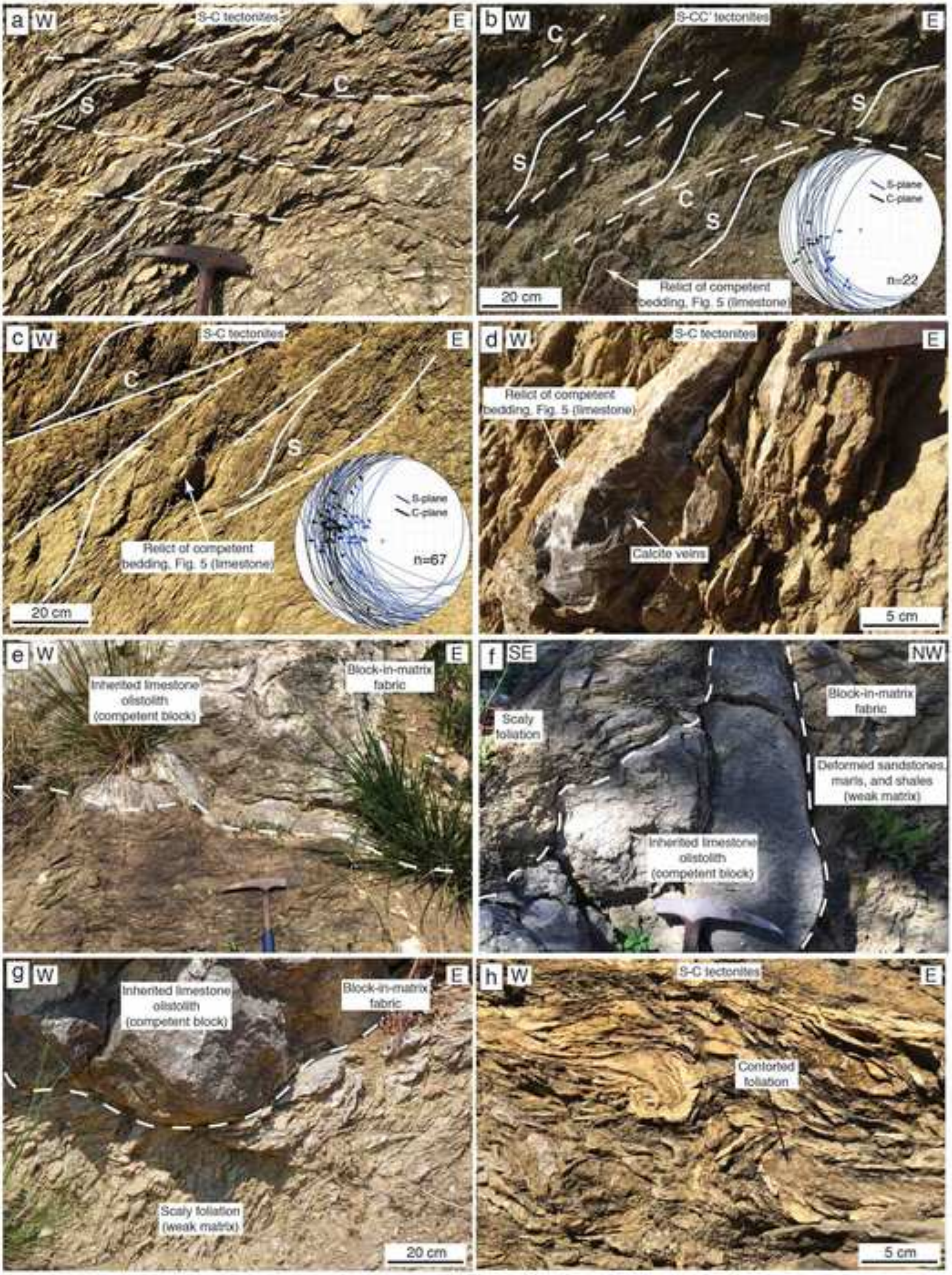

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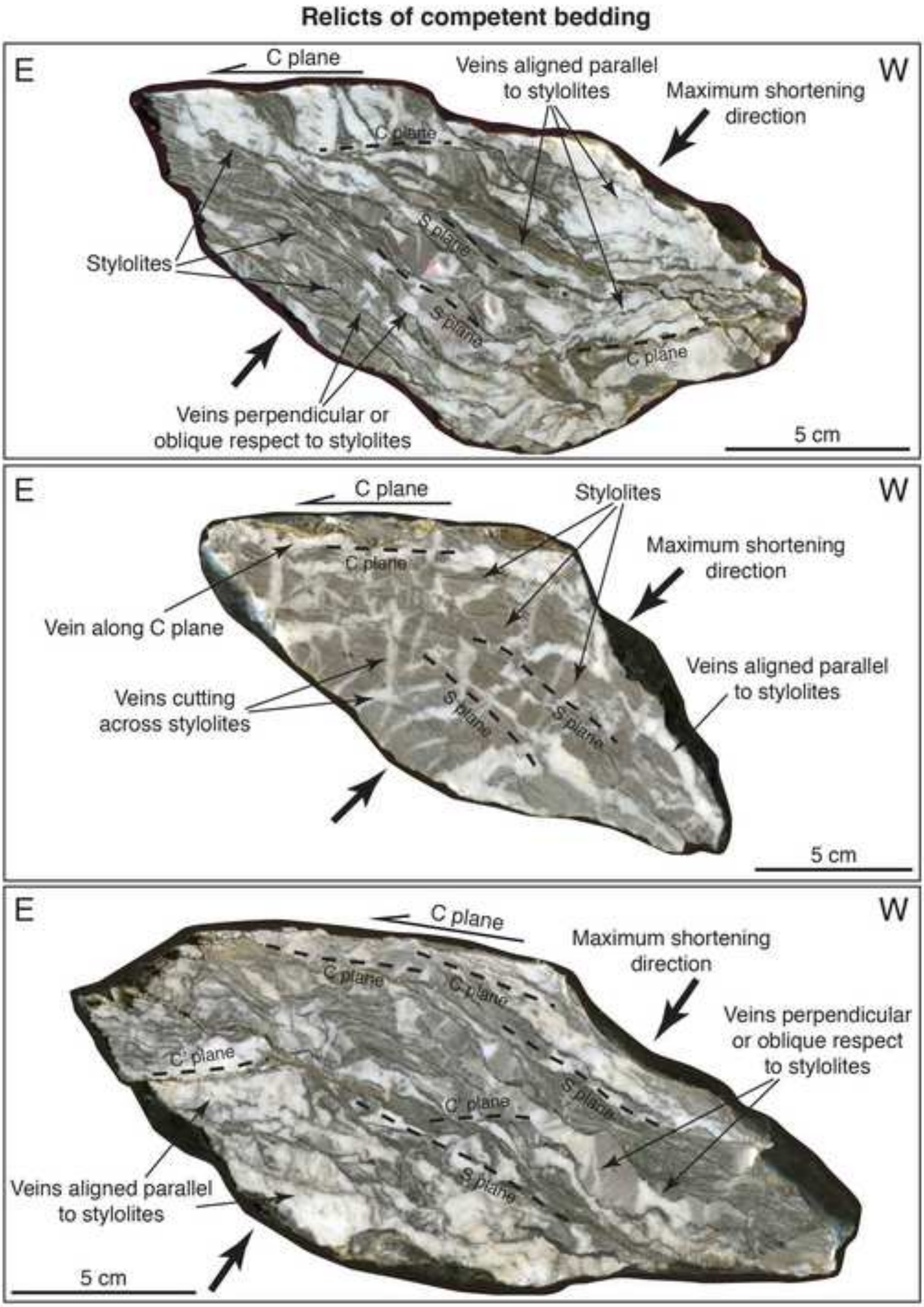

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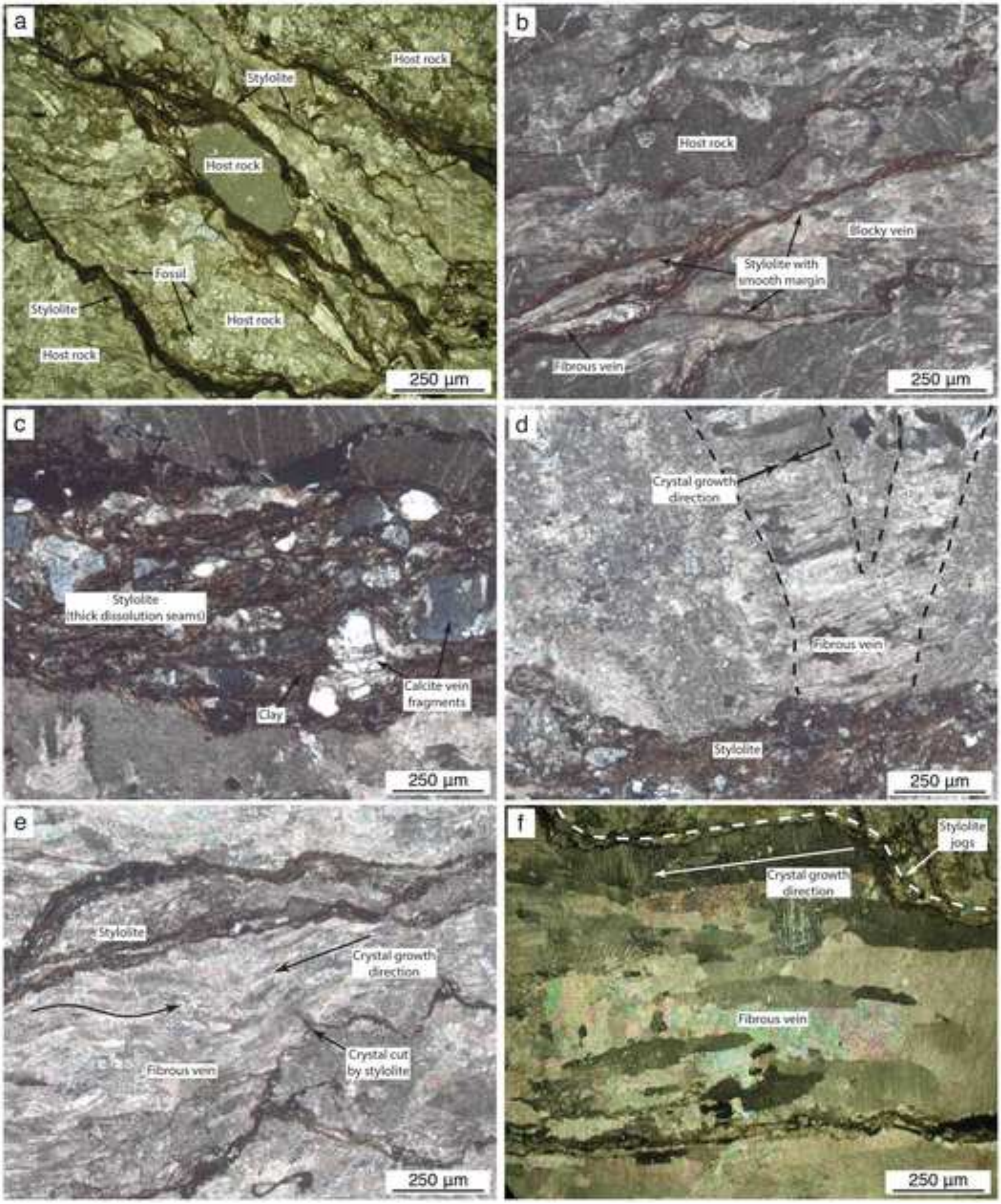

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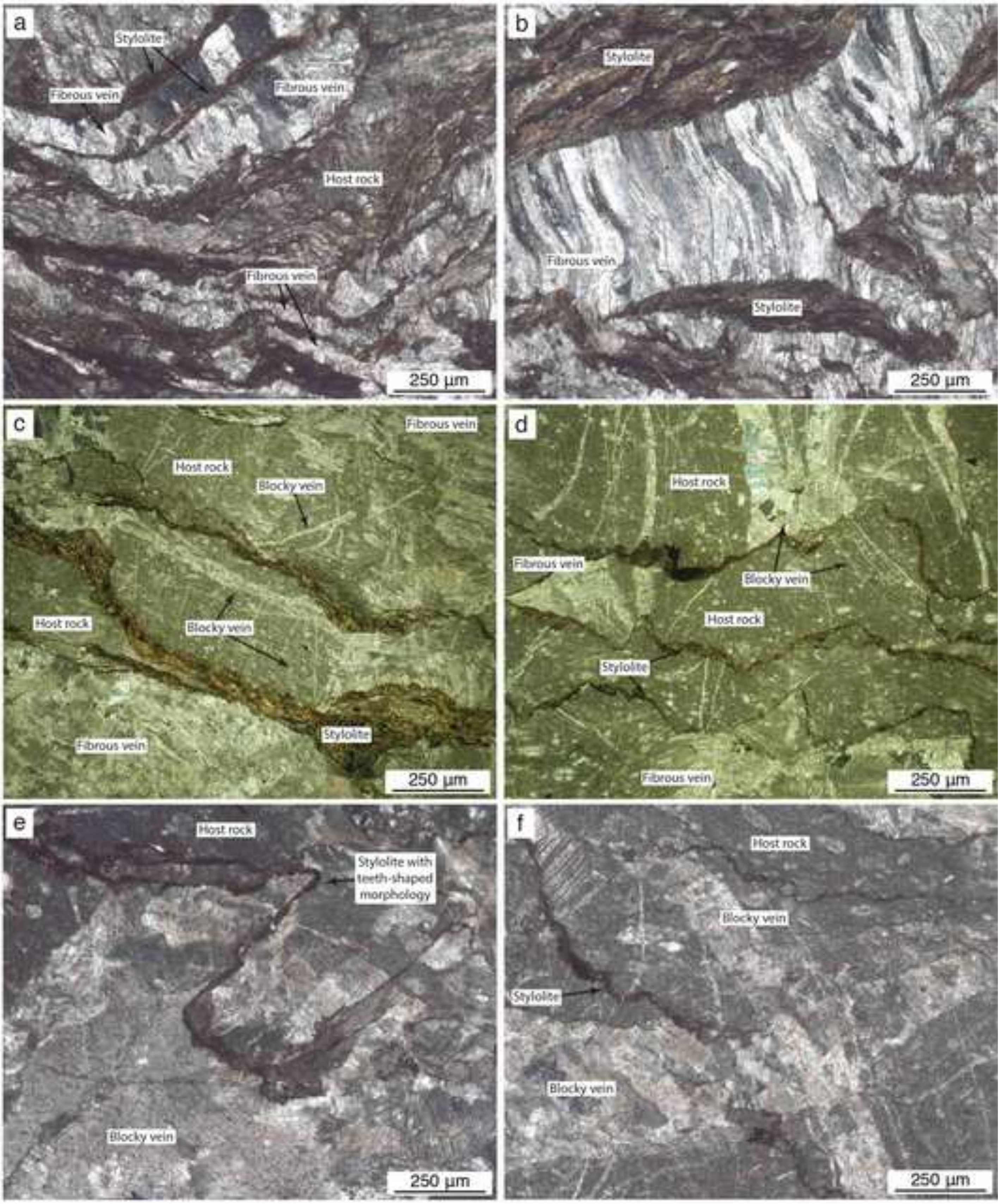

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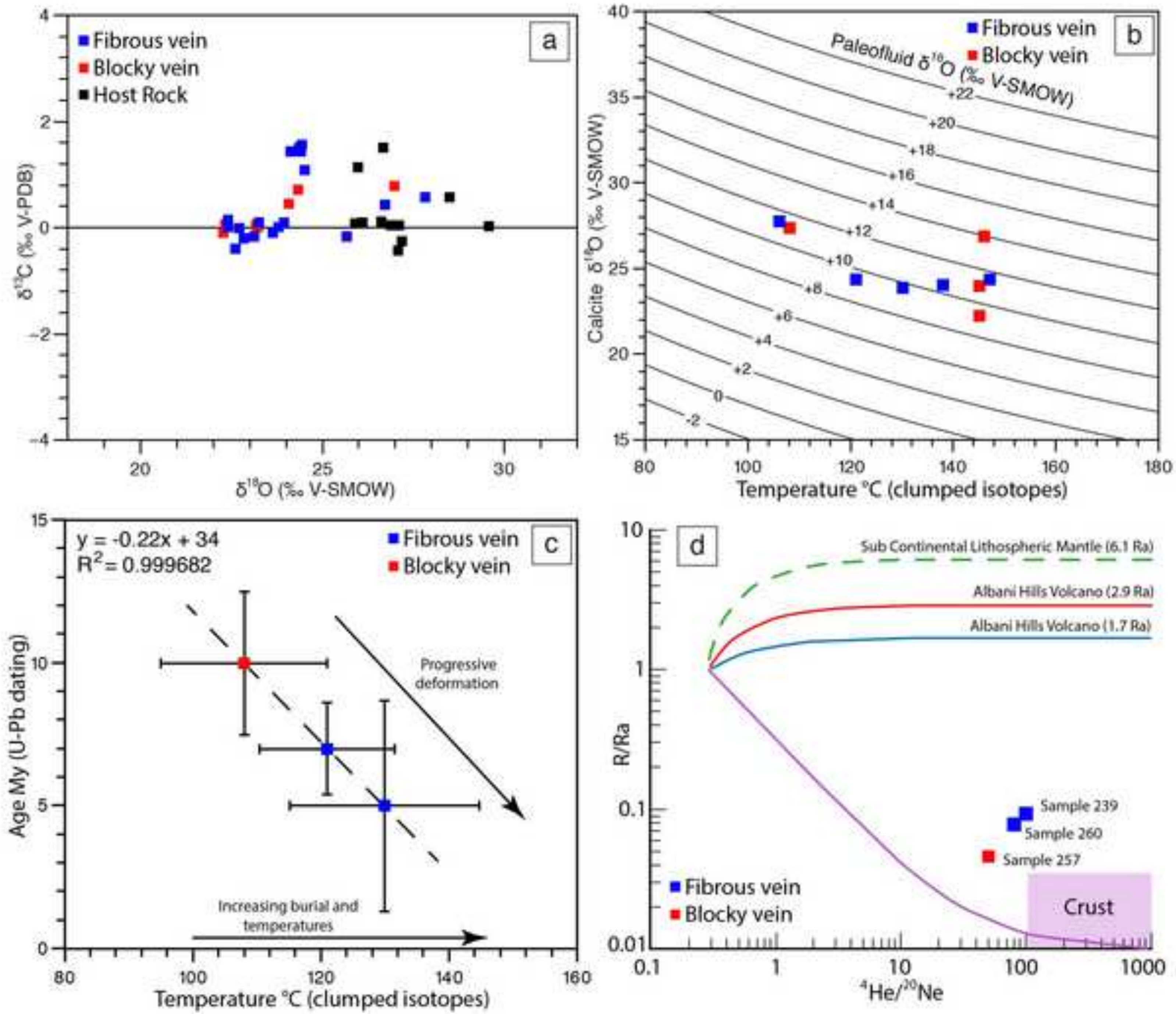

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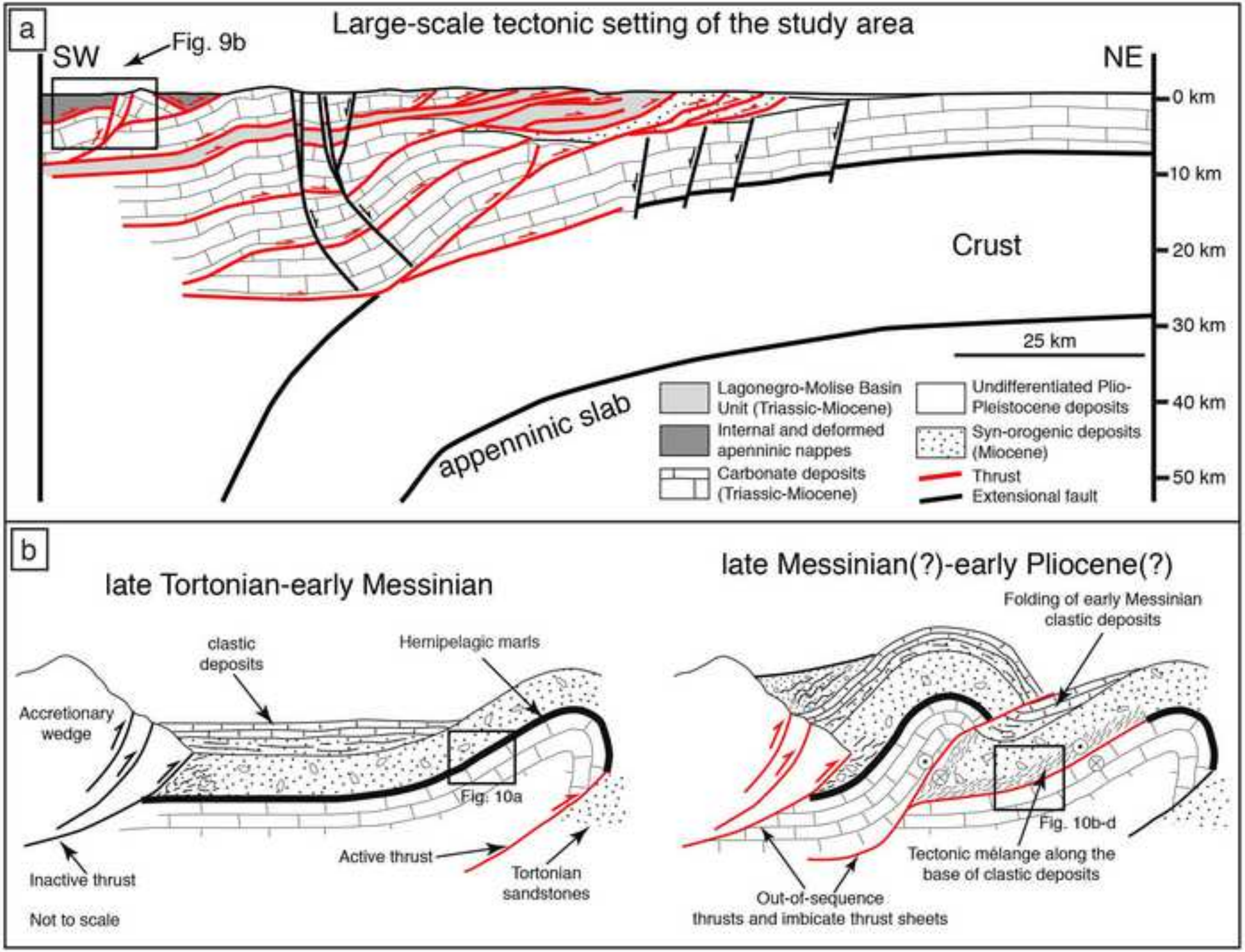

Figure 10
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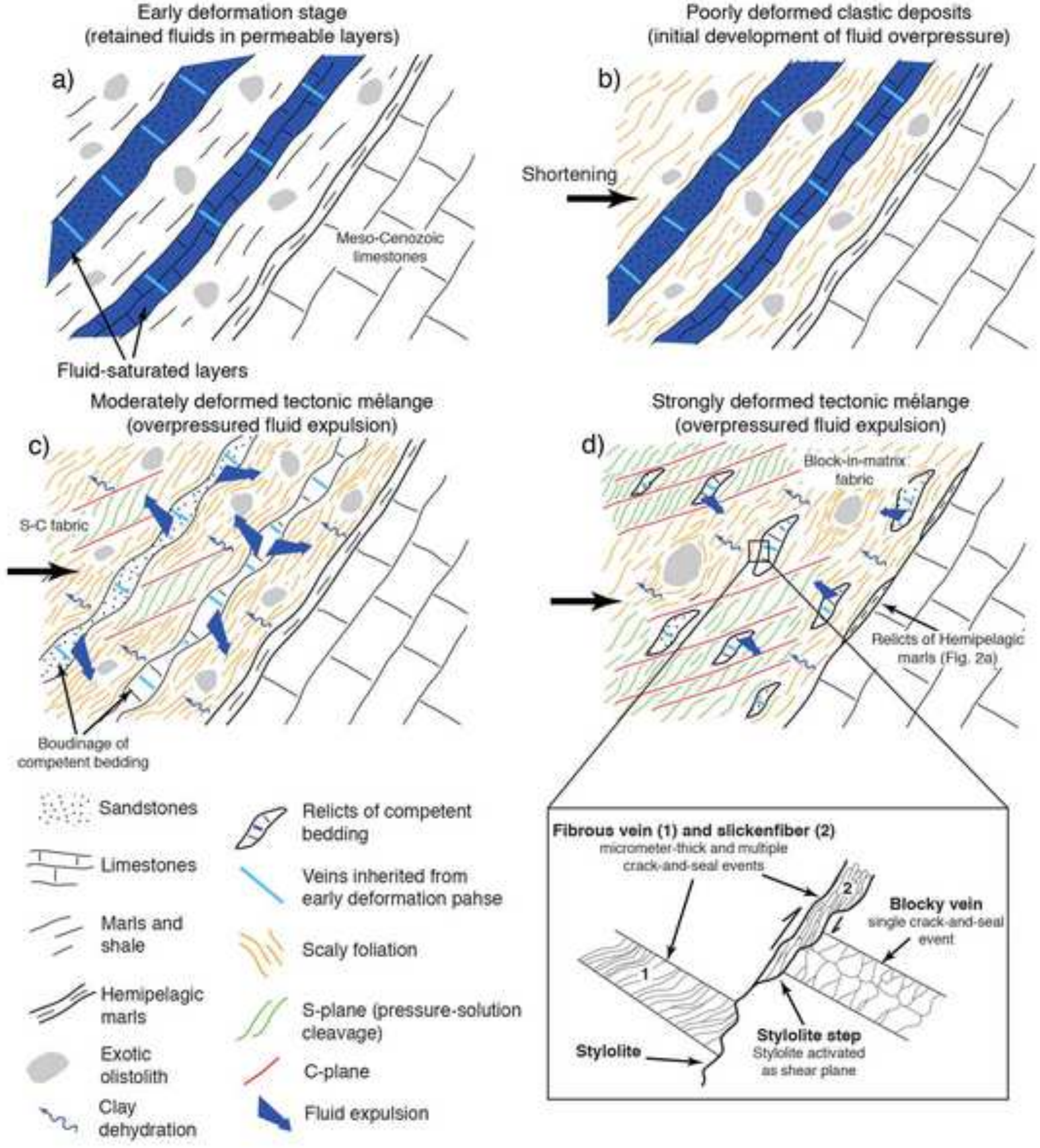

***Declaration of Interest Statement**

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X The authors declare that they have no known competing financial interests or personal relationships that could have appeared to influence the work reported in this paper.

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