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Role of inherited compositional and structural heterogeneity in shear zone development at mid-low levels of the continental crust (the Anzola shear zone; Ivrea-Verbano Zone, Southern Alps).

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Abstract:	The formation of shear zones is crucial to understand the deformation of the crust and the evolution of rifted margins. However, despite their intrinsic importance, a detailed characterization of the compositional and structural patterns of shear zones is often lacking, resulting in poorly constrained models of shear initiation. In this contribution, we reconstruct the pre-shearing lithological, structural and metamorphic proprieties of rocks forming a major, rift-related shear zone with the aim to assess the conditions promoting the strain localization. We focus on the Anzola shear zone, a major extensional structure from one of the best-preserved cross-sections through the middle to lower continental crust of a fossil passive margin, the Ivrea-Verbano Zone (Southern Alps, Italy). Until now, the Anzola shear zone is believed to have developed within a rheologically hard and isotropic gabbro rather than in the surrounding weaker and anisotropic volcano-sedimentary metamorphic sequence. New geological mapping shows that significant pre-existing heterogeneities related to the composition and deformation structures, characterize the Anzola shear zone. Field relationships and geochemistry reveal that the (ultra-)mylonitic rocks overprinted a multi-lithological sequence that have already experienced Variscan deformation and late Variscan High-Temperature metamorphism, at the boundary between amphibolite and granulite facies. Our in-depth trace elements study is shown to be a powerful tool in reconstructing the pre-shearing relationships between wall rocks and mylonites and determining the protoliths of tectonites. Estimated P-T conditions indicate that mylonitic deformation started at high temperature (~820°C) with presence of melt and continued as solid-state deformation down to amphibolite facies (~650°C), following a retrograde path. We argue that strain localization was promoted by the combination of rheological boundaries derived from pre-existing conditions, including: i) compositional and structural anisotropies of the volcano-sedimentary metamorphic sequence contrasted by ii) the close intrusion of a nearly isotropic gabbro and iii) the presence of melt within the metamorphic boundary depicted by the transition between granulites (dominated by anhydrous minerals) and amphibolite facies (dominated by hydrous minerals). Our findings finally suggest that pre-existing significant heterogeneities relate to rock composition, deformation and metamorphism represent the preferential loci for strain localization controlling the initiation and development of rift-related structures in the mid to lower crust of passive margins.

Role of inherited compositional and structural heterogeneity in shear zone development at mid-low levels of the continental crust (the Anzola shear zone; Ivrea-Verbano Zone, Southern Alps).

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Highlights

1. Field mapping and chemical analysis of an extensional shear zone from middle/lower crust
2. Trace element analysis reveals composite protoliths of the mylonites
3. Anisotropic and/or hydrated rocks are folded with isotropic and/or anhydrous layers
4. Mylonites postdates folding and upper-amphibolite-granulite facies metamorphism
5. Pre-existing contrasted rheological boundaries promoted strain localization

Abstract

The formation of shear zones is crucial to understand the deformation of the crust and the evolution of rifted margins. However, despite their intrinsic importance, a detailed characterization of the compositional and structural patterns of shear zones is often lacking, resulting in poorly constrained models of shear initiation. In this contribution, we reconstruct the pre-shearing lithological, structural and metamorphic proprieties of rocks forming a major, rift-related shear zone with the aim to assess the conditions promoting the strain localization. We focus on the Anzola shear zone, a major extensional structure from one of the best-preserved cross-sections through the middle to lower continental crust of a fossil passive margin, the Ivrea-Verbano Zone (Southern Alps, Italy). Until now, the Anzola shear zone is believed to have developed within a rheologically hard and isotropic gabbro rather than in the surrounding weaker and anisotropic volcano-sedimentary metamorphic sequence. New geological mapping shows that significant pre-existing heterogeneities related to the composition and deformation structures, characterize the Anzola shear zone. Field relationships and geochemistry reveal that the (ultra-)mylonitic rocks overprinted a multi-lithological sequence that have already experienced Variscan deformation and late Variscan High-Temperature metamorphism, at the boundary between amphibolite and granulite facies. Our in-depth trace elements study is shown to be a powerful tool in reconstructing the pre-shearing relationships between wall rocks and mylonites and determining the protoliths of tectonites. Estimated *P-T* conditions indicate that mylonitic deformation started at high temperature (~820°C) with presence of melt and continued as solid-state deformation down to amphibolite facies (~650°C), following a retrograde path. We argue that strain localization was promoted by the combination of rheological boundaries derived from pre-existing conditions, including: i) compositional and structural anisotropies of the volcano-sedimentary metamorphic sequence contrasted by ii) the close intrusion of a nearly isotropic gabbro and iii) the presence of melt within the metamorphic boundary depicted by the transition between granulites (dominated by anhydrous minerals) and amphibolite facies (dominated by hydrous minerals). Our findings finally suggest that pre-existing significant heterogeneities relate to rock composition, deformation and metamorphism represent the preferential *loci* for strain localization controlling the initiation and development of rift-related structures in the mid to lower crust of passive margins.

Keywords: rheological boundaries, granulite-amphibolite facies transition, continental crust, strain localization, Tethyan rifting, Ivrea-Verbano Zone

1. Introduction

Shear zones accommodate deformation during rifting evolution from the central ridges toward the passive margins. How strain is distributed within the lithosphere at different scale and where it localizes to initiate the development of shear zones, are key questions in geodynamic studies (Lavier and Manatschal, 2006; Pennacchioni and Mancktelow 2007; Whitney et al., 2013).

Strain is localized in spatially restricted but continuous bands of reduced viscosity. The process of strain localization has been attributed to several microscale factors: brittle fracturing, grain size reduction with and without deformation mechanism changes, reaction softening, phase mixing, thermal softening and the presence of melt/fluid (e.g., Arzi, 1978; Brodie and Rutter, 1987; Mei et al., 2002; Rybacki and Dresen, 2004; Svahnberg & Piazolo 2010; Lee et al., 2020; Gardner et al., 2020; Casini et al., 2021). These microscale processes may be accentuated by pre-existing heterogeneities such as mechanically weak layers (Passchier, 1982), stress concentrations around hard layers including compositionally contrasted rock pairs (e.g., Pennacchioni and Mancktelow 2007), inherited fault zone where fluid flow within damage zone will further enhance strain localization (e.g., Austrheim, 2013). Among these, lithological heterogeneities likely represent the most common factor in controlling strain localization at all scales (Fossen and Cavalcante, 2019 and references therein), since compositional boundaries may often coincide with significant viscosity contrast (Treagus & Sokoutis, 1992; Smith et al. 2015).

In summary, pre-existing and/or inherited structural and compositional heterogeneities are commonly loci of shear zones development. In such zones of strain localization, fluid flow is common resulting in micro to macro-scale rheological effects based on metamorphic reactions involving hydration and metasomatism (e.g., Austrheim, 1987; Yardley et al., 2014). In turn, metamorphism triggering a change from a hydrous to anhydrous assemblage (e.g., reactions defining the amphibolite to granulite facies) is commonly accompanied by the release of free fluids and/or partial melting depending on the prevailing PT conditions (i.e., migmatites) that may further contribute to strain localization at or near metamorphic facies boundaries (e.g., e.g., Austrheim, 2013; Getsinger et al., 2013; Yardley et al., 2014; Smith et al. 2015; Gardner et al., 2020).

At the regional scale, pre-existing compositional/mechanical heterogeneities and their position (in depth) in the continental lithosphere have a first-order control on the development of shear zones driving the evolution of rifted margins (Michibayashi and Mainprice, 2003; Lavier and Manatschal, 2006; Mohn et al., 2012; Manatschal et al., 2015; Langone et al., 2018; Petri et al., 2019). Mechanically weak boundaries were, indeed, documented to produce detachment faults and decoupling horizons, especially at mid-crustal depth in magma-poor rifted margin during major lithospheric thinning (Prosser et al., 2003; Mohn et al., 2012). Nevertheless, detailed reconstruction of their compositional and geometrical pattern is often poorly constrained due to lack of correlations between field relationships, microstructure and geochemical characterization. The Ivrea-Verbano Zone (IVZ; Fig. 1), in the Italian Southern Alps, preserves remnants of the former Alpine Tethys rifted margin (e.g., Rutter et al., 1993; Petri et al., 2019) where localization of detachment faults has been related to inherited structures (Beltrando et al., 2015; Manatschal et al., 2015). However, the role of geometrical and compositional heterogeneities of these structures at the outcrop scale has been poorly investigated. Therefore, new detailed petrological, structural and geochemical investigations are required for a robust interpretation of the mechanisms leading to the initiation of deformation.

In this paper, we investigate a rift-related shear zone developed in the middle-lower crust with the aim to determine compositional, structural and/or metamorphic gradients or breaks from wall rock to shear zone centre. We selected the Anzola shear zone (Val d'Ossola, Northern Italy) because it represents a well-exposed high-temperature extensional fault (Rutter et al., 2007; Simonetti et al., 2021) allowing us to explore the relationships between strain localization and compositional/structural heterogeneities. Until now, the Anzola shear zone has been described as a fault zone affecting gabbroic rocks bounded by amphibolite facies metapelite and felsic granulite (e.g., Brodie, 1981). Previous works assumed that the compositional homogeneity of the gabbro minimizes the role of initial chemical/structural heterogeneity in the shear zone localization. This leads to consider the strain localization as driven by processes involving dynamic recrystallization and metamorphic reactions (Brodie, 1981; Altenberger, 1997; Stünitz, 1998). Moreover, the role of the amphibole in both mechanical and metamorphic processes has been stressed to explain the paradox of shear zone localization in the mechanically stronger layer (i.e., gabbros). This contrasts the fact that, strain localization is usually expected in the surrounding weaker quartz-feldspar-rich layers (Rybacki and Dresen, 2000), unless there are specific processes, such as grain size reduction leading to diffusion creep, that make them stronger (e.g., Pearce and Wheeler, 2011). On the base of microstructural and chemical changes of amphiboles, shear zone development was initially defined under prograde *P-T* conditions (Brodie, 1981), however, other observations led to attribute strain localization during retrograde *P-T* path (Altenberger, 1997; Stünitz, 1998).

In this work, we perform new detailed geological mapping of the Anzola shear zone with the aim to investigate the underlying reasons for strain localization. We combine the analysis of field relationships with petrological and geochemical characterization of samples collected across the shear zone. For the first time we integrate bulk and mineral analyses of major elements with the trace elements geochemical data. We found that the trace elements study is the most powerful tool in order to test the links between wall rock and shear zone and thus determine the protoliths of mylonites. The coupling of different geothermobarometers was used to determine the pre- and syn-deformation conditions of the studied area. Our results highlight that the Anzola shear zone is characterized by a complex pattern of lithotypes, which experienced multiple tectono-metamorphic stages before shear zone development. We argue that pre-existing lithological and metamorphic meso- to micro-scale variations associated with the presence of melt promoted strain localization resulting in the Anzola shear zone. We finally discuss these new data in the framework of the regional rifting evolution.

2. Geological setting

2.1 The Ivrea-Verbano Zone

The IVZ in north-western Italy represents a cross-section through the middle to lower continental crust of the Southern Alpine basement (Fig. 1). During Jurassic extension, the IVZ was subjected to a first uplift phase, that was accelerated and converted in substantial vertical exhumation during the Alpine collision and associated open folding. As such the IVZ escaped alpine metamorphic overprint (Henk et al., 1997; Rutter et al., 2007; Wolff et al., 2012). Northward, the IVZ is separated from the Alpine metamorphic rocks (i.e., Sesia Zone) by the Insubric Line, while south-eastward, it is divided from the Serie dei Laghi Unit by the Late-Variscan Cossato–Mergozzo–Brissago Line, which is in turn locally crosscut by the Pogallo Line (Fig. 1; Boriani et al., 1990).

Three main units are described in the IVZ (Fig. 1), from NW-SE: i) the mantle peridotites; ii) the Mafic Complex, and iii) the Kinzigite Formation.

The main peridotitic bodies – e.g., Finero and Balmuccia - crop out close to the Insubric Line. These peridotitic lenses lie in different stratigraphic levels of the crust. The crustal emplacement of these rocks is considered a consequence of the tectonic evolution of the Kinzigite Formation that occurred in the end of the Variscan orogeny and before the intrusion of the Mafic Complex (Quick et al., 2003).

The Mafic Complex is composed of gabbros and diorites, variably overprinted by deformation and metamorphism, intruded into the metasedimentary sequences (i.e., schists and gneisses) of the Kinzigite Formation. In the south-western part of the IVZ, the transition between the upper and lower Mafic Complex corresponds to a zone, identified as the “paragneiss-bearing belt”, where paragneiss septa (i.e., depleted granulite) are interlayered with igneous rocks (Fig. 1; Quick et al., 2003). The Mafic Complex mainly intruded during Permian (290-270 Ma) with underplating dynamics and coeval with the acid magmatism and volcanism recognised in the upper crust (Peressini et al., 2007). Mantle-derived mafic magmatism probably started before the Permian, as indicated by the presence of mafic sills intruded during Carboniferous (e.g., ~314 Ma; Klötzli et al., 2014), and locally occurred even during Triassic and Jurassic (e.g., Zanetti et al., 2013; Denyszyn et al., 2018).

The Kinzigite Formation is considered as the originally upper part of the tilted crustal section made up of a heterogeneous group of metasediments, belonging to a volcano-sedimentary sequence, comprising mainly metapelites with intercalated metabasic rocks (i.e., volcanic sediments and MORB-like lavas) and minor marbles/calc-silicates and quartzites (Fig. 2; Zingg, 1990; Schmid, 1993; Quick et al., 2003; Kunz et al., 2014). At a regional scale, preserved peak metamorphic grade decreases from granulite (~900 °C and 900 MPa) to amphibolite facies (~600 °C and ~400 MPa) from NW to SE (Brodie and Rutter, 1987; Zingg, 1990; Schmid, 1993; Redler et al., 2012; Kunz et al., 2014). Long-lasting high-grade metamorphism developed between the Late Carboniferous (~316 Ma; Ewing et al., 2013) to the Early Permian, coevally with the Mafic Complex intrusion (~290 Ma; Ewing et al., 2013).

At the mid to lower crustal levels the transition between amphibolite to granulite facies is marked by a ~1-5-kilometre-wide zone that experienced pronounced migmatization processes (Fig. 2A; e.g., Redler et al., 2012; Kunz et al., 2014). In the north-eastern part of the IVZ, this transition zone hosts two mylonitic shear zones, namely the Anzola (Brodie and Rutter, 1987) and the Forno-Rosarolo shear zones (Siegesmund et al., 2008; Fig. 1). Although described as separate shear zones and often named in different way (see Simonetti et al., 2021), these structures have been interpreted as a single fault system with vertical foliation and NNE-SSW strike in present-day orientation (e.g., Rutter et al., 2007).

2.2 Tectonic evolution

The structure of the IVZ is the result of a complex tectonic evolution that developed between the Variscan and the Alpine orogeny. During the late stages of Variscan convergence (290-320 Ma), the IVZ went through a lithospheric delamination that triggered magmatic underplating and lead to polyphase deformation under amphibolite to granulite facies conditions (Handy et al., 1999). Variscan large-scale folding phases are described in the Kinzigite Formation of the IVZ (Brodie and Rutter, 1987; Rutter et al., 2007). Polyphasic folding is well preserved in the main upright fold of the Massone antiform that extends within the Kinzigitic Formation around 40 km NE-SW with a hinge line strongly curved through an angle of 115° within the axial plane. In particular, the Massone axial planar crenulation cleavage deforms pre-existing schistosity and lineation associated with an earlier folding phase, finally resulting in a Type 2 interference geometry with perpendicular fold hinges and axial planes (Fig. 2A; Rutter et al., 2007).

Since the Permian (270-290 Ma), the IVZ experienced post-orogenic extension associated with magmatic underplating of mafic melts occurred along fault systems in all crustal levels (Boriani et al., 1990; Handy et al., 1999; Rutter et al., 2007; Siegesmund et al., 2008). Stretching led to post-orogenic extension and incipient exhumation of the lower crust supported by mylonitic shearing, as testified by the Cossato-Brissago-Mergozzo Line (CMBL; Fig.2A; Brodie & Rutter, 1987; Henk et al., 1997).

In the Triassic-Jurassic time interval (230-180 Ma), the IVZ was involved in complex and polyphase rift tectonics (Beltrando et al., 2015; Petri et al., 2019 and reference therein). Crustal thinning was accommodated by several shear zones active during different phases of rifting (Mohn et al., 2012; Manatschal et al., 2015) at different crustal levels (e.g., Beltrando et al., 2015; Petri et al., 2019 and reference therein). The most prominent rifting-related structure in the southern portion of the IVZ is the Pogallo Line, which is interpreted as a low-angle normal fault that accommodated thinning between Triassic and Jurassic age (ca. 210 and 170 Ma; Zingg, 1990; Mulch et al., 2002; Wolff et al., 2012) under decreasing temperatures from amphibolite- to greenschist-facies conditions. In the northern sector of the IVZ, Late Triassic to Early Jurassic ductile shear zones developed within (ultra)mafic rocks (e.g., Langone et al., 2018; Corvò et al., 2020) at granulite- to greenschist-facies conditions (Brodie, 1981; Kenkmann and Dresen, 2002; Langone et al., 2018). Important rift-related structures have been described also in the central part of the IVZ, i.e., the Forno-Rosarolo and the Anzola shear zones (Figs. 1, 2A; Siegesmund et al., 2008; Simonetti et al., 2021). Onset of extension occurred since the upper Triassic as suggested by U-Pb geochronological data relative the emplacement of felsic dykes in lower crust lithologies (e.g., Schaltegger et al., 2015; Bonazzi e al., 2020 and references therein). Exhumation of the lower crust occurred along a large, noncoaxial mylonitic shear zone, i.e., Pogallo shear zone, which was linked to asymmetrical rift basins in the upper crust (Handy et al., 1999).
As last stage, during Alpine orogeny (20-50 Ma), the IVZ accomplished the present-day sub-vertical attitude through a rotation of ca. 60° around a horizontal axis striking parallel to the Insubric Line in a clockwise sense

viewed to the north (Henk et al., 1997; Rutter et al., 2007; Wolff et al., 2012). Consequently, the entire sequence experienced large-scale open folding (i.e., Proman antiform, Fig. 2A) and brittle faulting (Rutter et al., 2007).

2.3 The Anzola shear zone

The Anzola shear zone crops out in a dismissed quarry, 500 m east from the Anzola village in Val d'Ossola (Fig. 2B; Brodie., 1981; Brodie et al., 1989, Stünitz, 1998). Here, mylonitic amphibolites were the objects of several structural, microstructural and geochemical studies (Brodie, 1981; Brodie and Rutter, 1987; Brodie et al., 1989; Rutter et al., 1993; Altenberger, 1997; Stünitz, 1998; Rutter et al., 2007). The Anzola shear zone has always been described as an extensional amphibolite-facies mylonitic belt overprinting mainly gabbroic rocks.

Brodie (1981) was the first to study one mylonitic band of the Anzola shear zone, focusing on the role of deformation on mineral and rock chemistry. The author showed that the plagioclase and amphibole composition, varies in relation with the intensity of shearing, i.e., Fe# ($Fe^{2+}/(Fe^{2+}+Mg)$) and Ca increase within amphibole and plagioclase, respectively. These changes in mineral chemistry together with preliminary microstructural investigations suggested that shear zone developed with increasing temperature during prograde regional metamorphic conditions from low- to high-grade amphibolite facies.

Altenberger (1997), in agreement with Brodie (1981), highlighted an increase in modal abundance of amphibole and grain size contrast within mylonitic layers with respect to the wall rocks. Moreover, the author reported a more extensive recrystallization of amphibole with respect to plagioclase. According to Altenberger (1997), the pre-existing heterogeneity (i.e., general sample layering and grain-size variations) was the *locus* of concentrated shear deformation.

The structural and geochemical features of the shear zone were further investigated by Stünitz (1998) who described a 10 metres wide N-S striking mylonitic belt showing a vertical foliation, discordant on the Variscan metamorphic schistosity (i.e., fold axial plane cleavage), with sharp boundaries with the surrounding host rocks. According to the author, two types of shear zones were recognisable on the base of their microstructure, grain size and mineral abundance. He confirmed that syn-tectonic recrystallization of plagioclase, clinopyroxene and amphibole produces compositional differences between porphyroclasts, and recrystallized grains as previously recognised by Brodie (1981). For what concern the porphyroclasts, recrystallized clinopyroxene has higher Mg# values (Mg/Mg+Fe^{2+}) and Al contents, syn-kinematic hornblende is characterized by lower Ti and higher Mg# values whereas plagioclase grains have lower anorthite and orthoclase components. Stünitz (1998) concluded that the Anzola shear zone recorded *P-T* conditions indicative of the transition from amphibolite to granulite facies, likely with a prograde initiation as suggested by Brodie (1981). However, the main activity of the shear zone was attributed to a retrograde *P-T* path, under amphibolite facies conditions (from about 550 to 650 °C) at pressures lower than 800 MPa.

Brodie et al. (1989) provided an attempt to date shear zone activity. Ar-Ar radiometric dating of hornblende were carried out from the undeformed gabbro and mylonitic amphibolites obtaining a minimum ^{40}Ar–^{39}Ar age of about 247 Ma for unsheared gabbroic rocks and about 210-215 Ma for syn-kinematic amphibole grains. The authors suggested that crustal extension started around 280 Ma, after the emplacement of the gabbro, and was the beginning of a long-lasting period of crustal thinning and cooling of more than 100 Myr. According to the authors, the considerably younger dates (215 and 210 Ma) obtained from sheared metabasic rocks suggest that cooling and mylonitic shearing may be related.

Despite the several studies on the microstructural and geochemical evolution of the shear zone, the *PT* conditions of deformation, i.e., prograde for Brodie (1981) and retrograde for Stünitz (1998), as well as the timing, are still poorly constrained. Furthermore, a detailed geological fieldwork, including the description of all the lithologies and their structural relationships, is lacking. In order to fill this gap, we carried out a geological mapping at 1:2.000 scale and we collected several samples across the shear zone including footwall and hanging wall rocks (Figs. 2, 3). Sampling localities are reported in Fig. 2B and Table S1.

3. Field observations and sampling strategy

Field relationships and macroscopic features are described moving from SE (hanging wall) to NW (footwall) across the shear zone (Figs. 2, 3). Studied outcrops comprise hanging wall and footwall rocks up to a distance of about 0.8 and 2.0 km, respectively, from the shear zone and following increasing metamorphic conditions from amphibolite- to granulite facies (Figs. 2, 3).

The hanging wall (SE) is characterised by upper amphibolite facies rocks with different composition: paragneisses (sample AN11) and minor mafic gneisses (AN10; Figs. 2B, 3A), both showing evidences for partial melting. Migmatitic paragneisses are metatexites with thin (up to ~1 cm) leucosomes occurring as patches or along the folded foliation (Fig. 3B). Melanosome mainly consists of fine-grained biotite, garnet, sillimanite, quartz, and feldspars. Mafic gneisses consist of plagioclase, clinopyroxene and amphibole, and appear as centimetres-meters thick intercalations within the paragneisses (Fig. 3A).

The overall thickness of the Anzola shear zone is around 100 m (Fig. 3A) and is characterized by a well-developed mylonitic fabric steeply dipping (~70°) towards SE (Fig. 2B) and a left-lateral south-westward shear-sense (Rutter et al., 2007). The shear zone is dominated by paragneisses and amphibolites (Fig. 3A). Paragneisses prevail in the SE portion, whereas amphibolites are abundant towards the NW boundary (Fig. 3A). All rock types show compositional layering on a centimetre to decimetre scale (Fig. 3B-G, S1A). Paragneisses are characterized by alternating coarse-grained levels of plagioclase, garnet, biotite and fine-grained layers with more abundant biotite and sillimanite (Fig. 3C-E). Mylonitic fabric is easily recognizable in hand specimen by a well-developed foliation plane defined by the alignment of biotite, and garnet porphyroclasts contrasting quartzo-feldspatic domains (e.g., AN09, AN09C, AN22A; Fig. 3E). Garnet porphyroblasts are distinctly rounded (Fig. 3C-E) in contrast to garnet-bearing rocks outside the shear zone (Fig. 3B). Locally, boudinaged leucocratic rocks characterized by feldspar and greyish porphyroclasts up to 2 cm occur within paragneisses (e.g., AN22B, Fig. 3F; S1C). Amphibolites involved in the shear zone consist of predominant amphibole, clinopyroxene and plagioclase with rare occurrence of garnet (AN01, AN06I-L; Fig. 3G). At the outcrop scale, strain partitioning defines an anastomosing pattern (Fig. 3A). Closer the SE boundary, a low-strain zone occurs passing from protomylonitic (AN22A; Fig. 3C) to a high-strain zone of (ultra)mylonitic paragneisses (AN09, AN09C; Fig. 3D, E). Overall, strain partitioning is limited as compositional bands and foliation are subparallel to each other precluding high rheological contrast which would result in boudinage or similar features (Fig. 3A). Towards the NW boundary, the shear zone is dominated by high-strain rocks mainly made of alternating mylonitic (AN01, Fig. 3G) and ultramylonitic amphibolitic layers (AN06I, L; Fig. 3A). Ultramylonitic layers are defined by very small grain sizes (<0.125 mm) and straight compositional boundaries between plagioclase- and amphibole-rich layers. This contrasts intermediate strain conditions at the SE boundary where rocks are dominated by protomylonitic and mylonitic paragneisses (Fig, 3A). Here, ultramylonitic layers are defined by very small grain sizes (<0.125 mm) and straight compositional boundaries between plagioclase- and amphibole-rich layers.

Towards NW, a lithologically heterogeneous domain separates the shear zone from a up to 800 m thick gabbroic body (Fig. 2B). Here, centimetre thick mafic gneisses, paragneisses and calc-silicates define folded layers (AN06C-G; Figs.2B, 3H, I) with axial planes steeply dipping towards SE and axes towards around E (Figs. 2, 3H), coherent with the main Variscan folding phase (D_2; see Rutter et al., 2007 for further details). Mafic gneisses and paragneisses have the same mineral assemblages as in the shear zone, i.e., mafic gneisses consist of clinopyroxene, plagioclase amphibole and minor garnet, while paragneisses are made of garnet, feldspars, sillimanite, quartz and minor biotite. Calc-silicates consist mainly of clinopyroxene, plagioclase, calcite, and epidote. Large crystals (up to 1-2 cm) of titanite are clearly observable in hand specimen (AN06G; Fig. S1E). These rocks are here documented for the first time (AN06G; Figs.2B, 3H, I). The contact between the folded gneisses and the gabbroic rocks is marked by lenses of calc-silicates (AN06C) and pockets of coarse-grained garnet, amphibole and plagioclase (Fig. 3I, L). All these rocks close to the contact with gabbroic rocks are here documented for the first time (i.e., AN06C-G; Figs.2B, 3H, I).

Proceeding to NW within the footwall of the shear zone, mafic rocks belonging to a gabbroic body occur (Gabbro of Anzola as defined by Cavallo et al., 2004; samples AN07B, AN08B, AN08C, AN20, AN21; Figs. 2, 3A). The gabbroic rocks are generally coarse grained and made of plagioclase and clinopyroxene and minor amphibole (sample AN07B/08B). Locally thin discontinuous layers or lenses richer in amphibole or plagioclase are recognisable (Fig. 3L). The gabbro within 5 m of the folded sequence shows a weak magmatic sub-vertical foliation dipping towards SE (Fig. 2B; 3L) which becomes indistinct towards the gabbro sensu stricto (Fig. 3A). The latter is defined by its isotropic texture characteristic for the main gabbro body. Locally, pegmatites and leucocratic dikes of plagioclase and pyroxene crosscut the gabbro (Fig. 3M).

Further to the NW, the gabbro is in contact with the felsic granulites (sample AN18, AN19; Fig. 2B), which become the dominant rock type showing the typical high-grade paragenesis consisting of abundant garnet, quartz, feldspars, sillimanite and rare of absent biotite. The contact between felsic granulites and gabbro is not well-exposed, however, a weak foliation is locally recognisable (Fig. 2B).

Felsic granulites are characterized by brownish needles of rutile easily recognisable on the fresh surfaces of the samples. The texture is prevalently isotropic, with a weak foliation steeply dipping towards SE (Fig. 2B) still concordant with the regional attitude.

To characterize the lithological, microstructural, petrological and geochemical variations across the shear zone, we collected samples across the wall rocks and the mylonitic/ultramylonitic shear zone (Figs. 2, 3, S1). A detailed list of samples is reported in Supplementary material (Table S1). Samples (around 40 and initial label AN = Anzola) were subdivided in (Figs. 2, 3): i) proto- and ultra-mylonitic rocks from the shear zone (sample labelled AN01A-I, AN06I-L, AN09, AN09C, AN22A, AN22B; Figs. 2, 3) and metamorphic rocks representative of the wall rocks at the footwall and hanging wall, i.e., paragneisses, mafic gneisses and felsic granulites (samples labelled AN10, AN11, AN06C-H, AN18, AN19; Figs. 2B, S1). Samples from the gabbro of Anzola are considered as a subcategory of the wall rocks and are identified as gabbroic rocks *(*samples labelled AN07, AN08B, AN08C; AN20, AN21). Mineral abbreviations are after Whitney and Evans (2010).

4. Petrography and microstructure

4.1 *Wall rocks*

4.1.1 Hanging wall rocks

Migmatitic paragneisses (sample AN11) are characterized by a layered texture (stromatic migmatites following Sawyer, 2008) consisting of alternating thin (up to 5 cm-sized) leucosomes and thicker (up to 10 cm) biotite-rich mesosomes (Fig. 4A). The boundaries between stromatic leucosomes and melanosomes are generally marked by a local increase of biotite towards the mesosomes. Leucosomes consist of large garnet grains including clusters of quartz, abundant rutile, and graphite (<1 cm; Fig. 4A). The mesosome is medium- to coarse-grained, composed of biotite, garnet, prismatic, aligned sillimanite, K-feldspar, plagioclase, and quartz (Fig. 4A; Table S1). Clear evidence for partial melting is especially in the form of resorbed biotite and sillimanite associated with films of K-feldspar and quartz. Accessory phases are rutile, ilmenite, graphite, apatite, zircon, and monazite. Rutile is abundant and associated with ilmenite.

Rocks characterized by a grano-nematoblastic texture and dominant mineral assemblage of Amph + Cpx + Pl + Scp (Table S1) are intercalated with the migmatitic paragneisses (sample AN10; Fig. 4B). The average grain size is around 0.5 mm. These samples are strongly banded alternating dark layers of subhedral amphibole and plagioclase and whitish layers of anhedral plagioclase, subhedral clinopyroxene and subhedral scapolite (Fig. 4B; Table S1). From these observations, hereafter, we identified these samples as scapolite-bearing gneisses. Accessory minerals are ilmenite, apatite, zircon and titanite. The latter is the most abundant and is present in subhedral or rounded grains with grain size up to 250 µm. Rare and altered epidote is present as evidence of retrograde metamorphic reactions.

4.1.2 Footwall rocks

Mafic gneisses (sample AN06F) show a grano-nematoblastic texture and consist of two levels made of Cpx + Pl ± Grt ± Amph and of Amph + Cpx + Pl + Grt, respectively (Fig. 4C). In the first layer, amphibole occurs only as interstitial grains, whereas garnet is rare. Conversely, the amphibole-rich portion is characterized by major modal abundance of garnet occurring as both poikiloblasts enclosing plagioclase, clinopyroxene and amphibole, or as coronitic texture surrounding plagioclase (Fig. 4D) or fine-grained mineral aggregates. Abundant amphibole (up to 40 vol.%) occurs as both interstitial mineral phases and as large grains including plagioclase and clinopyroxene. The fine-grained alteration shows a pale yellowish colour and composition comparable to chlorite (Fig. 4C). Relicts of brown amphibole are visible suggesting that the fine-grained aggregates represent an alteration product of amphibole. A thin layer of chlorite has been observed also between garnet and plagioclase in the coronitic texture. Amphibole contains several tiny oxide exsolutions along the cleavages. Ilmenite is the most abundant accessory mineral and locally is partially replaced by rutile and titanite. As other accessory minerals, apatite occurs in both layers, whereas titanite is present only in the Amph-poor portion.

Paragneisses (sample AN06D) show a weak foliated texture characterized by large garnet porphyroclasts surrounded by a matrix mainly composed of quartz, plagioclase and K-feldspar. The mineral assemblage consists of Pl + Kfs + Qz + Grt + Sil + Bt (Fig. 4E; Table S1). Garnet occurs mainly as large, fractured (up to 2 mm) grains with a high density of inclusions of quartz, sillimanite, biotite, plagioclase, and accessory minerals such as rutile and zircon. Prismatic subhedral grains of sillimanite are present as both isolated grains along the foliation or aggregates surrounded by tiny film or atollar garnet, plagioclase, and K-feldspar (Fig.

4E). Biotite is rare, commonly chloritized and associated with garnet. White mica locally replaces the sillimanite. Rutile, ilmenite, graphite, apatite, zircon, and monazite (up to 100 µm) are accessory phases. Rutile reaches grain size up to 500 µm and is the most abundant among the accessory mineral especially observable as inclusion in garnet or along the matrix.

Calc-silicates are coarse-grained (mm-cm sized) and consists of plagioclase, epidote and titanite (sample AN06C, AN06G; Fig. 4F). Sigmoidal/lozenge shaped crystals of titanite of about 1x2 mm preferentially are dispersed in the matrix or found associated with the clinopyroxene, as well as radial aggregates of epidote up to 2 mm in size (AN06G; Table S1). Apatite is abundant and form aggregates. At the contact with gabbroic rocks, calc-silicate (AN06C) are characterized by granoblastic texture and consist mainly of Cpx + Cal + Pl ± Scp and minor quartz and K-felspar (Fig. 4F; Table S1). Titanite, apatite, zircon and sulphides are common accessory minerals. Both samples show evidence for retrograde replacement reactions that have promoted the formation of chlorite, white mica, epidote and albite at the expense of the primary assemblages. In the sample AN06G, especially, the primary mineral assemblage is almost completely overprinted.

4.1.2.1 Gabbroic rocks

Gabbroic rocks (samples AN07B, AN08B, AN08C, AN20, AN21A) show a mainly coarse-grained granoblastic and nearly isotropic texture (Fig. 4G; Table S1, Fig. S1B). It consists of mm- to cm- sized euhedral-subhedral grains of plagioclase with polysynthetic twinning, euhedral-subhedral clinopyroxene and subhedral amphibole (Fig. 4G). Predominantly accessory phases are ilmenite and apatite, mainly included in pyroxene. The inner portion of the body (i.e., AN08C, AN20) displays an isotropic texture with equigranular crystals (in the order of mm-µm size). Towards the shear zone, the gabbro (AN21, AN08C) shows well-defined foliation marked by amphibole aligned almost parallel to the schistosity of the shear zone rocks (Fig. 3A). This observation is in contrast with those made by Stünitz, (1998), which observed a discordant attitude between the vertical foliation of the shear zone and the metamorphic layering of the host rocks (see figure 3 of the cited work). Closer to the shear zone, the gabbro is coarse grained (cm-sized) with large clinopyroxene grains (i.e., AN07B; Fig. S1E) and locally with big garnet grains (up to 2 cm in size; AN07B/AN08B; Fig. S1E). The gabbroic body is locally crosscut by leucocratic dykes showing coarse-grained granoblastic texture of plagioclase and clinopyroxene (AN21B; Fig. 3L).

4.1.2.2 Felsic granulites

Felsic granulites (AN18, AN19) consist of coarse-grained garnet (0.5–2 mm), plagioclase, sillimanite, K-feldspar, quartz, and subordinated biotite (Fig. 4H; Table S1). The texture is mainly granoblastic and no evidence of deformation were observed. Garnet occurs as rounded/elongated porphyroblasts with variable concentrations of inclusions of plagioclase, quartz, rutile and zircon (Fig. 4H). Sillimanite shows a variable modal abundance among the samples (more abundant in granulite AN18) occurring mainly in prismatic subhedral grains and locally as inclusions in garnet. Plagioclase is the dominant feldspar; K-feldspar locally occurs in association with garnet and biotite (Fig. 4H). Biotite is rare and is mostly chloritized at the rim. Retrograde white mica is observable around garnet grains. Accessory minerals are apatite, zircon, monazite, rutile and graphite, which is less abundant than in samples from the upper amphibolite and transition zones (i.e., AN11, AN09). Rutile is the most abundant among the accessory minerals occurring as rounded grains and often in association with ilmenite.

4.2 *Shear zone rocks*

4.2.1 Amphibolites

A detailed petrographic characterization of mylonitic amphibolites was carried out on several thin sections, representative of the main grain size, mineral abundance and textural variations (AN01A-I; Fig. 5). This sample group corresponds to the mylonitic rocks described in the previous studies (Brodie., 1981; Brodie et al., 1989; Altenberger, 1997; Stünitz, 1998) who interpreted these as derived from the adjacent gabbroic body.

Mylonitic amphibolites consist of alternating bands of plagioclase- and amphibole/clinopyroxene-rich layers (Fig. S1A). Mylonitic texture is characterized by core and mantle structure where porphyroclasts (diameter of 0.5-1.0 mm) of plagioclase, amphibole or clinopyroxene and, rarely, garnet, are surrounded by a fine-grained recrystallized matrix composed of alternations of monophasic layers (either plagioclase, amphibole or clinopyroxene) with mixtures of plagioclase + clinopyroxene (± garnet and titanite) or plagioclase + amphibole (Fig. 5A-E). Modal abundances of amphibole (28-42%), plagioclase (28-36%) and clinopyroxene (15-25%), strongly varies along layers (Fig. S1; Table S1).

In Cpx/Pl-rich layers of the amphibolites (i.e., AN01D, Fig. S1A, D), clinopyroxene is preserved mainly as porphyroclasts (CpxI; Fig. 5G) showing tails of recrystallized clinopyroxene (CpxII; Fig. 5G) and/or amphibole (AmphII; Fig. 5C). Porphyroclasts are locally replaced by a greenish amphibole, which also occurs in fractures within the porphyroclasts or as equant grains adjacent to clinopyroxene grains (e.g., thin section AN01D, Fig. 5C, G). Clinopyroxene porphyroclasts can include inclusions of plagioclase, amphibole, calcite and ilmenite.

Amphibole porphyroclasts show intracrystalline deformation as evidenced by strong undulose extinction and deformation bands. Their long axes are orientated parallel to the foliation and are characterized by tails of recrystallized amphibole grains, which usually are beard-like overgrowths on the porphyroclasts (Fig. 5B). Brown-amphibole occurs as porphyroclasts, whereas green-amphibole (AmphI) predominates in the fine-grained recrystallized matrix (AmphII) (Fig. 5C, E). Amphibole porphyroclasts can include apatite, plagioclase, clinopyroxene and zircon.

Rarely, large porphyroclasts of sub-rounded pinkish garnet were observed along the clinopyroxene/plagioclase-rich layers. These garnet grains locally show a poikilitic texture with clinopyroxene and plagioclase (e.g., thin sections AN01, D, H, I; Fig. 5E).

Most of the plagioclase is completely recrystallized as fine-grained matrix (PlI) but it locally occurs as porphyroclasts or ribbon (PlI) (Fig. 5A-G). Rounded or ellipsoidal plagioclase porphyroclasts show glide-controlled deformation, such as (bent) twins and undulatory extinction (Fig. 5D).

Locally, calcite is present in the strain shadows as termination of the porphyroclasts or between amphibole porphyroclasts (e.g., AN01B; Fig. 5H). Titanite, ilmenite, apatite and zircon are common accessories minerals (Fig. 5C, F). Ilmenite is quite abundant in mylonitic samples occurring as interstitial grain with low dihedral angles and irregular shapes along the foliation (Fig. S2A, B). Elongated or sigmoidal shaped crystals of titanite preferentially occur within the Cpx/Pl-rich layers but also as inclusions in amphibole porphyroclasts (e.g., AN01A, B, D; Fig. 5C). In some cases, titanite occurs in big crystals with up to 500 µm wide x 1 mm long showing the typical lozenge shape and double set of twinning (Fig. 5F). Usually, when zircon is very common, titanite is less abundant (e.g., AN01C). Late alteration product such as calcite, chlorite, epidote, and mica are concentrated in fractures and cross-cutting veins.

Ultramylonitic amphibolites (i.e., AN06L; clasts <10% and matrix grain size <0.125 mm) have a mineral assemblage mainly constituted by porphyroclasts of amphibole and plagioclase (up to 1 mm large), surrounded by a finer matrix (average of 100 µm) made by recrystallized amphibole, plagioclase and clinopyroxene (Fig. 6A; Table S1). Biotite porphyroclasts are mainly aligned along the main foliation. Clinopyroxene is mainly preserved as recrystallized grains (~150 µm) in the matrix (Fig. 6A). The rare garnet occurs in poikilitic texture with plagioclase and ilmenite and has a size of 250 to 500 µm. Quartz was locally observed as cluster surrounded by biotite and matrix forming minerals. Rounded apatite and anhedral ilmenite are abundant as accessory minerals (size up to 250 µm), several grains of zircon with size up to 100 µm occur. Chlorite, epidote and rare crystals of white mica are present as alteration products in late fractures and/or veins.

4.2.2 Paragneisses

Sheared paragneisses (i.e., AN09, AN09C, AN22A; Fig. S1) have a mineral assemblage of Grt + Sil + Kfs + Qz + Pl ± Bt (Fig. 6; Table S1). Mylonitic domains are characterized by large garnet porphyroclasts up to 5 mm in size wrapped by a matrix of sillimanite, feldspar, and recrystallized quartz (Fig. 6B, C). Randomly oriented quartz, plagioclase, biotite, and rutile are observed as inclusions in garnet porphyroclasts (Fig. 6B-E). Syn-kinematic biotite (BtII) occurs along the main foliation, in strain shadows around garnet and feldspar porphyroclasts (Fig. 6C, D). Locally, relicts of biotite (BtI) are present in low strain domains with anhedral shape and included in garnet porphyroclasts (Fig. 6C, D). Sillimanite is observable both as big porphyroclasts (up to 1-2 mm long) and along the mylonitic foliation with a smaller grain size (up to 250 µm; Fig. 6B). Feldspars occur as porphyroclasts up to 500 µm in size and syn-kinematic mineral along the main foliation showing evidence of ductile deformation such as undulose extinction and deformation lamellae (Fig. 6B-F). Quartz is mostly present within the matrix showing irregular grain boundaries with sometimes low dihedral angles (Fig. 6B, C; S2C, D). Internally, some undulose extinction is seen (Fig. 6B, C). Rutile (with average grain size of 150 µm), ilmenite, monazite (grain size up to 500 µm) and zircon are abundant as accessory minerals and occur in the fine-grained matrix (Table S1).

Ultramylonitic paragneisses occur as thin, mm-cm sized layers (sample AN09C) among the mylonitic domains. The mineral assemblage consists of rounded garnet, sillimanite porphyroclasts embedded in a recrystallized matrix of plagioclase and biotite (Fig. 6C-D). In this domain, sillimanite porphyroclasts are surrounded by secondary white mica (Fig. 6D).

Locally, paragneisses contain boudinaged leucocratic layers (AN22B) showing a mylonitic fabric characterized by K-feldspar, corundum and minor Bt + Grt + Pl + Spl (Fig. 6F; Table S1). K-feldspar porphyroclasts are the main constituent of the leucocratic layer and are characterized by plagioclase exsolution. The greyish corundum porphyroclasts are up to 2 cm in size. Garnet occurs as porphyroclast (grain size up to 5 mm) including biotite, plagioclase, and zircon. Green spinel occurs as a minor constituent of the corundum-bearing leucocratic sample (AN22B), it is mostly associated with garnet, corundum and biotite (Fig. 6E). Locally a greenish spinel surrounded by biotite is recognisable also within the mylonitic paragneiss (AN22A; Fig. 6F). Biotite showing kinking occurs along the foliation as well as are included within garnet and spinel. Plagioclase is mainly hosted within strain shadows around garnet. Several grains of monazite up to 500 µm in size is observable as accessory mineral.

5. Bulk and Mineral chemistry across the shear zone

5.1 Bulk Rock

Bulk rock composition was determined for representative samples across the Anzola shear zone rocks. A full description of the analytical methods is presented in Appendix, whereas results and related tables are available in Figure 7 and in the Supplementary material (Table S2; Figure S3).

Both major and trace elements bulk rocks composition highlight a strong lithological variability across the Anzola transect.

SiO_2 contents range between 46 to 49 wt.% for mafic rocks of the hanging wall, shear zone and footwall (Scp-bearing gneiss, mylonitic amphibolites, gabbros, respectively), whereas it is between 48-65 wt.% for calc-silicates from the footwall and paragneisses from hanging wall, shear zone and footwall (migmatitic paragneiss, mylonitic paragneisses, felsic granulites, respectively; Fig. 7A).

Paragneisses and felsic granulites are characterised by Al_2O_3 in the range of 21.5 and 25.0 wt.%, except for one felsic granulite (AN19) that contain lower Al_2O_3 content around 16.4 wt.%, probably due to the scarce abundance of sillimanite. Mafic rocks and calc-silicates show even lower values in Al_2O_3 ranging from 11.0 to 16.4 wt.% (Fig. 7A).

TiO_2 contents span mainly between 1.0 to 1.2 wt.% and reach the highest values within mylonitic amphibolites and rutile-rich granulite, maybe due to the abundance of titanium-rich accessory minerals such as titanite, ilmenite and rutile (Fig. S3A).

The MgO versus CaO diagram was plotted to emphasise the lithological heterogeneity along the Anzola profile (Fig. 7B). Paragneisses and felsic granulites, both undeformed and mylonitic samples, show low MgO and CaO values ranging between 2.0-4.0 wt.% and 0.5-5.0 wt.%., respectively. Conversely, mafic rocks have MgO and CaO values that do not exceed 4 wt.% and 10 wt.% in that order. Specifically, mylonitic amphibolites have values of MgO ~6.0 wt.% and CaO ~12.5 wt.%, lower with respect to the Scp-bearing gneisses from the hanging wall (MgO = 6.0 wt.% and CaO = 16 wt.%) and gabbroic rocks (MgO ~8.5 wt.% and CaO ~15.75 wt.%). Calc-silicate from the footwall stands out with respect to the other lithologies having the higher CaO content of ~20 wt.% and lower MgO of ~3.0 wt.%.

Bulk trace elements that allow to distinguish the different rock types (i.e., mafic from the metasedimentary rocks) are in particular the REE elements and Cr versus Co (Fig. 7C, D; S3B; Table S2).

The concentrations in High-REE (HREE) have a similar flat trend for all the studied samples, but generally metasedimentary rocks have REE concentration 40 times higher than CI chondrite, whereas mafic rocks show lower concentration below CI chondrite (Fig. 7C, D). The highest and lower values are reached for felsic granulite (AN18) and Scp-bearing gneiss (AN10), respectively. Light-REE (LREE) show more significant differences among lithologies (Fig. 7C, D). Overall, the mafic rocks have LREE concentrations lower than metasedimentary rocks (Light-REE>100CI). Among the mafic protoliths, the highest values come from the inner portion of the gabbroic body (AN20), whereas the hanging wall Scp-bearing gneiss is the lithology that shows a significant depletion in LREE.

Paragneisses, felsic granulite and calc-silicate show negative anomaly in Eu (except that for felsic granulite AN18), with lower REE values for calc-silicate (AN06D) and felsic granulite (AN18; Fig. 7C).

Calc-silicate mimes the general patterns of paragneisses and felsic granulites with lower concentration in LREE. HREE concentrations have a flat pattern comparable with those of the mafic rocks (Fig. 7C, D).

5.2 Mineral chemistry: Major elements

In this section, we reported the results of mineral composition from mafic rocks (i.e., amphibole, clinopyroxene, feldspar, garnet), calc-silicates (i.e., clinopyroxene, feldspar), paragneisses and felsic granulite (i.e., feldspar, garnet, biotite, spinel). Details on the methods are reported in Appendix; results are reported in Supplementary material (Table S3) and Figures 8 and 9. Links to the petrographic relationships are highlighted through reference to figures 4-6.

5.2.1 Amphibole

Amphibole is mostly pargasite, (Na + K) > 0.50 a.p.f.u. and Si < 6.5 a.p.f.u., with a few edenite and hornblende grains analysed within mylonitic amphibolites (Fig. 8A; cf. Fig. 5B). In the Amph-rich mylonites (i.e., AN01A, AN01C), the amphibole composition tendentially changes from pargasite in the brown porphyroclasts (AmphI) to green edenite in the syn-kinematics grains (AmphII; Fig. 8A; cf. Fig. 5H).

5.2.2 Clinopyroxene

In the studied samples, clinopyroxene is prevalently diopside (45 < Wo < 50 mol%, Fe^{2+} < 0.5 a.p.f.u.; Fig. 8B). Exceptions are the gabbroic rocks (AN08B, C, AN07) where clinopyroxene has mostly an augitic composition (Wo < 45 mol%; Fig. 8B; cf. 4G) and in the Cpx/Pl-rich mylonitic amphibolites (AN01D; cf. 5G) where has hedenbergitic composition (45 < Wo < 50 mol%, Fe^{2+} > 0.5 a.p.f.u.), as well as clinopyroxene from the calc-silicates (i.e., AN06C; Fig. 8B, cf. Fig. 4F).

5.2.3 Plagioclase

Plagioclase from the mafic rocks, ranges in composition mainly from andesine to labradorite (Fig. 8C). Anorthite content up to 90-100 mol% were obtained from the calc-silicate sample (AN06C; Fig. 8C).

Plagioclase in the gabbro (AN08B, C, AN07) has labradorite composition with anorthite content between 50-60 mol%. Plagioclase from mylonitic amphibolites shows lower anorthite contents (<50 mol%), particularly in the Amph-rich mylonites (30-40 mol%) and, generally for all syn-kinematic grains (PlII). Few plagioclase grains found as inclusion in amphibole (i.e., AN01D) are oligoclase, with an anorthite content around 25-35 mol%.

Paragneisses and felsic granulite contain both plagioclase and K-feldspar, except for the felsic granulite which present only plagioclase (AN18; Fig. 9A; Table S3). The latter ranges in composition from labradorite to andesine: the highest anorthite contents were obtained for the low deformed migmatite (AN11, average An = 64 mol%). Mylonitic paragneisses are characterized by anorthite contents ranging from 40 to 55 mol%, with the highest values related to plagioclase included in garnet (55 mol%), whereas felsic rocks from the footwall (AN06D, AN18, AN19) show mainly andesine composition (30 mol% < An < 40 mol%). K-feldspar has orthoclase content mainly in the range of 80-95 mol% (Fig. 9A). The mylonitic paragneiss (AN09) is characterised by lower orthoclase contents (75 mol%).

5.2.4 Garnet

As shown in the ternary diagram (Fig. 9B), garnet is significantly variable among sample types. All samples fall close to the almandine + spessartine vertex (> 50%). The grossular content of garnet allows to distinct three main groups associated with different lithotypes: i) garnet from Cpx/Pl-rich mylonitic layers from amphibolites shows the highest grossular content (~35-40%); ii) intermediate values (~20-25%) characterize garnet from the Amph-rich mylonitic amphibolites and gabbroic rocks; iii) paragneisses host garnet with the lowest grossular content (~5%). On the other hand, starting from the Alm + Sps vertex, the increase in the pyrope component (15%→40%) is observed from corundum-bearing leucocratic rocks (AN22B) across mylonitic paragneisses (AN22A, AN09C3, AN09) and from hanging wall paragneiss, migmatite (AN11), to the footwall paragneiss (AN06D) and felsic granulite (AN18, AN19). Moreover, pyrope content minor or greater than 35% (dashed grey line in Fig. 9B) marks samples from either the footwall or hanging wall, respectively. The increase in pyrope content over 35% also correlates with the switch from amphibolite to granulite-facies (Fig. 9B). Interestingly, this change of metamorphic facies is also marked by a slightly decreases of the grossular content within granulitic samples.

5.2.5 Biotite

Overall, biotite from paragneisses and felsic granulite has Fe# ($Fe^{2+}/Fe^{2+}+Mg$) that ranges between 0.25 and 0.60, with any specific distinction on the base of the metamorphic grade or deformation (i.e., BtI, BtII; Fig. 9C, S3C; Table S3). High Fe# values are recorded by corundum-bearing leucocratic rock (AN22B) and chloritized felsic granulite (AN19), whereas low Fe# values are shown by mylonitic paragneiss (AN09) and felsic granulite (AN18). Overlapping of Fe# values are present only by two mylonitic paragneisses (AN09C, AN22A; Fig. 9C). Al contents range mainly between 1.2-1.6 a.p.f.u. for all rock types except for biotite from corundum-bearing rock where Al is higher (from about 1.7 to 1.9 a.p.f.u.). Ti contents range between 0.15 to 0.45 a.p.f.u. with an average value of 0.3 a.p.f.u., low values for felsic granulite (AN18), corundum-bearing rock (AN22B) and scattered biotite from ultramylonitic paragneiss (AN09C3; Fig. S3C).

5.2.6 Spinel

Spinel occurs as minor component in mylonitic paragneiss (AN22A) and corundum-bearing leucocratic rock (AN22B; Table S1). Spinel grains occurring in the two samples show the same petrographically features. Nevertheless, the spinel crystals in mylonitic paragneiss are hercynite/gahnite in composition with ZnO content around 12 wt.%, whereas in corundum-bearing sample spinel is hercynite with ZnO content around 0.9 wt.% (Table S3).

5.3 Mineral chemistry: Trace Elements

Trace elements concentrations were determined for main mineral phases, i.e., clinopyroxene and amphibole, for representative mafic rocks and calc-silicates: Scp-bearing gneiss (AN10) from the hanging wall, mylonitic amphibolites (AN01A, AN01D), calc-silicate (AN06C) and gabbros (AN07B, AN08B, AN08C) from the footwall (Fig. 3-6; Table S1). The reason why we focused only on these samples is to verify if the Anzola shear zone developed partially at the expense of the gabbroic body (e.g., Brodie, 1981; Altenberger, 1997; Stünitz, 1998; Rutter et al., 2007). In particular, we focused on the Rare Earth Element (REE) concentrations with the aim to diagnose the protoliths of mylonitic amphibolites. Results are reported in Supplementary material (Table S4). In Figure 10 we plotted clinopyroxenes and amphiboles REE patterns for the studied samples normalised to chondrites values according to McDonough and Sun, (1995).

Clinopyroxene REE patterns show concentrations between 1- and 10-times CI for all samples except for gabbroic rocks that have higher concentration (>10 times CI; Fig. 10A). Mylonitic amphibolites have similar clinopyroxene REE behaviour despite the different compositional layering (Amph-rich, Cpx-Pl rich). Scp-bearing gneiss from the hanging wall is strongly depleted in LREE (around 1 time CI). Mylonitic amphibolites and calc-silicates share a similar relatively flat REE patterns with a slight increase in the HREE concentrations. Clinopyroxene REE patterns for gabbroic samples were distinguished on the base of the grain size (e.g., Fig. S1E – Gabbroic rocks AN07B): coarse-grained (CpxI) and fine-grained (CpxII). Coarse grained parts share a similar pattern to the fine-grained one and negative anomaly in Eu, whereas fine-grained parts show a more flatted trend.

Amphibole REE patterns show comparable behaviour between 10- and 100-times CI for all samples, except for the REE pattern of porphyroclast (AmphI) in Amph-rich mylonitic amphibolites and the Scp-bearing gneiss from the hanging wall that show strong depletion in LREE (Fig. 10B). Amphibole from Amph-rich mylonites shows similar REE patterns but with significantly different concentrations where porphyroclasts (AmphI) are overall richer in REE up to about 3 times the normalized REE values from syn-kinematic grains (AmphII; Fig. 10B; c.f., Fig. 5B). An apparent negative Eu anomaly was observed for the amphibole from the Amph-rich mylonites, whereas it is only slightly evident for the other samples. Amphibole REE patterns for gabbroic samples were distinguished, as clinopyroxenes, on the base of the sample grain size (e.g., Fig. S1E – Gabbroic rocks AN07B): coarse-grained (AmphI) and fine-grained (AmphII).

6. Geothermobarometry

Pressure (*P*) estimates in the range between 0.7-0.8 GPa were determined for the aluminosilicate bearing mylonitic rocks, namely paragneisses and felsic granulites, using the Garnet-Aluminosilicate-Plagioclase geobarometer (GASP; Holdaway, 2001) considering sillimanite as the stable aluminosilicate. *P* conditions of the IVZ have been presented by several authors (e.g., Zingg, 1980; Henk et al., 1997; Barboza and Bergantz, 2000; Luvizotto and Zack, 2009; Redler et al., 2013; Kunz et al., 2014) giving an average value of 0.7 GPa for granulite to amphibolite facies boundary. The mineral assemblages of mafic rocks are not suitable for pressure estimates by conventional geobarometers.

Temperature (*T*) conditions were determined by combining the following geothermometers: Zr-in-Rutile (Kohn, 2020), Ti-in-Amphibole (Liao et al., 2021) and Garnet-Biotite (Grt-Bt) exchange (Holdaway, 2000). Results are reported in Figure 11 and Supplementary material (Table S5, S6). Since the Zr-in-Rutile and Ti-in-Amphibole geothermometers are pressure dependent, we set the pressure at 0.7 GPa in accordance with our and literature estimates. The applied geothermometers allow to assess both pre-shear and syn-shear conditions. The pre-shear conditions have been estimated from samples avoiding mylonitic deformation through the Zr-in-Rutile geothermometer and the Grt-Bt in several lithologies ranging from upper amphibolite to granulite facies conditions, while the Ti-in-Amphibole geothermometer was used only in a Scp-bearing gneiss from the hanging wall (Fig. 11). Syn-shear temperature conditions were constrained on mylonitic amphibolites (Ti-in-Amphibole) and paragneisses (Grt-Bt). The Ti-in-Amphibole was used for the different generations of amphiboles, i.e., the porphyroclasts (AmphI) and the syn-kinematic recrystallised gains (AmphII). For the Grt-Bt geothermometer we paired the chemical analyses of syn-kinematic biotite (BtII) with garnet rims (Fig. 6C, F). Constraints on the pre-shear conditions from the Zr-in Rutile geothermometer define a slightly decreasing trend from granulite facies (816±27 °C) to upper amphibolite facies rocks (716±65 °C, Fig. 11). Our estimates agree with data from similar rocks in the studied area (Luvizotto & Zack, 2009). The Grt-Bt geothermometer yielded higher temperatures in the footwall (up to 900±20 °C for the felsic granulites) with respect to the migmatites in the hanging wall (600±40 °C), partially overlapping estimates from the Zr-in-Rutile (Fig. 11). The Ti-in-Amphibole geothermometer provided temperature of 818±28 °C in agreement with other geothermometers.

Syn-kinematic conditions from mylonitic samples were defined by Grt-Bt and Ti-in-Amphibole geothermometers providing temperatures between ~550 and ~700 °C and ~750 and ~900 °C, respectively. As regard the Ti-in-Amphibole, the *T* values gradually decrease as function of the textural positions. Amphibole included in clinopyroxene and porphyroclasts display systematically high temperature (830±48 -

890±34 °C), whereas recrystallized grains show decreasing temperature from 837±22 to 786±50 °C (Fig. 11; c.f., Fig. 5).

7. Discussion

7.1 Pre- and syn-deformational PT conditions

Mineral assemblages from the footwall rocks (Grt + Sill + Pl + Kfs + Qz ± Bt; Fig. 11) indicate granulite facies conditions at *T* of ca. 800-900 °C compatible with our temperature estimates (Zr-in-Rutile and Grt-Bt geothermometers) and to the literature data (Zr-in-Rutile from felsic granulites; Luvizotto & Zack, 2009; Fig. 11). Hanging wall rocks are characterized by: i) the stability of amphibole and the lack of orthopyroxene in mafic rocks (Fig. 4, 5); ii) the occurrence of sillimanite and minor biotite in paragneisses (Fig. 6); iii) the lack of muscovite (except for retrograde white mica; Fig. 6D). These observations indicate conditions below the Opx-in isograd and above the Kfs + Sill-in isograd suggesting that the pre-shear conditions of the studied samples resulted between upper amphibolite to granulite facies conditions, in agreement with thermometric results (~550-800 °C: this work and literature data; Fig. 11) and metamorphic field gradient studies (e.g., Zingg, 1980; Fig. 2A).

Temperature estimates from mylonitic rocks revealed interesting features. The Ti-in-Amphibole geothermometer applied to recrystallized amphibole grains of the mylonitic amphibolites provides high *T* conditions for deformation (837±22 to 786±50 °C; Fig. 11). These estimates agree with the occurrence of interstitial melt during the earliest stage of deformation as documented by interstitial ilmenite, with irregular, low dihedral boundaries and with elongate grain shapes along grain boundaries (Fig. S2A, B; i.e., AN01D, AN01E; e.g., Ghatak et al., 2022). Similar features were even shown by quartz in mylonitic paragneisses (Fig. S2C, D; e.g., Piazolo et al., 2020). However, the *T* obtained from mylonitic paragneisses are systematically lower with respect to *T* estimated by Ti-in-Amphibole geothermometer for recrystallized grains (Fig. 11). This difference would indicate that the two geothermometers recorded different stages of the deformation, in agreement with the observations of local occurrence of retrograde minerals within the matrix of mylonitic paragneisses. Here, we observed sillimanite porphyroclasts always surrounded by white mica, especially within the ultramylonitic portions (Fig. 6D).

T estimates for mylonitic amphibolites were determined in previous studies on the base of microstructural and chemical changes in amphiboles (i.e., variation of Mg# from porphyroclasts to recrystallized grains) within and outside the mylonitic bands (Brodie, 1981; Stünitz, 1998). Brodie (1981) reported increasing *T* in mylonitic amphibolites during prograde regional metamorphism from low- to high-grade amphibolite facies (between 500-700 °C). Conversely, Stünitz (1998) stated that the deformation of amphibolites occurred during a retrograde *P-T* path under amphibolite facies conditions at approximately 550 to 650 °C and confining pressures of probably less than 0.8 GPa. Literature data are in perfect agreement with our *T* estimates based on Grt-Bt geothermometry, resulting to lower *T* than determined by Ti-in-Amphibole geothermometry (up to ca. 200 °C). Based on our results, we suggest that the shear zone was active from granulite to amphibolite facies conditions (Fig. 11). Lithologies recorded different stages of deformation starting at high *T* conditions, in presence of melt and in a diffuse mode, as still recognisable within mylonitic amphibolites, and continued under lower *T* conditions with a progressive strain localization as documented by the (ultra-)mylonitic paragneisses (Fig. 6D).

7.2 Protoliths within the Anzola shear zone are compositionally heterogeneous

New geological and structural mapping, petrographic and geochemical analyses reported in this work document that the Anzola Shear zone is characterized by significant lithological heterogeneity. Mylonites and ultramylonitic bands developed within two main lithologies, namely the paragneisses and amphibolites (Fig. 3A). Minor lithologies such as corundum-bearing felsic rocks and calc-silicates are interlayered within paragneisses and amphibolites, respectively (Fig. 3). Such lithological variations and the relative abundances among rock types reflect the rock associations of both the footwall and hanging wall. The latter is made up of upper amphibolite facies rocks with different composition: paragneisses and minor mafic gneisses, both showing evidence for partial melting (Fig. 3A, B). The footwall is characterized by centimetre thick folded layers of mafic gneisses, paragneisses and calc-silicates (Fig. 3A, G). Petrographic observations, bulk rock (Fig. 7) and mineral chemistry (Figs. 8, 9) indicate that the mylonitic and ultramylonitic paragneisses share more similarities with hanging wall lithologies with respect to footwall ones (Figs. 3, 7, 9). Although the lithologies between gabbroic rocks and the shear zone at the footwall are similar to the hanging wall rocks they experienced textural and chemical modifications due the intrusion of the gabbroic body.

As it concerns the protoliths of the mylonitic amphibolites we exclude an apparent involvement of gabbroic rocks as sustained by previous references (e.g., Brodie, 1981; Altenberger, 1997; Stünitz, 1998). Firstly, the gabbro is compositionally homogeneous, and it does not show apparent anisotropies except for a weak foliation at the contact with host rocks (Fig. 3I). The granoblastic and isotropic texture of the gabbroic samples is incompatible with the apparent compositional banding, i.e., Amph-rich and Cpx/Pl-rich layers, characterizing the mylonitic amphibolites (Fig. 3G, S1A, B). Moreover, these latter are generally richer in amphibole, which occurs only as thin layers or lenses within the gabbroic rocks (Fig. 3G, I). Further evidence for a different origin of mylonitic amphibolites is that they contain also abundant zircon and titanite grains, locally up to hundred microns in size that were never observed within the mafic body (Figs. 4G, 5).

Furthermore, mineral compositions, and in particular trace elements, provide fundamental information in order to shed light on different rock types involved by shearing (Figs. 7, 8). Plagioclase from gabbroic rocks have higher anorthitic contents (i.e., labradorite) with respect to plagioclase porphyroclasts analysed within the mylonitic amphibolites (i.e., andesine; Fig. 8C). Analogously, clinopyroxene from gabbroic samples are typical augite, whereas Amph-rich mylonites and Scp-bearing gneiss from the hanging wall contain diopside (Fig. 8B). Besides that, clinopyroxene porphyroclasts within the Cpx/Pl-rich mylonitic layers (e.g., AN01D) are hedenbergite as well as the composition of clinopyroxenes occurring in the calc-silicates at the contact with the shear zone (e.g., AN06C; Fig. 8B). Amphiboles are mainly pargasite in composition with minor hornblende as recrystallized grains within Cpx/Pl-rich mylonites (Fig. 8A).

As major element compositions, even the clinopyroxene REE patterns mark differences among mylonitic amphibolites and gabbroic rocks (Fig. 10A). Clinopyroxene from mylonitic amphibolites, indeed, have general lower REE concentration with respect to those within the gabbro; on the contrary, they partially mimic the composition of Cpx within the Scp-bearing gneiss from the hanging wall (Fig. 10A). Similarities were observed also between clinopyroxene from Cpx/Pl-rich mylonitic layers and the clinopyroxene within calc-silicates from the footwall (Fig. 10A). The REE patterns of clinopyroxene from gabbroic rocks share many similarities with clinopyroxene from other mafic bodies of the IVZ (e.g., Mazzucchelli et al., 1992, Zanetti et al., 2013 and Berno et al., 2020). In particular, the comparison of the REE patterns highlights a strong affinity of the Anzola gabbro with the mafic intrusion of the adjacent Val Strona di Omegna (Berno et al., 2020; Fig. 10A). The REE patterns of amphibole from mylonitic samples are extremely variable (Fig. 10B). Such variations can be attributed to control by variations in: i) the rock types/bulk chemistry, i.e., Amph-rich (amphibolite) vs Cpx/Pl-rich (calc-silicate) layers and ii) the textural features, e.g., porphyroclasts (AmphI) vs syn-kinematic grains (AmphII). Analogously to clinopyroxene also the REE patterns of amphibole from mylonites do not show strong affinity with REE patterns of amphibole from gabbroic rocks (Fig. 10B). Only the REE patterns of AmphII partially overlap those obtained from gabbroic rocks but differ for a more pronounced LREE fractionation over HREE and a more apparent Eu anomaly. As for clinopyroxene, the amphibole compositions of Anzola gabbro share many similarities with amphibole REE patterns of other mafic intrusions of the adjacent Val Strona di Omegna (Berno et al., 2020; Fig. 10B). It is worth to note that the REE patterns of AmphI share many similarities with those obtained from amphiboles within hornblendites fractionated from dioritic magmas and intruded in metamorphic roof (Liou and Guo, 2019). Even in our case, we cannot exclude that the REE patterns of AmphI are inherited from a magmatic protolith such as amphibole-rich dikes or sills intruded within metamorphic roof of the gabbroic intrusion. Moreover, similar mafic dykes and sills have been documented within the migmatitic roof of the Mafic Complex exposed in other localities of the IVZ (Quick et al., 1993).

Summing up, both major and trace element compositions of clinopyroxene and amphibole indicate a predominant metamorphic affinity for the protoliths of mylonitic amphibolites. Beside these, AmphI-rich mylonitic layers may represent former dykes/sills related to the gabbroic rocks that intruded in the metamorphic roof before shearing. Combining field data with petrography and geochemistry we infer that the Anzola shear zone developed from a heterogeneous protolith consisting of alternating paragneisses, mafic gneisses and calc-silicates of the Kinzigite Formation (i.e., a metamorphic volcano-sedimentary sequence).

7.3 Formation of a major crustal shear zone along pre-existing rheological boundaries

In this work, we show that the Anzola shear zone developed in a complex compositional, structural and metamorphic setting dominated by alternating paragneisses and supracrustal mafic rocks (Fig. 12). High strain deformation overprinted metamorphic rocks that already experienced both Variscan and late Variscan tectono-metamorphic and magmatic stages (Fig. 12A). In particular, the shear zone took advantage of the pre-existing transition zone between upper amphibolite and granulite facies metamorphic rocks, which is characterized by abundant migmatites (Fig. 12B). Gneissic rocks involved in high strain shearing were also characterized by abrupt changes i.e., marked heterogeneity in composition, textural features (layering) and geometric features at different scales (Fig. 12A). Geometric anisotropy stem from inherited Variscan structures, such as folds and associated shear planes, which characterize the gneissic protolith of the shear zones (Fig. 12A). The attitude of fold limbs, almost parallel to the shear planes, is consistent with preferential shear along them (e.g., Austrheim, 1987; Piazolo et al. 2002; Hughes et al., 2014; Fossen et al., 2019). As such, our results from the Anzola shear zone show that pre-existing features have a marked influence on the nucleation sites of subsequent ductile shear (e.g., Prosser et al., 2003; Pennacchioni and Mancktelow, 2007; Smith et al., 2015, Gardner et al., 2017). From micro- to large scale, the lithological and structural inherited heterogeneities had a first-order control driving the development of mylonites within gneissic rocks instead of within the isotropic gabbroic body (Fig. 12B).

However, the lithological and structural variability alone does not explain why deformation localised in both the stronger mafic rocks and weaker paragneisses and calc-silicates. Another important factor to consider is the availability of fluids or melts, which may change the response of the system by localizing deformation into high strain zones (Davidson et al., 1992; Getsinger et al., 2013; Yardley et al., 2014, Piazolo et al. 2020). Even little amounts of free fluid (e.g., water or melt) along the grain boundaries may have a major rheological effect on lower crustal shear zone by facilitating deformation and phase nucleation as well as the stabilization of syn-deformational paragenesis (e.g., Okudaira et al., 2015; Menegon et al., 2017). In our case, the compositional variations are associated with alternating layers with different H_2O bulk contents (Table S2), as well as by the changes in modal abundances of OH-bearing mineral, i.e., amphibole and biotite, within the layering of the mylonitic rocks (Tables S1; Fig. S1A, D, E). The mobilization of free water or melt from these variable water storages may be related to the dehydration reactions involving OH-bearing mineral during the high temperature stage of deformation. We documented that at the beginning of deformation, the high temperature conditions (>800°C) promoted dehydration reaction and partial melting leading to the generation of melts and water (Fig. 11; S2). The addition of melt to the pre-existing lithological and structural anisotropies accentuated the weakening mechanisms (e.g., Mei et al., 2002; Piazolo et al., 2002; Stuart et al., 2018) acting in the multilayer amphibolite-granulite transition zone. The lower grade evolution of the shear zone took place under water-present conditions as documented by the formation of abundant syn-kinematic amphibole and biotite (Figs. 5-6; S1A; Vauchez et al., 2012; Casini et al., 2021).

What remains still enigmatic is why deformation does not affect the rocks directly at the contact with the gabbro. These rocks show the same lithological and structural features as those characterizing the protoliths of mylonitic rocks. However, here, we observed local evidence for de-hydration reactions due to contact metamorphic effects. The static growth of coronitic garnet replacing amphibole is documented in the mafic gneisses from the footwall (Fig. 4D). In this zone we observed also larger sizes of mineral phases and metasomatic effects leading to the formation of reaction boundaries between gabbros and their host rocks (marbles/calcsilicates). These features have never been documented in other mafic rocks of the entire crustal section exposed in the Ivrea-Verbano zone. On the other hand, a similar feature of intrusion related rheological hardening has been observed and described at a small scale in granulite-facies metamorphic rocks from the Fiordland, New Zealand (Smith et al. 2015). Therefore, a possible reason why this meter-thick region escaped deformation could be the local (meter scale) contact metamorphic effect due to the intrusion of gabbroic rocks promoting dehydration reaction that changed the rheological behaviour of these gneisses (Fig. 12A, e.g., Harlov, 2012). Alternatively, it could be that the original magmatic contact between the gneiss and the gabbroic rocks is curvilinear whereas deformation occurred mainly along planar planes saving up some gneisses in bights of the contact. To evaluate the hypotheses discussed, more detailed investigations temporally constraining the evolution of the Anzola shear zone with respect the gabbroic body would be needed.

7.4 The role of the Anzola shear zone during the Triassic-Jurassic rifting phases

In the following section, we discuss the implications of our findings in the framework of the geodynamic evolution of IVZ, with particular focus on Val d'Ossola section (Fig. 12).

After the Variscan orogeny, the IVZ experienced post-orogenic extension (Handy et al., 1999). From the Early to Middle Permian (290–260 Ma), the lithospheric extension was associated to intense magmatic activity (e.g., Petri et al., 2019 and references therein). Consequently, heat advected by hydrous melts combined with lithospheric extension was responsible for widespread regional high temperature metamorphism (e.g., Ewing et al., 2015).

In the meantime, mafic magmas intruded at mid crustal levels (ca. 280 Ma; Peressini et al., 2007; Brodie et al., 1989) within metamorphic rocks showing evidence for partial melting and intense folding (Fig. 12A; C). The rocks at the contact with mafic intrusion reached high temperature conditions developing locally features comparable to those due to contact metamorphism (e.g., Barboza & Bergantz, 2000). Consequently, all these features promoted the formation of crustal boundary characterized by significant rheological and compositional heterogeneities (Fig. 12A).

From the Early Mesozoic, multiple episodes of rifting affected the IVZ. During these events, deformation was mainly accommodated by several shear zones active at different crustal levels (Beltrando et al., 2015; Manatschal et al., 2015). Although precise geochronological estimates are still lacking, the Anzola shear zone has been interpreted as a main extensional structure accommodating the Triassic-Jurassic rift-related deformation (Fig. 12D; Simonetti et al. 2021 and references therein). The nucleation of these crustal scale shear zones is interpreted as primary driven by tectonic inheritance (Lavier and Manatschal, 2006; Petri et al., 2019). The impact of compositional and structural heterogeneities results in potentially abrupt contrasts in the rock strengths, promoting strain localization in the weaker layers. Furthermore, the temperature dependence of viscosity highlights the crucial role played by thermal softening (Smye et al., 2019). Thermochronological constraints and thermal models (Siegesmund et al., 2008; Smye et al., 2019) predict the onset of deformation at high-T conditions in the IVZ middle-lower crust prior to the initiation of mantle exhumation at ~180 Ma. Later, fault-related advective infiltration of high-temperature fluids accompanied the crustal-mantle exhumation.

Our study confirms that also at small scale the locus of strain localization was driven by rheological boundaries originated from lithological and structural contrasts occurring at the amphibolite to granulite facies transition (Fig. 12A, B). Deformation localized at high-T conditions (>800°C), supporting the evidence of a heated IVZ crustal section at the beginning and during rifting. Here, melts/fluids assisted-deformation accompanied the shear zone since the onset throughout all its evolution. Small amount of melt helped to localize deformation at high-T conditions (>800°C), while the ongoing deformation produced permeable pathway were fluids migrated, promoting positive feedback with the deformation processes (Austrheim, 1987, 2013; Getsinger et al., 2013; Lee et al., 2020; Casini et al., 2021; Piazolo et al. 2020). The Triassic-Jurassic mantle derived magmatic rocks and fluids, which have been extensively documented within middle-lower crustal lithologies of the IVZ (Schaltegger et al., 2015; Denyszyn et al., 2018; Bonazzi et al., 2020; Corvò et al., 2020), were the potential sources for melt/fluid components associated with the thermal pulses.

Finally, we argue that the combination of local melting, and the intrinsic rheological heterogeneity of poly-tectonometamorphic complexes with geometric and compositional variability originating from pre-existing structural features (e.g., folds) and metamorphic assemblages allowed the development of the crustal-scale shear zones during Triassic-Jurassic rifting since high-temperature conditions.

8. Conclusions

In this work, we present new field observations, petrological and geochemical data from a major extensional shear zone – the Anzola shear zone – of the middle/lower crust of the Ivrea-Verbano Zone (Southern Alps) with the aim to define the nature of pre-shear heterogeneities and reconstruct the factors that promoted strain localization. While the shear zone has been previously described as developed within an isotropic gabbroic body, our new geological mapping revealed instead significant pre-existing lithological and structural heterogeneities. Mylonitic rocks derive from a metamorphic volcano-sedimentary sequence, in which the gabbro was intruded before the shear zone development. In particular, petrographic and geochemical data indicate that the protoliths of (ultra-)mylonitic rocks coincide with the folded paragneisses, mafic gneisses and calc-silicates rather than the nearly isotropic gabbro. Noticeably, trace elements characterization resulted as the most powerful tool to constrain the geochemical links between minerals from the wall rock and shear zone. Petrographic observations suggest that deformation took place under amphibolite facies conditions and was firstly aided by the local presence of interstitial melt and subsequently by fluids. The application of different geothermometers (Zr-in-Rutile, Grt-Bt, Ti-in-Amphibole) and a geobarometer (GASP) allowed to constrain P-T conditions of deformation started at high temperature (*T:* ~820°C; *P*: 0.8 GPa) and continued down to lower amphibolite facies (T: ~650°C *P*: 0.7 GPa) following a retrograde path. Overall, the Anzola shear zone is characterized by a complex pattern of lithotypes that experienced, before shearing, multiphase Variscan folding followed by high-temperature metamorphism at the upper amphibolite to granulite facies transition. The summing up of contrast between i) hard and relatively weak lithologies (isotropic gabbro and granulites vs. folded metavolcanites and metasediments), ii) layering of anhydrous- and hydrous-dominated metamorphic assemblages (transition between amphibolite to granulite facies) and iii) the heterogeneous presence of melt promoted the strain localization at high-T conditions. In particular, the rheological contrast between the isotropic/dry strong gabbro/granulite and the weaker multi-lithological assemblages containing melt, provides the ideal place where strain was localized. This study confirms that the combination of tectonic/metamorphic inheritance and thermal processes have first-order control driving the nucleation of shear zones.

In this frame, the Anzola shear zone played a significant role in accommodating the deformation during crustal thinning and exhumation phases during the Triassic-Jurassic time.

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Figure Captions

Fig. 1 – Geological sketch map of the Ivrea-Verbano Zone, modified after Ewing et al. (2015) and Simonetti et al. (2021). The locations of high-temperature shear zones are after Rutter et al. (1993). The studied area is reported in the black square.

Fig. 2 – Schematic geological maps showing the main lithologies, tectonic structures and isograds through the Ivrea-Verbano Zone for A) the Val d'Ossola (modified after Rutter et al., 2007). Isograds (Kfs + Sill in and Opx in) are from Zingg (1980) and mark the transition between amphibolite to granulite facies. IL = Insubric Line; CMBL= Cossato-Mergozzo-Brissago Line. In the legend on the right: AF = Amphibolite facies; GF = Granulite facies. B) Detailed geological map and structural data of the study area. Sampling across the Anzola shear zone (red coloured), from footwall (NW) to hanging wall (SE), is marked with stars. Samples analysed for geochemical analyses are labelled in

italic characters. In the legend on the right, the patterns labelled as MB, MP, CS refer to the different rock types: metabasite, metapelite, calc-silicate, respectively.

Fig. 3 – A) Sketch of the Anzola shear zone transect (not at scale; approximate thickness of the shear zone ~100m) and representative outcrop pictures (B-M). Shear zone boundaries are defined on the base of foliation orientation (see Figure 2B and the text for geological mapping and measurements). B) Migmatitic paragneiss (AN11). C) Protomylonitic paragneiss (AN22A). D) Mylonitic paragneiss outcrop (i.e., AN09). E) Ultra-mylonitic paragneiss (sample AN09C). F) Boudinaged leucocratic rock (AN22B). G) Mylonitic banded amphibolites (AN01). H) Folded domain made by calc-silicates and gneisses. I) Contact between calc-silicates and gabbroic body. L) Coarse-grained pocket (AN07) at the contact with gabbro (AN07). M) Leucocratic dyke cutting the gabbro.

Fig. 4 – Microphotograph reporting the main petrographic features of the wall rocks. A) Migmatitic paragneiss (AN11) made of melanosome mainly constitutes by garnet rutile and prismatic sillimanite and leucosome made by Qz and Fsp; B) Scapolite-bearing gneiss (AN10) with mineral assemblage made by Cpx + Amph + Pl + Ttn; C) Mafic gneiss (AN06F) from the folded layers at the footwall made by Amph + Pl + Cpx + Chl + Ilm; D) Backscattered electron image (BSE) of sample (AN06F) showing garnet corona around plagioclase and surrounded by Amph + Cpx + Qz + Ap + Ilm; E) BSE image of paragneiss (AN06D) from folded layer at the footwall showing the garnet relict surrounded by a film of Sil + Grt + Pl + Kfs. F) Calc-silicate (AN06C) with very big crystals of titanite and mainly constitute by epidote and chlorite and calcite; G) Gabbros (AN07/AN08) sample with typical mineral assemblage composed by Cpx + Pl + Amph; H) Felsic granulite (AN18) mainly constitute by feldspar, biotite and garnet.

Fig. 5 – Microphotograph of the main petrographic and microstructural features of the mylonitic amphibolites (AN01). A) Cpx porphyroclast and recrystallized amphibole (AmphII); B) Amphibole porphyroclasts (AmphI); C) Amphibole porphyroclasts in Pl/Cpx-rich mylonite and titanite grains; D) Plagioclase porphyroclasts (PlI) and matrix rich in amphibole and plagioclase; E) Garnet relict surrounded by Amph, Pl and Ttn; F) Big crystals of titanite; G) Clinopyroxene in Cpx/Pl-rich layer (AN01D), with recrystallized CpxII and PlII; H) Calcite in Cpx porphyroclasts and surrounded and included amphibole.

Fig. 6 – Microphotograph reporting the main petrographic and microstructural features of the mylonitic paragneisses. A) Back-Scattered Electron (BSE) image of mafic ultramylonite (AN06L) with amphibole and plagioclase porphyroclasts and matrix mainly made of Amph + Pl + Cpx; B) Mylonitic felsic granulite (AN09) with evident foliation marked by quartz + feldspatic and/or garnet + sillimanite layers; C) Ultra-mylonitic felsic granulite (AN09C) with layer enriched in Grt + Bt + Fsp and layer constitutes by Pl + Qz; D) Ultra-mylonitic felsic granulite (AN09C) with layer enriched in porphyroclast of prismatic sillimanite with an alteration rim of muscovite surrounded by fine matrix made of Bt + Fsp. E) Felsic mylonitic gneiss (AN22A) with mineral assemblage made of Grt + Kfs + Pl + Sil + Spl + Bt; F) Corundum-bearing leucocratic rock (AN22B) made of Grt + Kfs + Pl + Bt + Spl + Crn.

Fig. 7 – Bulk rock diagrams: A) Al_2O_3 (wt.%) versus SiO_2 (wt.%). B) MgO (wt.%) versus CaO (wt.%). C) REE patterns for footwall and hanging wall rocks; D) REE patterns for shear zone rocks. All data are normalised to chondrites values according to McDonough and Sun, (1995).

Fig. 8 – Classification and geochemical diagrams for main mineral phases for the mafic rocks and calc-silicates. Mineral composition diagrams for: A) Amphibole; B) Clinopyroxene; C) Plagioclase.

Legend of the symbols is reported; in particular, “porphyroclasts” refered to AmphI, CpxI and PlI whereas “recrystallized” refered to AmphII, CpxII, PlII in figure 5 of petrography.

Fig. 9 – Classification and geochemical diagrams for main mineral phases for paragneisses and felsic granulites. Mineral composition diagrams for: A) Plagioclase; B) Garnet. C) Aluminium (a.p.f.u.) versus Fe# (Fe^{2+}/Fe^{2+}+Mg) diagram for Biotite; BtI and BtII from figure 7 were distinguished in filled and open diamond, respectively. Legend of the symbols is reported.

Fig. 10 – REE diagrams for A) Clinopyroxene. In colours our samples, in thick dark-grey line data for mafic intrusion from Val Strona di Omegna (Berno et al., 2020), in light grey band data for Finero mafic complex from Zanetti et al. (2013), in dashed lines data for Mafic Complex from Mazzucchelli et al. (1992). Gabbro data are distinguished on the base of the grain size (coarse grained- CpxI; small grained-CpxII) B) Amphibole. In colours our samples, in dark-grey band data from Berno et al. (2020), in light grey band data from Zanetti et al. (2013). Data for Amph-rich mylonites and gabbro are distinguished on the base of textural features (i.e., porphyroclasts, recrystallized, coarse and small grained).

Fig. 11 – Pre-shear and syn-kinematic temperature estimates for the studied samples across the Anzola transect. Zr-in-Rutile temperature conditions for felsic granulite (R20) are from the studied area from Luvizotto & Zack, (2009) recalculated with the new calibration of Kohn, (2020). The grey box encloses the Zr-in-Rutile data and broadly defines the temperature gradient from granulitic to upper amphibolitic facies conditions. The temperature conditions for mylonitic amphibolites determined by Brodie (1981) and Stünitz (1998) are also shown as boxes a) and b), respectively. Ti-in-Amphibole estimates for Scp-bearing gneiss (AN10) and Grt-Bt data are reported as mediana values and relative standard deviations. For Zr-in-Rutile and Ti-in-Amphibole mylonites single crystals data mediana with standard deviations is defined by white circles/diamond and red/black lines.

Fig. 12 – Interpretation of the geological conditions that promoted the nucleation of the Anzola shear zone at the transition between middle to lower crust at the meso-scale (A, B) and regional scale (C, D), (modified after Gardner et al., 2017). A) Before the Anzola shearing, the transition between middle to lower crust is characterized by the juxtaposition of: i) a gradual transition from anhydrous conditions (granulite facies) to hydrated assemblages (amphibolite facies rocks) and, ii) pre-existing compositional and deformational heterogeneities (late Variscan undeformed isotropic gabbro and its contact aureole vs folded Variscan metamorphic volcano-sedimentary sequence). B) The combinations of these factors resulted in the formation of strong rheological contrasts that promoted the initiation of the Anzola shearing during Triassic, when the Tethyan rifting stage triggered renewed extensional tectonics. *P-T* conditions for deformation are reported as T_1, T_2, P_1, P_2 to indicate that deformation started at high temperature (T_1: ~820°C; P_1: 0.8 GPa) and continued down to amphibolite facies (T_2: ~650°C P_2: 0.7 GPa) following a retrograde path. Interpretative, paleo-cross section of the Val d’Ossola lithosphere: C) during Permian (pre-Anzola shearing) and D) during Triassic (during Anzola shearing). Note the migmatization at the middle-lower crust transition and the metamorphic halos (hardened aureole) developed due to the emplacement of the gabbroic body during Permian age. During Triassic, this layer is overprinted by the nucleation of the shear zone. Modified after Petri et al., (2019).

Supplementary materials

Figures Supplementary

Fig. S1 – Representative hand specimens and thin sections of the studied samples. A) Block of mylonitic banded amphibolites (sample AN01). B) Gabbro sample from the contact with garnet layer (AN07). C) Protomylonitic paragneiss (sample AN22B), blue rounded corundum porphyroclasts and leucocratic layer are well visible at the mesoscale. D) Representative thin sections obtained from sample in A) with evidence of textural, grain size and modal abundance differences distinguishable at the thin section scale. E) Representative thin sections of the Anzola shear zone transect from hanging wall to footwall rocks across the shear zone.

Fig. S2 – Microstructural features indicative of the former presence of melt: A, B) Back-Scattered Electron (BSE) images of ilmenite terminating with low dihedral angles (white arrows) in mylonitic amphibolites (e.g., AN01E). C) Microphotographs and BSE images highlighting quartz grain boundary low dihedral angles and interstitial textures (white arrows) in mylonitic paragneisses (e.g., AN09; AN22A).

Fig. S3 – Bulk rock diagrams. A) TiO_2 (wt.%) versus SiO_2 (wt.%). B) Co (ppm) versus Cr (ppm). All data are normalised to chondrites values according to McDonough and Sun, (1995). C) Ti (a.p.f.u.) versus Mg# = ($Mg/Mg+Fe^{2+}$) diagram for Biotite; BtI and BtII from figure 7 were distinguished in filled and open diamond, respectively.

Tables Supplementary

Table S1 - List of the locality, coordinates, textural features and mineral assemblages of the studied samples from Anzola Shear Zone transect from SE to NW.

Table S2 - Bulk rock major and trace element composition of representative studied rocks across the Anzola transect. Iron contents are expressed as all ferric.

Tab. S3 – Average major elements composition (wt.%) for feldspar, clinopyroxene, amphibole, garnet, biotite and other phases from the studied samples.

Tab. S4 – Average trace elements concentration (ppm) in clinopyroxene and amphibole from the studied mafic rocks.

Table S5 – Geothermobarometry results of the studied samples across the Anzola shear zone. Results are reported as mediana values and relative standard deviations.

Table S6 – Zr-in-Rutile geothermometer of the rutile-bearing studied samples across the Anzola shear zone.

Appendix - Analytical Methods

A.1 Scanning Electron Microscopy

Back-Scattered Electron (BSE) imaging of the studied samples texture and microchemical analyses on the mineral/matrix composition were also performed by Scanning Electron Microscopy (SEM) using a Tescan Mira3 XMU-series FESEM equipped with an EDAX-EDX (accelerating voltage 20 kV, beam intensity 16.5 nA, spot area 100×100µm, counts of 100 s., analyses and working distance 15.8 mm; University of Pavia, Italy); data was processed with EDAX Genesis software using the ZAF algorithms the correction method.

A. 2 Bulk rock geochemistry

Bulk rock major and trace element composition of 11 samples were determined by standard X-ray fluorescence spectroscopy (XRF) spectroscopy and Laser Inductively Coupled Plasma Mass Spectrometry (LA-ICP-MS) at the Activation Laboratories Ltd., Ancaster, Canada (4E-Research + ICPMS method). Table S2 gives the major and trace element composition for four representative samples from Anzola shear zone.

A.3 Electron Microprobe Analyses (EMPA)

The major-element composition of main mineral phases was determined by JEOL 8200 Electron Microprobe Analyses at the Dipartimento di Scienze della Terra "Ardito Desio", University of Milan. Average data and standard deviations are reported in Table S3.

A.4 Laser Ablation Inductively Coupled Plasma Mass Spectrometry (LA-ICP-MS)

Trace element concentrations have been determined by means of the LA-ICP-MS at the IGG-CNR, SS. of Pavia (Italy). Average values are reported in Table S4. A PerkinElmer SCIEX ELAN DCR-e quadrupole ICP-MS was coupled with a 266nm Nd:YAG laser (NewWave Research). Helium was used as carrier gas and mixed with Ar downstream of the ablation cell. Laser spot size was calibrated between 50 and 60 _m and laser. beam fluency at 8-9 J/cm^2. Data reduction was performed with GLITTER software, using the reference synthetic glass NIST (SRM) 610, NIST612 as external standards. ^{29}Si was used as internal standards for plagioclase; ^{44}Ca for clinopyroxene, and amphibole. Precision and accuracy were assessed via repeated analyses of basalt glass (BCR-2g) reference material, resulting better than 10% at ppm concentration level. The uncertainties related to the trace element data is of 1σ. Detection limits were typically in the range of 100-500 ppb for Sc, 10-100 ppb for Sr, Zr, Ba, Gd and Pb, 1-10 ppb for Y, Nb, La, Ce, Nd, Sm, Eu, Dy, Er, Yb. Hf and Ta, and usually <1 ppb for Pr, Th, and U.

A.5 Zr-in-Rutile thermometer

Rutile trace element analyses were carried out at the IGG-CNR of Pavia by using a 193nm Excimer Laser (Geolas) combined with an Agilent 8900 ICP quadrupole mass spectrometer. The samples were ablated in spots of 35/50 µm at a laser energy of 4 Jcm^{-2} and a repetition rate of 10 Hz. Signals were recorded for 50 s per spot after measuring a gas blank of 40 s for background subtraction before each spot. Plasma conditions and gas flow rates were adjusted by optimising He and Ar flow, monitoring $^{232}Th^{16}O^{+}/^{232}Th+$ ratios (always ≤ 0.2%) and $^{238}U^{+}/^{232}Th^{+}$ ratios (always between 0.95 and 1.05) while ablating NIST SRM 612 in no-gas mode (no reaction gas in the reaction cell). Individual mineral grains were ablated in a He atmosphere (flow rate 0.44 L/min) and the resulting aerosol was mixed with Ar carrier gas (flow rate 0.91 L/min) to be transported to the ICP-MS via signal smoothing device. Titanium, measured as 49Ti, was used as the internal standard element for each analysis. TiO_2 was assumed to be 100 wt.%. Analyses were calibrated against the NIST SRM 610 glass (GeoReM preferred values: http://georem.mpch-mainz.gwdg.de/). Accuracy of the results was confirmed by analyses of the well-known USGS reference glass BCR-2G (Jochum et al., 2005) using newly compiled values (GeoReM preferred values: http://georem.mpch-mainz.gwdg.de/) and NIST SRM 612 glass. Analyses were collected in MS/MS mode. The following masses were measured and the dwell times for each mass is given in brackets (in ms): 29Si (20), 45Sc (20), 49Ti (10), 51V (20), 52Cr (20), 53Cr (20), 57Fe (20), 88Sr (20), 89Y (20), 90Zr (20), 93Nb (20), 95Mo (20), 98Mo (20), 118Sn (20), 178Hf (20), 181Ta (20), 182W (20), 208Pb (20), 232Th (20), 238U (20). The software GLITTER ® was used to data reduction (Van Achterbergh et al., 2001). 29Si was measured in order to detect possible silicate inclusions while 57Fe was used as a good indicator in order to detect ilmenite.