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Geological map of the Klopai Pluton and its surroundings (Austroalpine Unit, Reschen Pass, Eastern Alps)

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Cover
Above: View of the study area, looking to east-south east. The highest peak on the left side is the Klopaierspitze. Reschen village and lake lies at the mountain side.
Below: Nauderer migmatites

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ABSTRACT

The Austroalpine Ötztal Nappe of the Eastern Alps preserves a polymetamorphic record ranging from pre-Ordovician to Alpine tectono-metamorphic events. In the Reschen Pass area (South Tyrol, Italy), Variscan mineral assemblages and a heterogeneous tectonic overprint coexist with possible pre-Variscan features, a rare occurrence in the Alpine Domain. We present a new 1:10,000 geological map of the area –geological mapping project, Autonomous Province of Bolzano– focusing on the Klopai tonalitic pluton and its surrounding lithologies. Detailed field mapping combined with petrographic characterization and structural analysis, reveal that migmatites occur exclusively at the contact with the pluton, whereas cordierite-bearing gneisses may mark the original contact aureole, suggesting a genetic link between pluton emplacement, contact metamorphism, and partial melting. Pegmatite dykes are associated with the pluton and its aureole, constraining the late magmatic stages. Our interpretation favours a relatively shallow emplacement (<10 km) during an early Paleozoic high-temperature event, broadly coeval with migmatization and regional crustal thinning. The overprinting Variscan deformation developed a penetrative foliation and high-grade mineral assemblages (staurolite, kyanite), while Alpine deformation is less recorded. These results refine the geological framework of the southwestern Ötztal Nappe, with implications for the reconstruction of the early Paleozoic geodynamics of the Gondwana margin.

Keywords: Klopai Pluton, Austroalpine, Ordovician magmatism, Ötztal Nappe, Eastern Alps.

INTRODUCTION

The European Alps, which formed during the Cretaceous-late Cenozoic convergence and collision between the European and African plates exhibit a complex architecture, derived from both the Alpine orogenic processes (e.g., Eoalpine and Neoalpine), and inherited from the Variscan Orogeny and the Permian to Jurassic rifting stages (Muttoni et al., 1996). The nappe-stack of the N-vergent part of the Alps (i.e., N of the Periadriatic Fault) comprises units derived from various palaeogeographic domains (Schmid et al., 2004; Handy et al., 2010; Pfiffner, 2014). The result is that most of the tectono-metamorphic units experienced a polymetamorphic evolution and several phases of deformation, where differences in structural position, lithological composition and strain partitioning have resulted in a heterogeneous metamorphic and deformational overprint, locally so subtle that older geological features have been variably preserved. This is the case of the Ötztal Nappe, one of the largest polymetamorphic units in the Eastern Alps, extending for a 50 km long and 20 km wide region across northern Italy and Austria. The Ötztal Nappe locally preserves late Carboniferous structures and metamorphic assemblages formed during the Variscan Orogeny (e.g., Gregnanin and Piccirillo, 1972; Thöni, 1981, 1986; Hoinkes and Thöni, 1993; Schweigl, 1995; Tropper and Hoinkes, 1996; Hoinkes et al., 1997) and even vestiges of older, Ordovician to Silurian high temperature event (e.g., Chowanetz, 1990; Schweigl, 1995; Bernhard et al. 1996; von Raumer et al., 2013; Zurbriggen, 2015, 2020; Neubauer et al., 2022; Piccin, 2026), making it a valuable archive for reconstructing the early geological history of the region.

In the SW Ötztal Nappe (Reschen Pass area, South Tyrol, NE Italy; Fig. 1) a combination of pre-Variscan features such as magmatic intrusions, contact metamorphic aureole, and migmatites occur within a relatively small area (~25 km²). The elsewhere penetrative Variscan deformation and metamorphic re-equilibration are here heterogeneously distributed, allowing the preservation of relict structures.

In this contribution, we present a geological map in scale 1:25,000 of this peculiar area, offering a comprehensive description of its lithologies and structural relationships. The primary goal is to establish a robust geological framework to support future investigations aimed at resolving the long-debated sequence of events –a geological "chicken-or-egg" dilemma: did the pluton intrude into already solidified metatexites, or did its emplacement and thermal influence trigger partial melting in the surrounding rocks?

GEOLOGICAL SETTING

The accretion of the Alpine orogen was already active in the Late Cretaceous with the subduction of the Meliata ocean and partly of the Western Alpine Tethys (Eoalpine event; Pomella et al., 2016; Dilek and Furnes, 2019), followed by the Cenozoic continental collision between the European Plate and the Apulian Plate (Neoalpine event; Dal Piaz et al., 2003). During the Jurassic, the Apulian Plate was bordered to the north by the newly formed Alpine Tethys and to the east by the westernmost branch of the Neotethys (including Meliata) Ocean, which were simultaneously undergoing closure (Stampfli et al., 2002; Schmid et al., 2008; Van Hinsbergen et al., 2020). In the Cretaceous, the Apulian Plate experienced a west-directed orogenesis, leading to an intracontinental subduction to the west and the closure of the Meliata ocean farther East, and generating a larger continental mass (Schmid et al., 2004, Pomella et al., 2016).

This tectonic phase, referred to as the Eoalpine event, was characterised by high-pressure metamorphism in several units of the Austroalpine domain (Hoinkes et al., 1991; Miller and Thöni, 1997; Thöni and Miller, 2000; Habler et al., 2006; Thöny et al., 2006; Zanchetta et al., 2013), associated to contemporaneous and slightly younger westward nappe thrusting (e.g., Ratschbacher, 1986; Schmid and Haas, 1989; Froitzheim and Manatschal, 1996; Thöny et al., 2008).

The North-directed Cenozoic orogeny (Neoalpine event) dragged the Apulian Plate northward. In this context, the Eoalpine Austroalpine nappe stack was transported northward as a coherent slice, overriding the oceanic and distal European margin units (Penninic and Subpenninic Domains), (Trümpy, 1975; Laubscher, 1983; Schmid et al., 2004, 2013). The collision with the European continent determined the final closure of the Alpine Tethys and led to the formation of the present Alpine chain.
A late Alpine post-collisional extensional phase fragmented the Austroalpine domain, and exhumed the Penninic and Subpenninic nappes, which are now exposed in tectonic windows — the Engadine, Tauern and Rechnitz Windows — via major normal faults: Brenner, Katschberg and Engadine Faults (Trümpy, 1977; Fügenschuh, 1995; Selverstone et al., 1995; Fügenschuh et al., 1997; Froitzheim et al., 1997; Bousquet et al., 1998; Neubauer et al., 1999; Rosenberg and Garcia, 2011; Scharf et al., 2013; Schmid et al., 2013; Schneider et al., 2015; Bertrand et al., 2017; Favaro et al., 2017).
The Ötztal Nappe (part of the Ötztal-Bundschuh Nappe System), is traditionally classified as part of the Austroalpine basement (*Altkristallin*; Pestal et al., 2009) plus its sedimentary cover (*Brenner Mesozoikum*), and lies between these two tectonic windows in its northern part (Fig. 1). The Ötztal Nappe is thrusted on top of the Silvretta-Seckau Nappe System (Ortler-Campo Nappe and Sesvenna unit) along the Schlinig Thrust to the west and the Vinschgau Shear Zone (VSZ) to the south (Schmid and Haas, 1989; Froitzheim et al., 1997). To the SE it is in tectonic contact along a wide mylonitic shear zone with the high pressure units of Schneeberg and Texel (Schmid and Haas, 1989; Viola et al., 2001; Sölva et al., 2005; Pomella et al., 2011, 2012; Montemagni et al., 2023). To the east, it is in direct contact with the Tauern Window, along the Brenner Normal Fault (Fügenschuh et al., 1997) and to the north, the tectonic boundary with the Northern Calcareous Alps is represented by the Inntal Fault: a left transpressional strike-slip fault (Ortner et al., 2006).
Our study area lies near the southwestern boundary of the Ötztal Nappe, tectonically interposed between two fragments of the Austroalpine basement, the overlying Matsch Unit and the Sesvenna unit with its preserved Permo-Mesozoic cover (Fig. 1; Hess, 1962, Thöni, 1973; Schmid and Haas, 1989; Habler et al., 2009). The contact with the Matsch Unit is typically marked by a ductile shear zone characterised by mylonites of variable thickness (Geological Sheet 012 "Silandro" at 1:50,000 - CARG - in progress). A Cretaceous top-to-the-west thrust represented by the Schlinig Line at its front and by its deeper portion, the mylonitic shear zone of the Vinschgau (VSZ) in the west, juxtaposes the Ötztal Nappe with the Permian-Mesozoic Engadine Dolomites (Thöni, 1980; Schmid and Haas, 1989; Montemagni et al., 2023) and the Austroalpine Ortles-Campo nappe, respectively.
The Ötztal Nappe displays a well-defined Eoalpine metamorphic gradient that increases from northwest (300 °C) to southeast (up to 550 °C) (Thöni, 1981; Purtscheller and Rammlmair, 1982; Purtscheller et al., 1987; Hoinkes and Thöni, 1987; Hoinkes et al., 1999; Raso et al., 2025). In the northwest, the lower greenschist-facies conditions were reached, while in the southeast, amphibolite-facies conditions were recorded (Hoinkes et al., 1991).
Therefore, in some areas, these rocks lack a clear Eoalpine tectono-metamorphic overprint (Niggli, 1960; Purtscheller and Rammlmair, 1982), allowing the preservation of pre-Alpine geological features. In fact, prior to the Permian rifting stages, the Apulian Plate was part of the Variscan orogenic belt (Berra et al., 2009). The Variscan event was responsible for the dominant metamorphic overprint of the Ötztal Nappe. The subsequent Alpine orogeny emplaced the Ötztal-Bundschuh Nappe System in a high structural position, just below the topmost Grauzug-Gurktal Nappe System (Schmid et al., 2004), resulting in a nearly continuous exposure of the Variscan basement (Pomella et al., 2016).
In the study area, the Variscan metamorphism is well preserved and the area is dominated by a lithological assemblage made of paragneiss that record Variscan amphibolite-facies metamorphism. In the northern part of the area, however, the presence of a large intrusive body (i.e., the Klopai tonalitic pluton) and adjacent migmatites are thought to give evidence of an older, high-temperature pre-Variscan event (Schweigl, 1993, 1995; Thöny et al., 2008). This metamorphic-magmatic high-temperature event, dated to the Ordovician–Silurian, was alternatively attributed to either crustal thinning and rifting or to a subduction accretion complex (Frisch, 1984; Bernhard et al., 1996; Schulz, 2008; Thöny et al., 2008; von Raumer et al., 2013; Zurbriggen, 2015, 2017, 2020; Siegesmund et al., 2021; Neubauer et al., 2022).
The geological variability of the study area provides a valuable framework for reconstructing the region's complex structural history, specifically for constraining both the pre-Variscan and Variscan tectono-metamorphic evolution. The preservation of primary contacts between different rock types –such as those between paragneisses and magmatic intrusions– offers key insights into the relative timing of geological events. Additionally, the paragneiss types exhibit a heterogeneous composition, which results in highly variable mineralogy even at the outcrop scale, including evidence of metasomatic reactions between metapelites and intercalated carbonate-rich layers (von Raumer and Bussy, 2004).

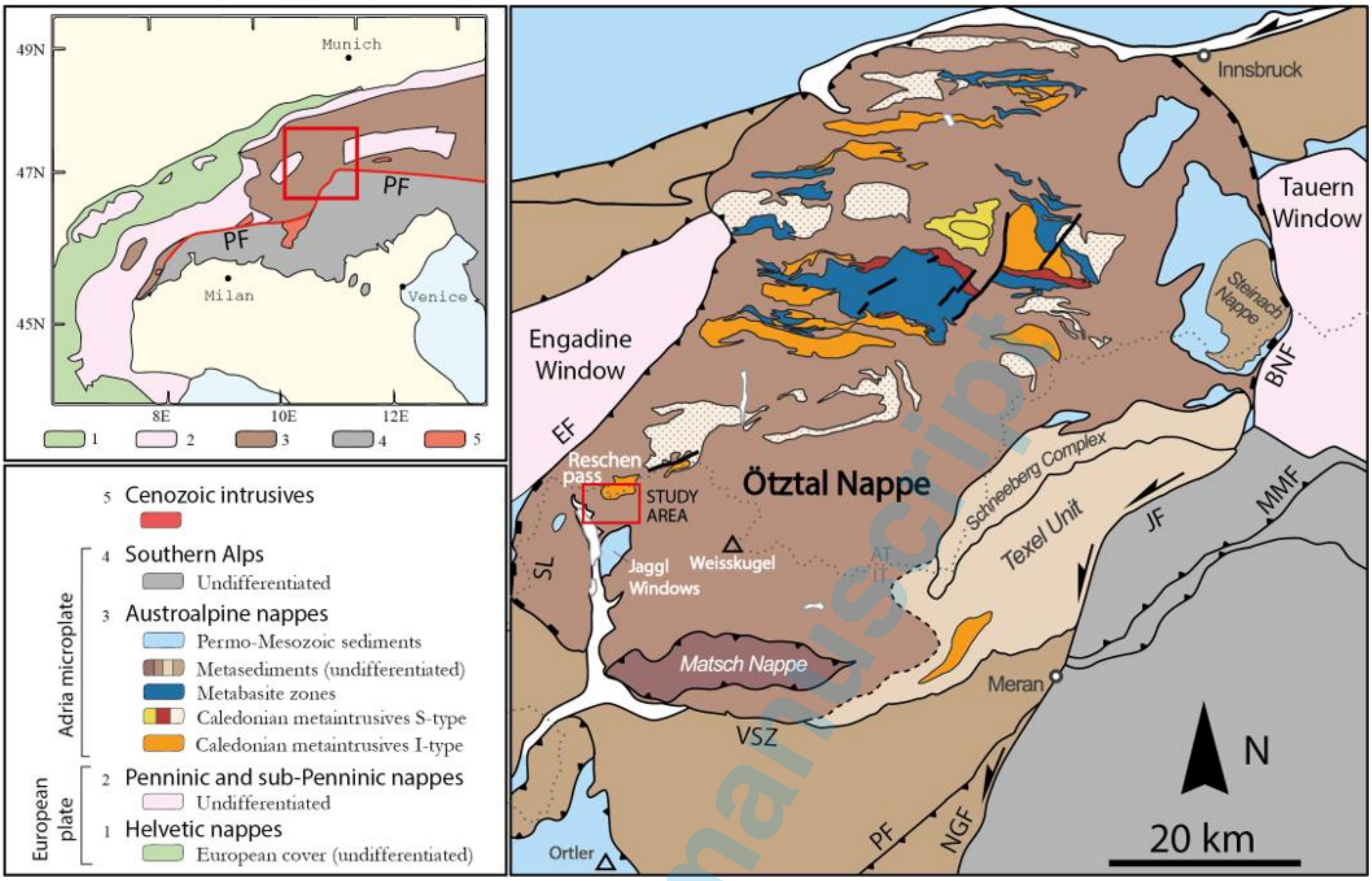

Fig. 1 - Upper-left, simplified tectonic sketch of the Alps (modified after Pfiffner, 2014); Right, lithological map of the Ötztal Nappe (modified after Schindlmayr, 1999). BNF: Brenner Normal Fault; JF: Jaufen Fault; MMF: Meran-Mauls Fault; NGF: North Giudicarie Fault; PF: Periadriatic Fault; VSZ: Vinschgau Shear Zone; SL: Schlinig Line; EF: Engadine Fault.

PETROLOGY AND METAMORPHIC EVOLUTION

The Ötztal-Stubai Complex is primarily composed of biotite-muscovite-plagioclase paragneiss and muscovite ± biotite micaschist, which locally contain kyanite ± staurolite ± garnet assemblages, developed during the Variscan metamorphic peak (550–650 °C and 4–7 kbar; Thöny et al., 2008). Sm/Nd dating on garnet and EMPA on monazite (CHIME method) in micaschist constrain this peak to 343 ± 1 to 331 ± 1 Ma (Hoinkes et al., 1997; Schulz et al., 2019), while post-peak Rb–Sr, K-Ar and Ar–Ar mica cooling ages range from 320 to 270 Ma (Thöni, 1981, 1986; Lichem, 1993; Schweigl, 1995).

The paragneisses are derived from highly heterogeneous Ediacarian to Cambrian metasediments (Siegesmund et al., 2021; Neubauer et al., 2022). Locally, calc-silicate bands –probably representing former marls– have also been observed, as in other sectors of the Alpine basement (Arnold, 1970; von Raumer, 1976; Zurbriggen et al., 1997; Schulz, 2008; “Heterogeneous gneisses”, Zanchetta et al., 2023; “Erlan-felses”, i.e., Cambrian and/or Neoproterozoic carbonate horizons, von Raumer and Bussy, 2004).

Amphibolites (± garnet ± epidote) and acidic orthogneisses may occur as layers (ranging from 10 to 1000 m thick) and bodies within the paragneisses. The intrusive precursors were emplaced during a high temperature pre-Variscan phase (670–750 °C and < 2.8 kbar, Thöny et al., 2008; 660-685 and >4kbar, Hoinkes, 1973 and Söllner et al., 1982; 570–630 °C and 7–9 kbar, Zurbriggen et al., 1997; 5 kbars, 8 kbar and ca. 750 °C, Klötzli-Chowanetz, 2001). The acidic orthogneisses occur in two types: one mainly containing K-feldspar and white mica, and another with a more granitic composition. Acidic orthogneisses have been dated at 435 ± 11 Ma (Rb/Sr whole-rock), 494 ± 73 Ma (Sm/Nd) and 455 ± 8 Ma (apatite whole rock; Hoinkes et al., 1997). On the other hand, the intermediate to mafic metaintrusives have been dated from 583 ± 78 Ma and 487±11Ma (Rb/Sr whole-rock dating on the Klopai tonalite; Schweigl, 1995; Klötzli-Chowanetz, 2001) to 480.5 ± 5 Ma (U/Pb zircon dating of the Fleres gabbro; Zanchetta et al., 2023). This high-temperature phase was linked by several authors to a period of crustal thinning leading to anatexis and the incipient formation of new oceanic crust related to the opening of the Rheic ocean (Bernhard et al., 1996; von Raumer et al., 2013; Neubauer et al., 2022), or to subduction-related magmatism during the Cenerian Orogeny (Zurbriggen, 2017; Dielforder et al., 2026; Žák et al., 2026), which may also have triggered partial melting of the metapelites (Verpeil, Winnebach and Nauders migmatites; Söllner et al., 1982; Klötzi-Chowanetz et al., 1997, Klötzli-Chowanetz et al., 2001; Thöny et al., 2008; Oester, 2017; Schaltegger et al., 2026).

All of these features occur in our study area. Here, the geology is dominated by Klopai Pluton, which is surrounded by different types of paragneiss ("Heterogeneous gneisses"). At the contact with the Klopai intrusion, the paragneisses exhibit signs of incipient melting. These partially molten rocks are referred to as "*Klopaier Metatexit*" (Schindlmayr, 1999), "Reschen migmatite" (Schweigl, 1995) or "Nauderer Gaisloch migmatite" (Thöny et al., 2008). However, the exact timing of anatexis remains debated, due to scattered radiometric data. Recent electron microprobe U–Th–Pb ages from monazite yield values between 431 ± 37 Ma to 472 ± 31 Ma (Thöny et al., 2008), while previous dating ranges from 430 ± 6 Ma to 585 ± 8 Ma (zircon growth, Schweigl, 1993; Klötzli-Chowanetz, 2001). These ages are broadly coeval with those of the other migmatites in the Ötztal Nappe [409 ± 50 Ma to 457 ± 42 Ma (U–Th–Pb monazite), Verpeil migmatites; 461 ± 4 Ma (Rb/Sr Wm) to 490 ± 9 Ma (U–Pb zircon) and 408 ± 46 Ma to 472 ± 36 Ma (U–Th–Pb monazite), Winnebach migmatites; Chowanetz 1991; Klötzli-Chowanetz et al., 2001; Thöny et al., 2008]. Within or near the Nauders Migmatite, fine-grained purple gneisses with a distinct spotted appearance are found ("*Scisti macchiettati*", Barattieri, 1994). These rocks are texturally similar to "*Fleckengneise*" of the Winnebach migmatite complex (Bidner, 1989), due to the presence of cordierite blasts. Given the wide age range for the granitoids and the neighbouring migmatite, spanning from 430 to 590 Ma (Söllner and Hansen 1987; Klötzli-Chowanetz et al.,1997; Klötzli-Chowanetz et al., 2001; Söllner, 2001; Thöny et al., 2008), it remains unclear whether the migmatization and cordierite formation in the paragneiss resulted from contact metamorphism due to the neighbouring intrusions or from an independent earlier regional high-temperature event.

Near the Klopai Pluton and its host rocks, pegmatite dykes are also observed, locally reaching several meters in thickness and cutting through both the metapelites and the magmatic rocks. These veins may represent fractionated products of the Klopai Pluton, potentially intruded during the late stage of magmatism. Pegmatites dated by Schweigl (1995) yielded scattered Rb–Sr ages from 474 ± 9 to 314 ± 6 Ma. More recent Sm/Nd ages suggest an emplacement age of around 458 ± 4.7 Ma (Thöni and Miller, 2004).

The western and southern portions of the study area expose weakly metamorphosed Permo–Triassic sedimentary rocks. These include a sequence of siliciclastic terrigenous sediments and a series of detrital carbonate rocks, along with lagoonal evaporite deposits overlying fossiliferous Anisian dolomites (the "Jaggl Window"; Richter and Zinkernagel, 1975). The contact between the cover rocks and the crystalline basement is typically tectonic and was interpreted either to be a tectonic window (Schmid and Haas, 1989) or as the displaced original sedimentary cover of the Ötztal Nappe (Froitzheim et al., 1997; Froitzheim et al., 2026).

METHODS

The area was surveyed on 1:10,000 scale using the classical field approach. The survey was carried out as part of Geological Sheet 'Malles/Mals' of the Autonomous Province of Bolzano. The topographic base map used for the survey is the IGM 1:25,000 topographic map with contour lines every 10 m superimposed from the digital terrain model of the Province of Bolzano. The geological map was compiled and georeferenced using the software ArcGIS Pro (© ESRI, Environmental System Research Institute) with the spatial reference system WGS 1984 UTM Zone 32N. The final layout was elaborated with the software Adobe Illustrator. Polished rock thin sections (30 μm) were prepared and investigated under transmission optical microscope for the petrographic characterization of the mapped lithologies. All mineral abbreviations are after Whitney and Evans (2010).

LITHOLOGIES IN THE SURVEYED AREA

The surveyed area is bounded to the South by the Langtaufers Valley, to the North by the Austria–Italy border, to the West by the Vinschgau Valley, and to the east by the ridge leading to Wölfeleskopf (Fig. 2). Geological units have been distinguished based on literature data, lithological associations, deformational history, and metamorphic grade, as inferred from characteristic mineral assemblages. A detailed description of all these lithologies is given below.

Lithological units of Permo-Mesozoic sediments

At the southern limit of the study area, along the bottom of the Langtaufers Valley on its right bank, outcrops of chloritised meta-sandstone and dolomitic marble are exposed (Fig. 2). The outcrops are often small in size and the rocks are deformed, showing both ductile structures (folds) and brittle features (cataclasites). They show a low-grade metamorphic overprint (greenschist facies). The metasediments likely represent the southwestern

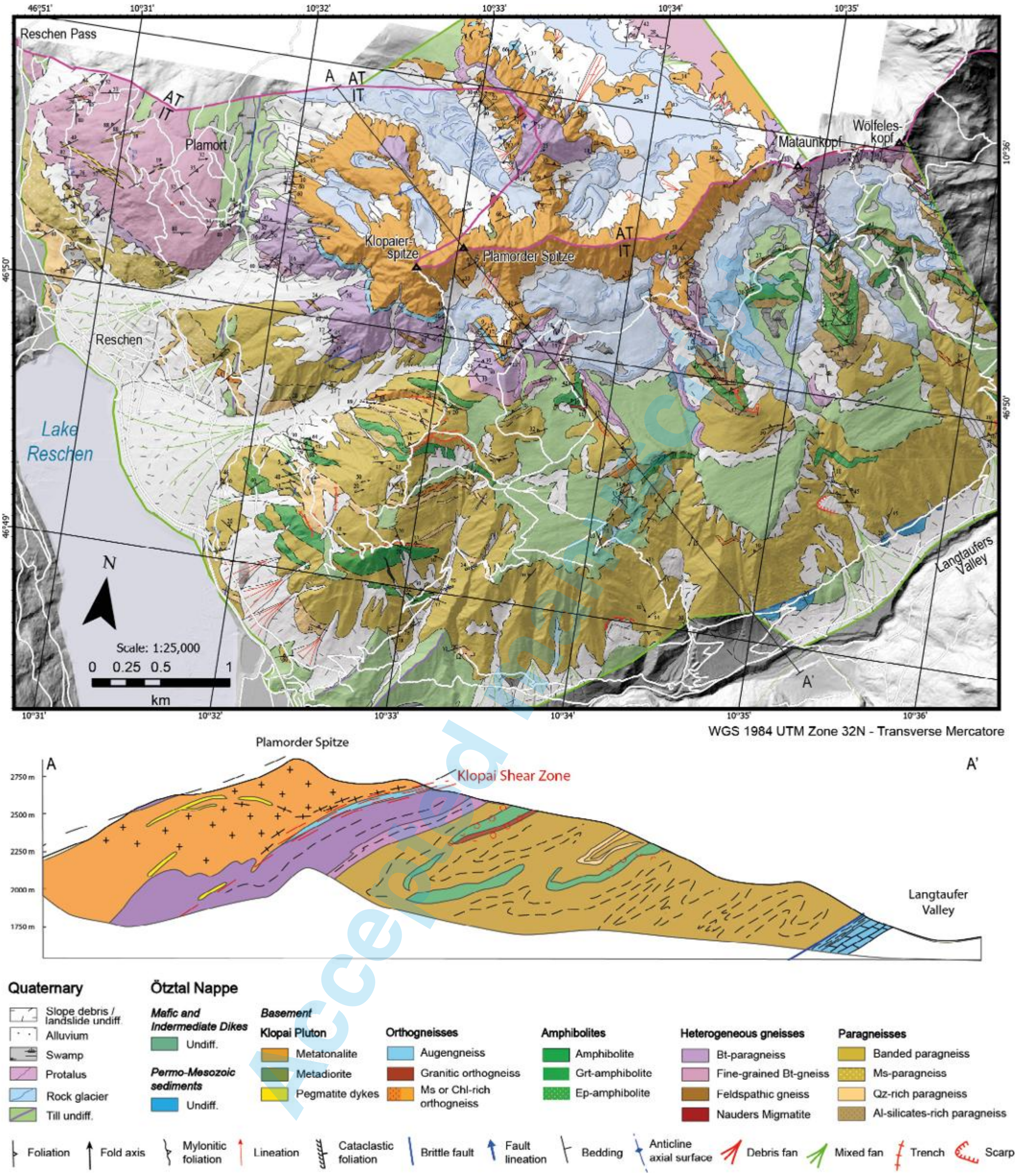

Fig. 2 - Simplified geological map of the Klopai Pluton and its surroundings. In the profile, crosses represent the undeformed portion of the pluton while elongated crosses and dashed lines mark the foliation trend.

termination of the Permo-Mesozoic sediments of the Jaggl Window (Fig. 1; see Section 3). The contact between these sedimentary rocks and the overlying Ötztal Nappe is tectonic in nature.

The meta-sandstone is sage-green, slightly silvery quartzitic schists, characterised by millimetric layers of chlorite and sericite. They crop out above dark grey carbonates (limestones and dolostones) with bedding thicknesses of approximately 10–15 cm. In contact with the meta-arenite, the strata become thinner and yellowish bands appear.

Lithological Units of the Ötztal basement

At the northern slopes of the Langtaufers Valley the base of the Ötztal Nappe mainly consists of banded paragneiss with more competent quartz-rich levels. The paragneiss typically contains plagioclase, quartz, biotite and minor muscovite. Further upslope, the paragneiss becomes increasingly micaceous and locally contains bands rich in kyanite and staurolite, which grow statically over the regional foliation. In this upper section of the slope, amphibolite layers become more abundant, comprising up to 40% of the rock mass. Garnet is present in both the paragneiss and occasionally in the amphibolite layers, especially near their contact zones.
Near the ridge marking the northern boundary of the surveyed area -coinciding with the Austria-Italy border- and in contact with the small intrusive body of the Klopai Pluton, a variety of lithologies appear in smaller proportions. These include biotite-rich feldspathic paragneiss, biotite-schist and fine-grained biotite-paragneiss. The occurrence of zoned calc-silicate layers and carbonate remnants is a common feature to all these biotite-rich rocks, which are grouped into the "Heterogeneous gneisses" (Zanchetta et al., 2023). A shear zone marks the contact between the "Heterogeneous gneisses" and the Klopai Pluton. This intrusion, covering an area of 6.5 km^2, is largely composed of tonalite and minor diorite that locally preserve their original magmatic texture. On the Austrian side, the upper contact of the Klopai Pluton with the host rock is locally preserved with primary magmatic features. Metatexite is in contact with the Klopai Pluton, as indicated by fragments of metatexite now found as xenoliths in the tonalite. Both the Klopai Pluton and the host rocks are cut by pegmatite dykes.

Paragneisses

<u>*Banded Paragneiss*</u>
In the lower part of Langtaufers Valley fine- to medium-grained banded paragneiss show orange-brown oxidised weathering crusts (Fig. 3A). They are mainly composed of quartz, plagioclase, and biotite, with minor muscovite. They often contain late stage stretched lenses of quartz, kyanite, and andalusite. In the lenses, the crystals of kyanite and andalusite are centimetre-sized and often substituted by white mica. Garnet is rare and small, with a maximum diameter of 2–3 mm.

<u>*Muscovite Paragneiss*</u>
Within the banded paragneiss, layers of grey, medium-grained paragneiss mainly composed of quartz, plagioclase, muscovite, and minor or absent biotite occur. Garnet can be an accessory mineral. Sometimes, this paragneiss appear less weathered and more compact than the banded ones. They break into angular, centimetre-sized fragments.

<u>*Quartz-rich Paragneiss*</u>
Similar to the banded paragneiss, but with a higher quartz content. At the outcrop scale, they have a more compact appearance, being less weathered. Quartz-rich paragneiss appear both as thin, un-mappable layers within the banded paragneiss and as thicker, meter-thick beds that form small walls along the slopes.

<u>*Al-silicates-rich Paragneiss*</u>
This lithology has a similar field aspect to the banded paragneiss, however they can contain locally abundant garnet, minor staurolite and kyanite (Figs. 3B, 4B). It is found in the higher part of the slope, from 2000-2200 meters up to the contact with the biotite-rich gneisses and schists. This paragneiss is mainly composed of plagioclase, quartz, biotite, garnet, muscovite, ±kyanite, ±staurolite. Staurolite, kyanite, and possibly sillimanite, are considered syn- to post-kinematic and were generated during the Variscan p-T evolution. Garnet appears close to the contacts with amphibolites where it usually becomes more abundant and larger (up to 3 cm), suggesting a reaction between the two lithologies. Moreover, garnet displays biotite coronas, likely due to a retrograde reaction with potassium-rich fluids (Clarke and Rottura, 1994). The boundary between this lithology and the banded paragneiss is gradational and therefore uncertain.

<u>*Amphibolites*</u>
Dark green amphibolite occur intercalated within the paragneiss, as tabular bodies with thicknesses ranging from about one meter to several tens of meters, concordant with the foliation (Figs. 3A, 4A). Amphibolite layers extend along the entire southern slope of the Klopaierspitze but are more abundant above 2000-2200 m, within the Al-silicates-rich paragneiss. They contain plagioclase and amphibole ± epidote. The texture is either banded or massive. When in contact with paragneiss, small garnet porphyroblasts of a few millimetres up to 1 cm in size appear. As in the paragneiss, garnet has biotite coronas. Amphibolite often show epidote ± carbonate veins.

<u>*Orthogneisses*</u>
Several types of orthogneisses crop out in the area. The first type bounds the bottom of the Klopai Pluton. It is a medium- to coarse-grained augengneiss. It is rich in porphyroclasts of K-feldspar, varying in size from a few millimetres up to 6-7 cm, quartz, plagioclase, muscovite and biotite. The K-feldspar can be from whitish to pinkish in color. The second type is a medium- to coarse-grained

orthogneiss with pink K-feldspar porphyroclasts, quartz, plagioclase and chlorite or muscovite. The third type is a medium-grained granitic “augengneiss”, with K-feldspar porphyroclasts up to 1-2 cm, quartz, plagioclase, biotite and muscovite.

Heterogeneous gneisses

Biotite paragneiss and schist

This unit includes folded, banded purple schist and gneiss with evident compositional layering: quartz-plagioclase, mica or amphibole-rich layers from few millimetres to decimetres in thickness. They contain dominant biotite, plagioclase and quartz, ± white mica, ± garnet ± amphibole, ± chlorite. Garnet crystals are locally abundant but usually indistinguishable to the naked eye. Tourmaline occurs as an accessory mineral. Calc-silicate bands are present and display a clear zoning from edge to core (Figs. 3C, 4C). The core contains associations of garnet, clinopyroxene, epidote (allanite and zoisite), rare K-feldspar, and quartz, associated with granular remnants of calcite. Garnet often give the vein cores a pink-orange colour. At the edge, biotite, plagioclase, quartz, and small garnets are present. The intermediate zone, which is green in colour, mainly contains amphibole, quartz, and white mica.

Fine-grained biotitic gneiss

Dark purple rocks, very fine grained and massive or poorly foliated, primarily composed of quartz, plagioclase, biotite, and minor muscovite (Figs. 3C and D, 4C and D). They break into medium-sized blocks with sharp edges. They contain calc-silicate rich bands. Sometimes, they display prismatic, stubby, dark spots composed of fine-grained aggregates of biotite, white mica, and rare relics of cordierite (Figs. 3D, 4D). In the literature, these rocks were often referred as speckled gneisses (“*Scisti macchiettati*”, Barattieri, 1994; “*Fleckengneisen*” Bidner, 1989). In the field, the dark aggregates appear to be overprinted by the foliation. The fine-grained biotitic gneisses are cut by pegmatite dykes.

Feldspathic gneiss

Within the biotite gneisses and schists, layers of a coarse grained paragneiss characterised by discontinuous domains rich in biotite and white mica, 1-2 cm in size plagioclase poikiloblast, and quartz occur (Figs. 3E, 4E). This gneiss also contains garnet, rare kyanite, and sillimanite (fibrolite variety). It contains rare K-feldspar.

Klopai Pluton

Metatonalite

The dominant lithology of the Klopai Pluton is a tonalitic orthogneiss, undeformed toward the core of the intrusion, with a texture ranging from gneissic to massive phaneritic, with fine to medium grain size. It is predominantly composed of plagioclase, quartz, dark green amphibole (hornblende, with the typical prismatic habit), brown biotite and minor K-feldspar, ± chlorite and sericite. Zircon and apatite are accessories mineral. The quantitative ratio between biotite and amphibole is not constant. The pluton crops out along the entire ridge of the surveyed area that marks the border with Austria. Within the tonalite, there are also enclaves of more mafic bodies (diorites), as well as centimetre-sized xenoliths of host rocks (“Heterogeneous gneisses”; Figs. 3F, 4F). The inclusions range from a few centimetres to one or two meters in diameter, with their aspect ratio varying depending on the degree of ductile deformation of the rock. A more detailed description can be found in Schweigl (1995). At the base of the Klopai Pluton, there are discrete 1 m thick layers of mylonite (hereafter refered to as the Klopai Shear Zone), where amphibole laths and mafic inclusions oriented and elongated in the shear direction can still be recognised. In other cases, the mylonite is too fine-grained to observe anything but grey-green compact layers.

Metadiorite

The Klopai Pluton includes small bodies of dark green-grey diorite (a few tens of metres in size), often undeformed. They contain mainly amphibole and plagioclase but lack of biotite. Due to incomplete magma mingling, these mafic bodies are also found as dark rounded enclaves within the tonalite.

Pegmatite dykes

Pegmatite dykes, ranging in thickness from a few centimetres to several metres, contain quartz, muscovite, K-feldspar, ± plagioclase, ± garnet, and ± tourmaline (Figs. 3G, 4G). Texture ranges from undeformed to highly foliated. When deformed, tourmaline forms black rounded aggregates.

Nauders Migmatite

In the surrounding area of the Klopai Pluton, the “Heterogeneous gneisses” often exhibit a migmatitic appearance (Figs. 3H, 4H). They occur in different textures: vein-type migmatite, nebulite, rafts, and schlieren (Sawyer, 2008). Specifically, in outcrop, the raft texture is predominant, showing often the dark, fine-grained and schistose melanosome embedded within a volumetrically dominant, fine-grained, leucocratic portion. The leucosome contains quartz, biotite and minor muscovite, plagioclase and K-feldspar is found only in small films at the grain boundary (Thöny et al., 2008). Accessory minerals are zircon, kyanite and sillimanite.

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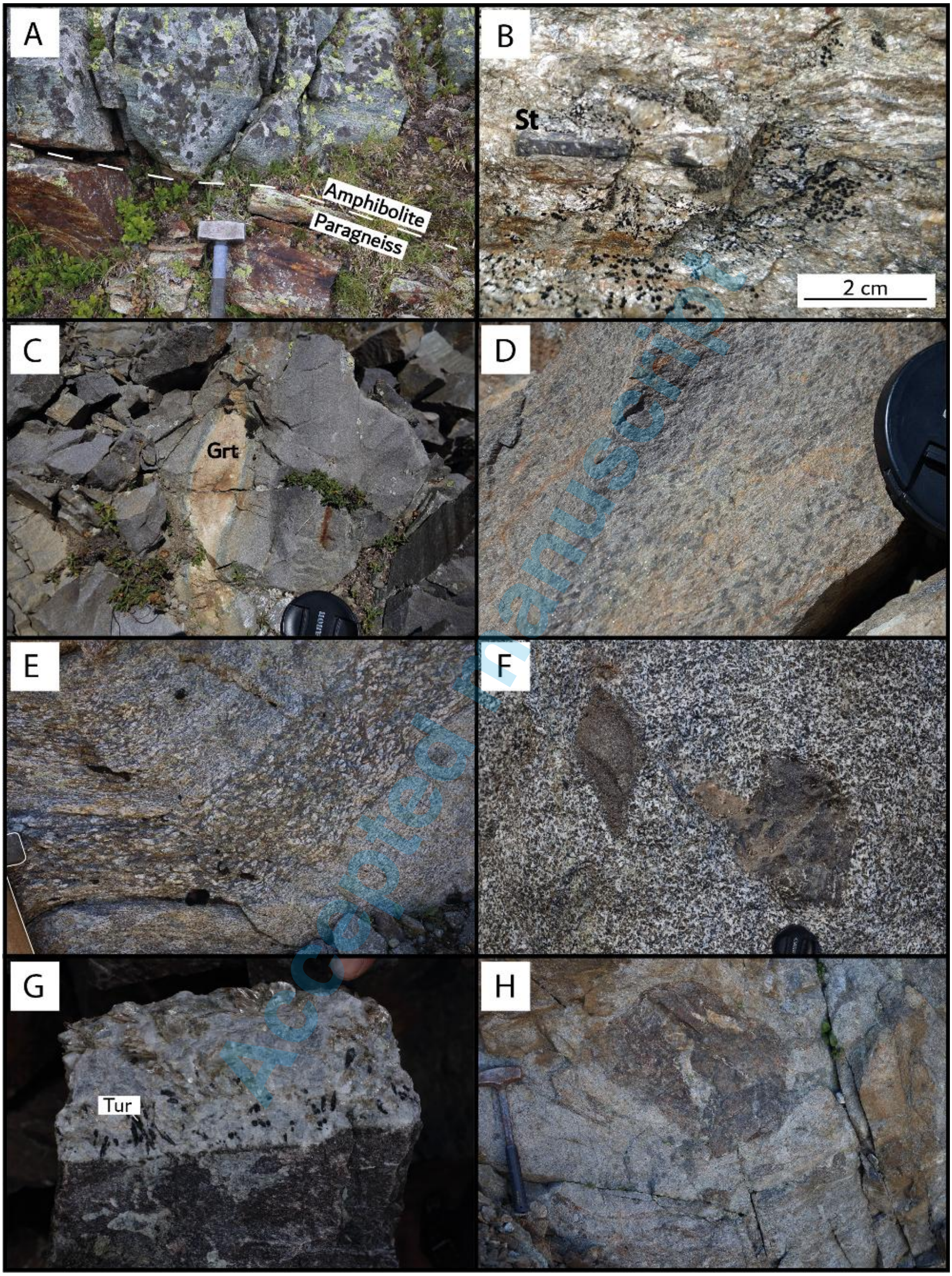

Fig. 3 - (A) Epidote-bearing amphibolite in contact with banded paragneiss; (B) Staurolite in Al-silicates-rich paragneiss; (C) Fine-grained biotitic gneiss with a calc-silicate-rich band; (D) Cordierite relict in fine-grained biotitic gneiss; (E) Feldspathic gneiss; (F) Klopai tonalite with xenoliths of host "Heterogeneous gneisses"; (G) Undeformed pegmatite dyke, (H) Nauders Migmatite.

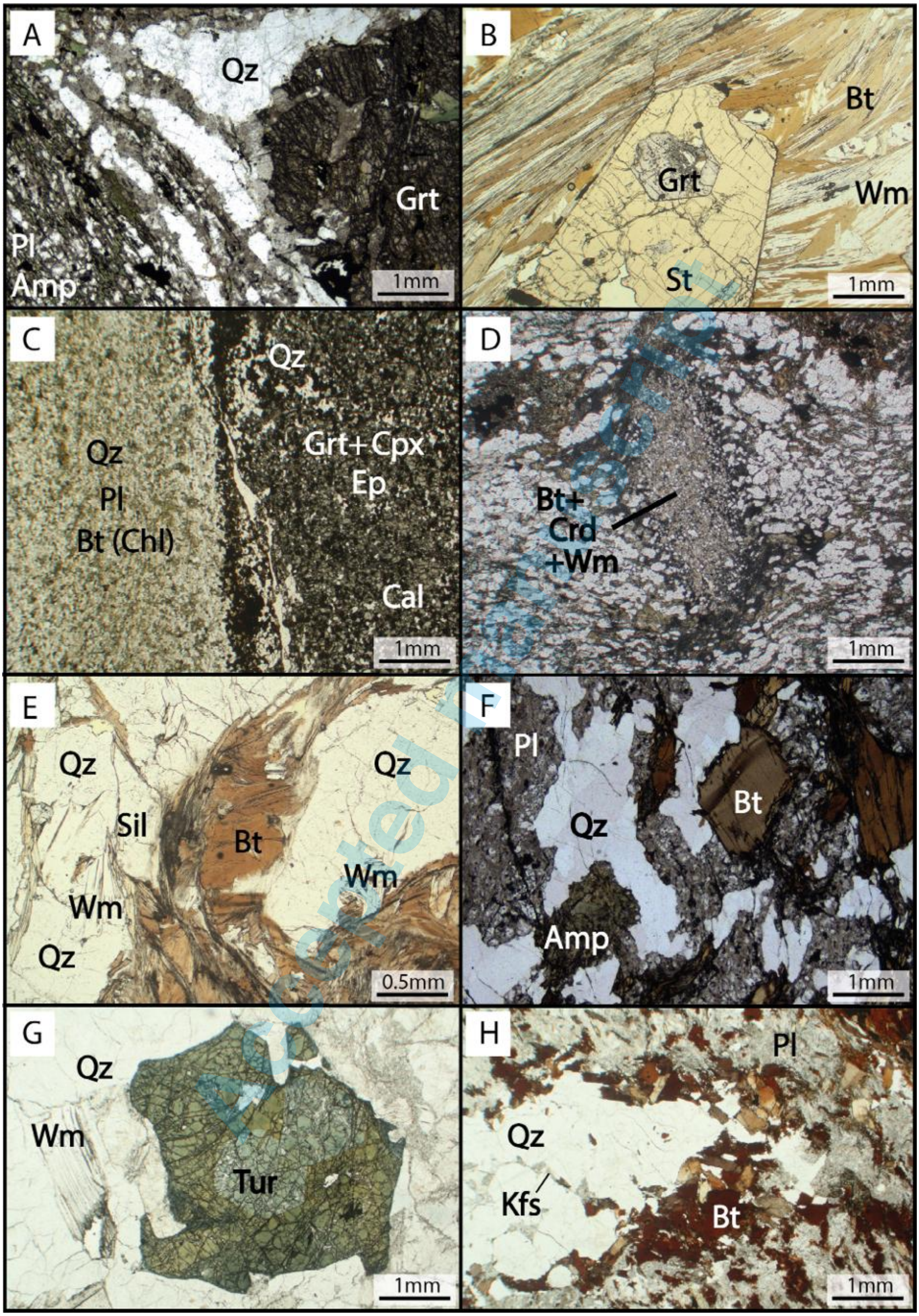

Fig. 4 - (A) Garnet-bearing amphibolite; (B) Staurolite and garnet in Al-silicates-rich paragneiss; (C) Fine-grained biotitic gneiss (left) with a calc-silicate-rich band (right); (D) Cordierite relict in fine-grained biotitic gneiss; (E) Feldspathic gneiss; (F) Klopai tonalite; (G) Tourmaline in undeformed pegmatite dyke, (H) Nauders Migmatite.

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Mafic and intermediate dykes

Dark green dykes of intermediate composition colour, with a porphyritic texture of few millimetre-sized plagioclase phenocrysts and quartz aggregates, in a matrix with submillimetric mafic minerals, have been found in the area. Their thickness ranges from 1 to 10 meters. All dykes intrude parallel to the foliation but preserve their original magmatic texture. Raso et al. (2025) studied a dyke (plagioclase + muscovite + chlorite + epidote + titanite + amphibole) from the Langtaufers Valley and constrained the P-T condition for the Eoalpine metamorphism between 300-400 °C and 4-6 kbar.

STRUCTURES IN THE SURVEYED AREA

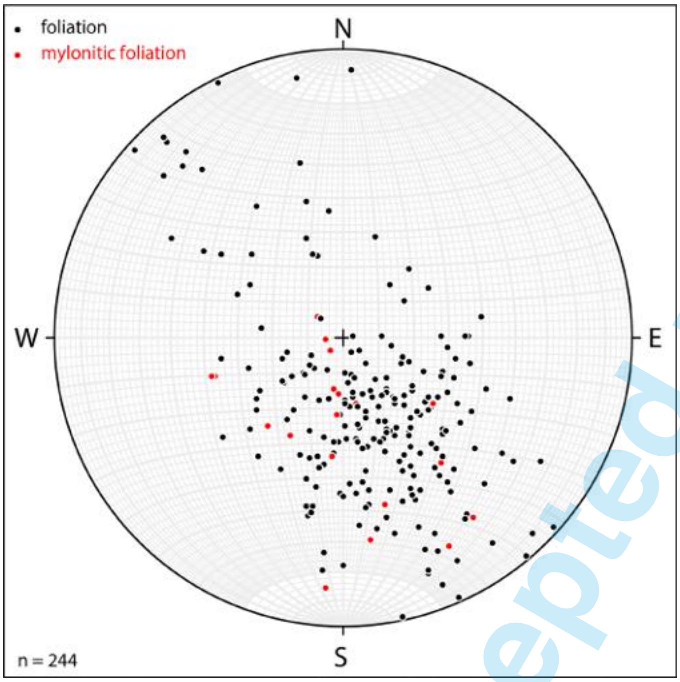

Fig. 5 - Poles to foliation and mylonitic foliation in the study area. Lower hemisphere, Schmidt Net.

The structural evolution of the southern part of the crystalline basement of the Ötztal Nappe was studied by van Gool et al. (1987), who documented four deformation phases. The Reschen Pass area is characterised by a pervasive ca. E–W striking foliation (S_1–S_2), which mostly dips N–NW at 10–50° (Fig. 5). The main foliation is also axial planar to isoclinal and intrafolial folds (F_2), well visible within the micaschist and paragneisses (Fig. 6A). The fold hinges are scattered, and plunge mainly to the N-NE but also to the S. The scattered appearance is caused by the refolding of the main foliation by F_3 open to close folds at all scales. The smaller scale F_3 folds are well visible in both amphibolites and paragneisses. F_3 folds are characterised by sub- horizontal hinges slightly plunging to the NE and sub-vertical or more recently inclined axial planes (Fig. 6B). F_3 folds can have a chevron appearance. Thöni (1988) pointed out that D_3 records a lower temperature deformation compared to D_2. The overprinting of the isoclinal F_2 folds by F_3 folding led to a mixed type 2/3 fold interference pattern (Ramsay and Huber, 1987; Egglseder and Fügenschuh, 2013).

In the area north of the Klopaierspitze, the contact between the pluton and the host rock ("Heterogeneous gneisses") is deformed by F_3 open folds, some of which reach tens of meters in size, with fold axes oriented approximately E–W, indicating a north–south compressional event (Fig. 6 B). Pegmatite dykes cutting through the tonalite are also folded mimicking the geometry of the contact. The timing and the presence of a pre-D_2 foliation remains debated; it may be related to the intrusion of the Klopai Pluton and depends also on whether the pluton intruded into a structured basement of pre-Ordovician–Silurian age.

The best-exposed and most continuous shear zone in the mapped area is the Klopai Shear Zone, located at the contact between the Klopai tonalite and the underlying "Heterogeneous gneisses". In this area, mylonites are localised in discrete bands ranging in thickness from meters to a few tens of meters and extending over several kilometres in length (Barattieri, 1994). The shear zone mainly involves the "Heterogeneous gneisses", the Klopai tonalite, and the pegmatite dyke. In some cases, the lithologies appear interfingered, with the presence of thin slivers and very narrow bands. The mylonites exhibit an orientation consistent with the regional foliation (Fig. 5) and E-W oriented lineations.

D_2 and D_3 structures record deformation related to the Variscan tectono-metamorphic event, whereas Eoalpine-related structures are visible only outside and south of the study area, in the proximity of the top-to-the-west-directed thrusting along the Vinschgau Shear Zone (see van Gool et al., 1987; Egglseder and Fügenschuh, 2013; Montemagni et al., 2023). In fact, the Eoalpine Cretaceous orogenic cycle includes various W to NW thrusts within the Austroalpine units, followed by a top-to-the-southeast extensional overprint. Whether the mylonites at the base of the Klopai Pluton (Klopai Shear Zone) are Eoalpine or related to the previous Variscan phase remains unclear.

Folds related to Neoalpine N-S collision (F_4) are rare in the study area (Fig. 6B), whereas brittle structures are well preserved (D_5, Fig. 7). In the area it is recorded by a series of brittle normal and strike slip faults oriented mostly NE-SW and NW-SE (Agliardi et al., 2008, 2009).

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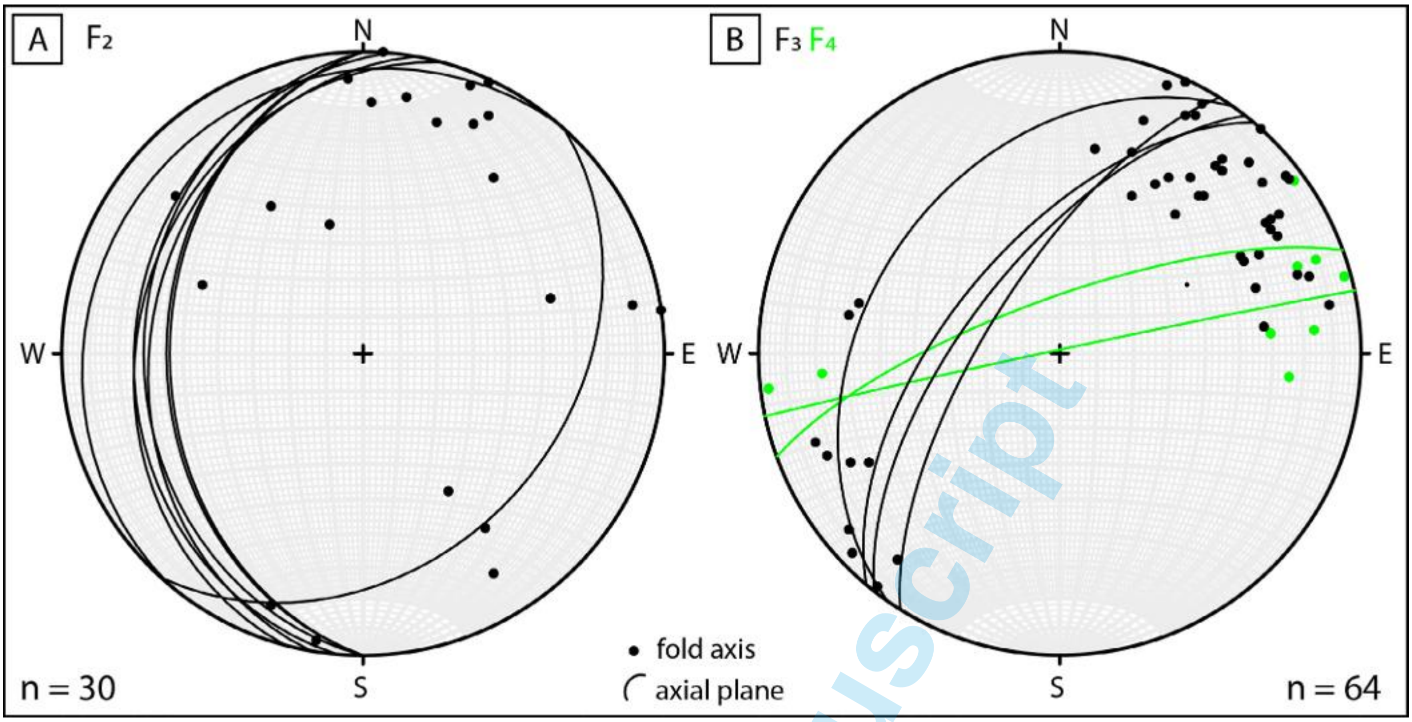

Fig. 6 - (A) axial planes and hinge lines of isoclinal F_2 folds, (B) axial planes and fold axis of F_3 (black) and F_4 (green) folds. Lower hemisphere, Schmidt Net.

DISCUSSION AND CONCLUSIONS

Though the area of the Klopai Pluton was represented in previous maps by Hammer (1925) and Schweigl (1995), this work provides a more accurate description of genetic and structural relationships, necessary for the understanding of the complex tectono-metamorphic history of this part of the Austroalpine basement. In addition, the higher level of detail –in terms of structural features and lithological units– compared to the 1:100,000 "Tre Venezie" geological map (Hammer, 1925), will be a useful starting point for the future realization of the CARG geological sheet of the area, our new map also highlights important differences with the interpretation of Schweigl (1995). Specifically, migmatites do not occur within the Klopai Pluton itself but are only distributed along the contact zone, while biotite- (and cordierite-) bearing rocks, which according to our field observation mark the boundary between the pluton and the enclosing paragneiss, suggest a genetic relation with the pluton, possibly delineating a pre-Variscan contact metamorphic aureole.

Paragneisses, amphibolites, and orthogneisses in the study area exhibit a complex metamorphic-deformation history, as is typically observed in the Ötztal basement, which has undergone at least three orogenic cycles. The oldest event, dated to the Ordovician, corresponds to the Klopai intrusion and the metamorphic aureola within the surrounding host rocks. Migmatites, the pluton itself, and pegmatite dykes appear genetically related: migmatites are found exclusively at the contact with the pluton, while pegmatite dykes are restricted to the pluton and its close surroundings. Despite the scattered ages available for the Klopai tonalite (583 ± 78 and 487 ± 16 Ma; Schweigl, 1995 and Klötzli-Chowanetz, 2001), the ages of migmatites (431 ± 37 Ma to 472 ± 31 Ma; Thöny et al., 2008) constrain the magmatic event to its youngest interval. Pegmatite dykes constrain the end of the magmatic activity at ca. 458 ± 4.7 Ma (Thöni and Miller, 2004), as they crosscut the intrusion, the migmatites and the surrounding host rocks. This timeframe is further supported by dating of analogous intrusive bodies in the surrounding region, suggesting a regional magmatic event between 480 and 460 Ma (Zantedeschi, 1991; Schulz et al., 2004; Zanchetta et al., 2023; Klug et al., 2025). For this reason, we favour the younger age for the emplacement of the Klopai Pluton.

The emplacement age of the Klopai Pluton is similar to the ages of migmatites and pegmatite dykes, but the ~20 Ma time span and the large analytical uncertainties prevent a more precise reconstruction of the temporal sequence. Specifically, it remains unclear whether the migmatites formed synchronously with the pluton or are slightly older. The presence of migmatite blocks within the pluton and at its contact suggests that they are not younger than the pluton. More precise geochronological

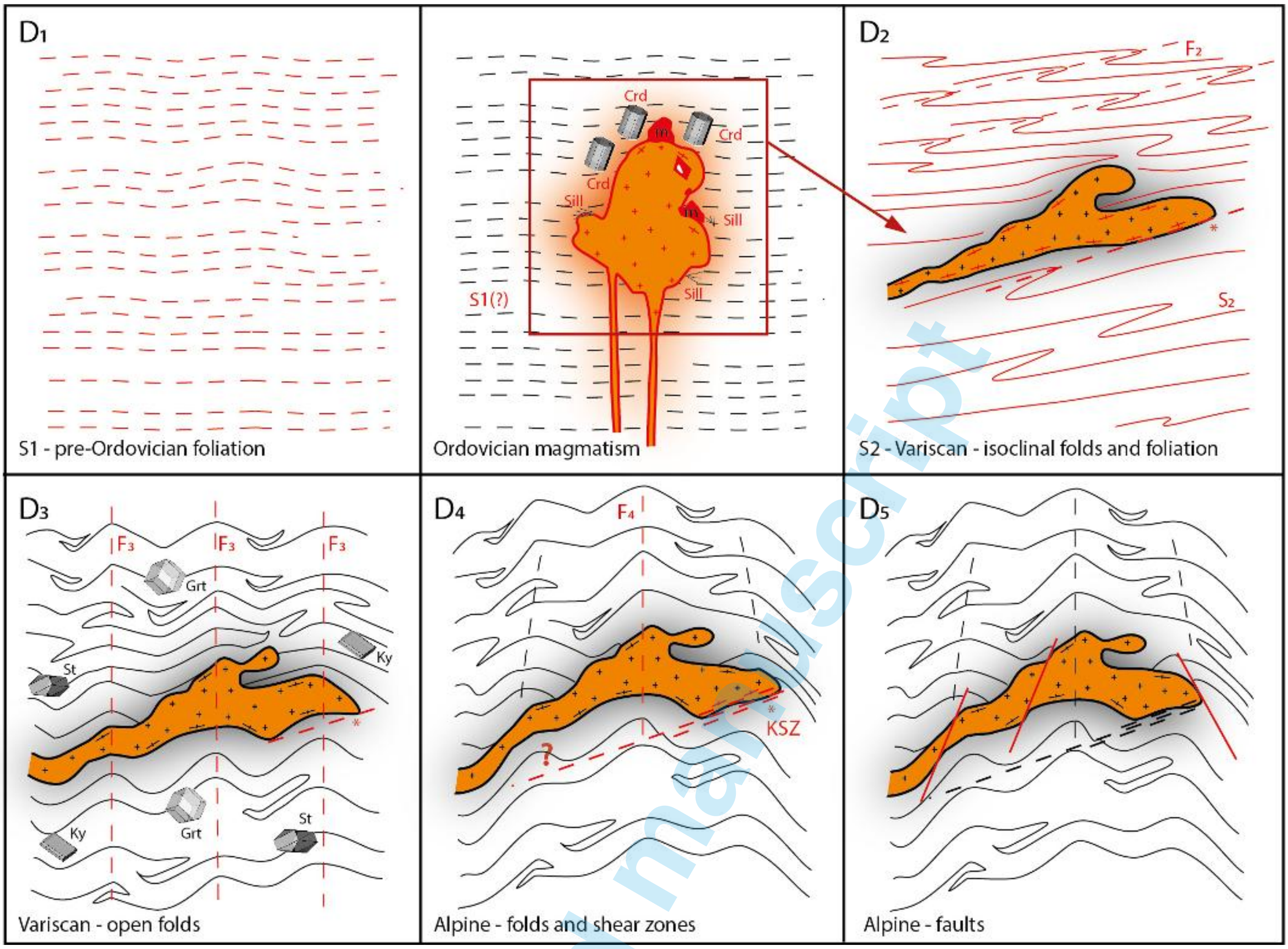

Fig. 7 - Schematic representation of the tectono-metamorphic evolution of the study area. In red, new structures; in black, old structures. The halo around the intrusion marks the metamorphic aureole. The Klopai Shear Zone (*KSZ) can be D_4 or older (Variscan).

data and detailed geochemical studies are needed to determine whether the migmatites resulted from a high-grade regional thermal event associated with extension or were generated by the intrusion of the pluton itself. The presence of cordierite in the contact aureola provides a P–T constraint for the emplacement environment, indicating relatively shallow intrusion conditions (<10 km depth).

The D_2 deformation phase of Variscan age corresponds to the development of the main foliation, which is associated with high-grade metamorphic minerals such as kyanite and staurolite. These minerals typically grow statically on the foliation during the Variscan metamorphic peak. However, not all paragneisses contain this type of mineral assemblage. In fact, high-grade minerals are absent in the paragneisses located closer to the valley bottom. In the absence of shear zones between the lower and upper parts of the valley, we suggest that the lack of Al-rich silicates in some areas is more likely related to the bulk chemical composition of the original sediments. For instance, although garnet is generally forming during high-grade regional Variscan metamorphism, in the study area it occurs exclusively at the contact between amphibolites and paragneisses. Biotite coronas around garnet would be compatible with cooling after the Variscan metamorphic peak. The D_3 phase folds both the main foliation and the contact between the Klopai Pluton and its host rocks. The D_4 deformational phase is poorly documented; only two structures may be attributed to it. The first is the Klopai Shear Zone beneath the Klopai Pluton. However, its age remains unconstrained and it may even predate D_4, possibly corresponding instead to D_2 (or D_3). The second is the tectonic contact between the Permo–Mesozoic sediments occurring at the base of the paragneisses, near the bottom of the Langtaufers Valley. However, this tectonic contact is never directly exposed in outcrop.

The D_5 stage is only expressed by discrete brittle faults related to the collapse or fragmentation of the orogen. In fact, the Neogene cycle corresponds to the lateral extrusion and exhumation of the Tauern and Engadine Window (Ratschbacher et al., 1991; Evans, 2011), or to the more recent activity of the Glorenza Fault (for a thorough discussion see Agliardi et al., 2008, 2009). However, this interpretation is not fully constrained, and the presence of older, pre-Miocene faults (i.e., latest Cretaceous to Paleogene) cannot be excluded.

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