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Crustal Shortening Mechanisms and Accretionary Prism Architecture in the Southern Apennines (Cilento–Mt Sirino), Italy

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Abstract

This study investigates the orogenic evolution of some key areas in the southern Apennines. The work focused on three key geological units: (i) the Lagonegro unit in the Mt. Sirino area (Basilicata region); (ii) the oceanic wedge of Ligurian Accretionary Complex (LAC); and (iii) the Cilento Group deposited in a wedge-top basin above the LAC. The latter are both studied in the Cilento area (Campania region). The Lagonegro unit exposed in the Tempa Roccarossa area hosts a tectonic *mélange*, formed in a shear zone associated with a regional thrust, that exhibits multiple deformation stages, including brittle and ductile shear, with a burial temperature range of 200°C–250°C (anchimetamorphism), corresponding to depths of 10–13 km. Kinematic vorticity and strain analyses indicate a dominant simple shear component with a secondary pure shear deformation, and a general flattening strain. The LAC shows polyphase deformation, characterized by veins, stylolites, cleavages, folds and faults. Thermal analysis using the Kubler Index (KI) and Raman spectroscopy indicates temperatures of 250–280°C (anchimetamorphism) and pressures of 0.3–0.4 GPa, corresponding to a burial depth of 11–15 km. This suggests a geothermal gradient of 17–25°C/km. Kinematic and strain analyses indicate similar features of the LAC deformation. In the Cilento Group basin, paleocurrent analysis reveals complex turbidite current flow directions, with sediment sources including the LAC, the overriding plate (Calabria-Peloritani Terrane), and volcanoclastics from synorogenic volcanic arcs. The study reconstructs the tectonic collapse of the Ligurian Accretionary Complex and the development of the wedge-top basin, indicating a transition from compressional to extensional tectonics. The integration of structural, petrographic, and geothermometric data provides new insights into the thermal history, deformation mechanisms, and sedimentary evolution of the Cilento region in the southern Apennines, contributing to a broader understanding of accretionary wedge formation at active plate margins. These findings enhance our understanding of the region's tectono-metamorphic and deformational evolution and provide a framework for understanding the interaction among fluid migration, shear zone dynamics, and sedimentation during orogenic processes. By applying these results to other active tectonic settings, the study offers a tool for regional tectonic reconstruction and helps clarify the processes governing accretionary wedge and orogenic mechanisms.

PREFACE

This thesis is a “thesis with publication,” consisting of seven chapters (including the introduction and concluding remarks). It is based on two publications and one manuscript (under preparation).

To better organise the thesis, each chapter's title primarily addresses a specific research topic. Chapter one offers a general introduction and problems/research questions, chapter two includes a detail description of the regional geology, chapter three include material and methods. Chapter four consists of one section on the Tempa Roccarossa mélange. Chapter five concerns the Ligurian units in Cilento and is based on three locations and outcrops: similarly, chapter six concerns the Cilento group deposits. Each of these chapters represents a manuscript; chapters two and four are already published, and a manuscript from chapter three is under preparation.

Finally, concluding remarks (chapter seven) are presented to provide a comprehensive overview of all three study areas, and a unique references list is provided at the end.

Dedications

To my supervisor, whose guidance, patience, and wisdom have been the foundation of this achievement, and my parents.

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Chapter 1 Introduction

1.1 Introduction

The present PhD thesis is based on an integrated geological study conducted in the Mt. Sirino (Basilicata region) and Cilento (Campania region) areas of the southern Apennines (Italy). This research focuses on three key geological units: (i) the Lagonegro-Molise Basin (LMB); (ii) the Ligurian Accretionary Complex (LAC), and (iii) the wedge-top basin deposits of Cilento Group (CiG). Together, these units provide a natural laboratory for investigating the tectono-stratigraphic evolution and deformation mechanisms associated with collisional orogenic processes, accretionary prism development, and syntectonic sedimentation in wedge-top basin settings.

The southern Apennines represent a sector of the Apennine–Maghrebide orogenic belt and offer a unique opportunity to study the complex interplay between thrust tectonics and sedimentation. In particular, the Tempa Roccarossa tectonic *mélange* of LMB (Mt. Sirino) provides key evidence for deformation associated with the crucial orogenic stage of the detachment and stacking of the LMB. On the other hand, the LAC (Cilento) recorded a multiphase deformation history associated with the first stages of obduction and frontal accretion of oceanic and transitional crustal materials. Further, the Cilento wedge-top basin deposits, consisting mainly of syntectonic turbidites and pelitic sequences, recorded the sedimentary response to progressive thrust migration during the Miocene and associated synorogenic extension associated with the allochthonous wedge collapse.

The rationale for this research is to better understand the structural architecture, deformation history, and evolution of these three interconnected geological units. While previous studies have focused individually on structural or sedimentological aspects of the southern Apennines, a comprehensive approach integrating meso- and microstructural analysis, strain quantification, and geochemical characterization across multiple tectonic settings is still lacking. By combining field observations with laboratory data, this research aims to fill that gap, providing new insights into the tectono-metamorphic processes that shape accretionary wedges and *mélange* complexes.

One of the most challenging aspects of understanding such systems lies in unravelling the mechanisms of strain localisation and the evolution from brittle to ductile deformation. In accretionary oceanic prisms like the LAC, deciphering the strain patterns and kinematics of

deformation is critical for reconstructing the regional tectonic evolution and for constraining the role of shear zones in accommodating large-scale tectonic displacements.

This study adopts a multi-scale analytical framework, beginning with detailed geological and structural surveys of selected sectors in Mt. Sirino and Cilento area, followed by targeted sampling of representative rocks across deformation gradients. Microstructure and strain analysis, kinematic vorticity estimates, and the identification of shear sense indicators in thin section provide constraints on the deformation regime and kinematic evolution of the major shear zones. Additionally, X-ray powder diffraction (XRPD), X-ray fluorescence (XRF), scanning electron microscopy with energy-dispersive spectroscopy (SEM-EDS), and chlorite–white mica geothermobarometry are employed to characterize the mineralogical and geochemical evolution of the deformed rocks, assess their diagenetic/metamorphic grade, and infer paleo-fluid conditions.

The integrated approach adopted in this research addresses several key objectives:

- To determine the thermobaric and diagenetic/metamorphic conditions associated with *mélange* formation in LMB.
- To determine the thermobaric and diagenetic/metamorphic conditions and reconstruct the structural evolution and pressure-temperature constraints within the LAC.
- To understand the basin geometry and source supply in the CiG wedge-top basin.

Despite the well-documented importance of *mélange*/shear zones and accretionary complexes in convergent settings worldwide, relatively few studies have combined structural, petrographic, and geochemical data to fully resolve the deformation mechanisms at work. In the southern Apennines, the Tempa Roccarossa area represents one of the most promising yet underexplored sites for addressing these issues. This research contributes to the growing body of work aimed at understanding the evolution of tectonically complex zones, with implications that extend beyond academic and applied geosciences.

The thesis is organized as follows:

1. Chapter I – Introduction, Geological Setting and Materials and Methods

This chapter provides a comprehensive literature-based overview of the southern Apennines, with specific emphasis on the Mt. Sirino and Cilento areas. It outlines the regional geological setting, stratigraphy, tectonic evolution, and previous research

relevant to the Tempa Roccarossa mélange, the LAC, and the CiG wedge-top basin. The chapter also details the scientific rationale, objectives, and significance of the study, followed by a summary of the analytical methodologies adopted throughout the research.

2. Chapter II – The Tempa Roccarossa Tectonic Mélange

This chapter explores the structural architecture, strain and burial history of the Tempa Roccarossa mélange. It provides detailed descriptions of mélange fabric, block-in-matrix relationships, shear zone structures, and strain partitioning. Microstructural and geochemical analyses, including mineral chemistry, XRPD, and SEM-EDS data, are used to reconstruct the deformation history, metamorphic conditions, and fluid-rock interaction within this high-strain zone.

3. Chapter III – The Ligurian Accretionary Complex

This chapter presents the original research results related to the LAC. It includes a general introduction to the structural and lithological characteristics of this unit, followed by a detailed description of field observations, strain analysis, petrographic data, and geochemical signatures. The deformation mechanisms, finite strain, metamorphic evolution, and implications for accretionary wedge development are discussed in detail.

4. Chapter IV – The Wedge-Top Basin of Cilento Group

This chapter focuses on the tectono-sedimentary evolution of the CiG wedge-top basin. It includes analysis of sedimentary structures, syntectonic sequences, and depositional patterns in basin-fill deposits. The chapter examines the timing and processes of basin development and their relationship with ongoing thrust propagation, drawing on sedimentological data.

5. Chapter V – Concluding Remarks

The final chapter integrates the findings from the previous chapters to develop a comprehensive tectono-stratigraphic model for the study area. It reconstructs the evolution of the LAC, CiG, and the LMB within the broader context of the southern Apennines. This chapter synthesizes the implications of mélange formation, strain

localization, and sedimentary basin development in relation to convergent margin dynamics and outlines potential directions for future research.

1.2 Problematics and Goals

The southern Apennines represent a key segment of the central-western peri-Mediterranean orogenic Alpine system and offer a unique opportunity to investigate the full progression of the Wilson Cycle, from oceanic opening and subduction to continental collision and post-orogenic processes. However, several critical aspects of the region's geodynamic evolution remain insufficiently constrained. In particular, the paleogeographic configuration and tectono-sedimentary evolution of the Cilento wedge-top basin, the thermobaric constraints, deformation conditions, and structural architecture of the major thrust/decollement/shear zones, and the deformation and diagenetic/metamorphic imprint of the Ligurian accretionary complex remain subjects of active debate.

1.2.1 Research questions

The tectonic *mélange* represents a structurally chaotic unit whose origin, internal dynamics, and metamorphic conditions are only partially understood. In particular, the relationship between *mélange* formation, subduction processes, and subsequent exhumation requires further investigation. Furthermore, the Ligurian accretionary complex, despite being a well-preserved remnant of the subducted Ligurian ocean, shows intricate deformation fabrics and metamorphic overprints that challenge straightforward tectonic interpretations.

Another major problem lies in reconstructing the paleogeography of the Cilento wedge-top basin during its synorogenic development, which is complicated by the complex interfingering of turbiditic units, variable sediment provenance, and diachronous thrust activity.

Specific research questions this Ph.D. addressed are as:

- How do the metamorphic conditions and deformation mechanisms within the tectonic *mélange* inform us about orogenic burial (crustal stacking) and exhumation in the Southern Apennines?
- What are the specific links between deformation mechanisms observed in the Ligurian Accretionary Complex and the overall tectonic and geodynamic processes associated with the closure of the Ligurian ocean?
- What is the detailed tectonic relationship between the Cilento wedge-top basin and the thrust belt propagation during the Oligocene–Miocene, and how does this affect the basin's evolution?

1.2.2 Goals

The primary goal of this study is to provide a comprehensive reconstruction of the tectonic evolution of the Southern Apennines through an integrated analysis of the Cilento wedge-top basin, the tectonic *mélange*, and the Ligurian accretionary complex. Specifically, the research aims to:

1. Constrain the diagenetic/metamorphic conditions of the tectonic *mélange* in Tempa Roccarossa, focusing on pressure–temperature estimates, and their implications for orogenic construction and related mechanics.
2. Analyse the structural and metamorphic evolution of the Ligurian Accretionary Complex to better understand deformation mechanisms at the frontal accretion and their linkage to crustal stacking and exhumation.
3. Reconstruct the paleogeographic and tectonic evolution of the Cilento wedge-top basin during the Oligocene–Miocene, with emphasis on basin evolution and dynamics in relation to thrust belt propagation.

Through a multidisciplinary approach combining field geology, structural analysis, geochemistry, and low-grade metamorphic petrology, this research seeks to refine the geodynamic framework of the Southern Apennines and to contribute to broader models of convergent margin evolution. By addressing these problems, the thesis will provide new constraints on the tectonic processes that governed the closure of the Ligurian ocean and the subsequent orogenic development of the Apennine chain.

Chapter 2 Regional Geological Framework

The southern Apennines, together with the northern Calabria, here and after called NCSA system (Figure 2-1 a), form a segment of the Alpine belt in the central-western Mediterranean Sea (Figure 2-1 b). The initial general model for the tectonostratigraphic structure of the orogenic wedge at shallow crustal levels was established in the 1960s and (Selli, 1962; Ogniben, 1969; D'Argenio et al., 1973; Amodio-Morelli, 1976) and is characterized by the superposition of various allochthonous thrust sheets that include both metamorphic and sedimentary sequences. In general, the chain is characterized, along a transect extending from northern Calabria to the southern Apennines (Figure 2-1 c, d), by a tectonic structure formed from Palaeozoic basement and Mesozoic sediments (overriding plate), which covers metamorphic units (originating from both oceanic and continental realms) that are in turn overlaid onto a non-metamorphic fold-and-thrust belt formed by continental sequences of the down going Adria plate. The significant advancements in our understanding of deep structural features have been primarily driven by hydrocarbon exploration, which involved several hundred boreholes reaching depths of up to 7 km (ViDEPI-Project, 2009–2015), alongside the CROP Project that included numerous seismic lines probing down to 15 km (<http://www.crop.cnr.it>). This encompasses the CROP04 profile traversing the southern Apennines (e.g., Cippitelli, 2007; Patacca & Scandone, 2007). Despite various and occasionally opposing interpretations of the tectonic structure, seismic and borehole data indicate the presence of a buried imbricated para-autochthonous belt consisting of Apulian Platform carbonates (APB) situated beneath the allochthonous Apennine wedge. More recent studies (Casero et al., 1988; Doglioni et al., 1996; Noguera & Rea, 2000; Shiner et al., 2004; Mazzoli et al., 2006; Cippitelli, 2007; Mazzoli et al., 2008; Scrocca, 2010; Carminati & Doglioni, 2012; Vitale et al., 2013b) highlight the general structure of the southern Apennines as defined by prevailing low-angle thrusts separating the major thrust-sheets, in turn, crosscut by high-angle thrusts enucleated in the APB. These different thrust faults resulted from alternating thin- and thick-skinned tectonics with a prevalence of the latter during the Pliocene that triggered deep breaching thrusting (e.g., Butler, 1987; Morley, 1988) or envelopment thrusting (Bally et al., 1966), with a significant surficial out-of-sequence thrusting that involved the wedge-top basin deposits (Vitale et al., 2017; Vitale et al., 2020a; S. Vitale et al., 2020b).

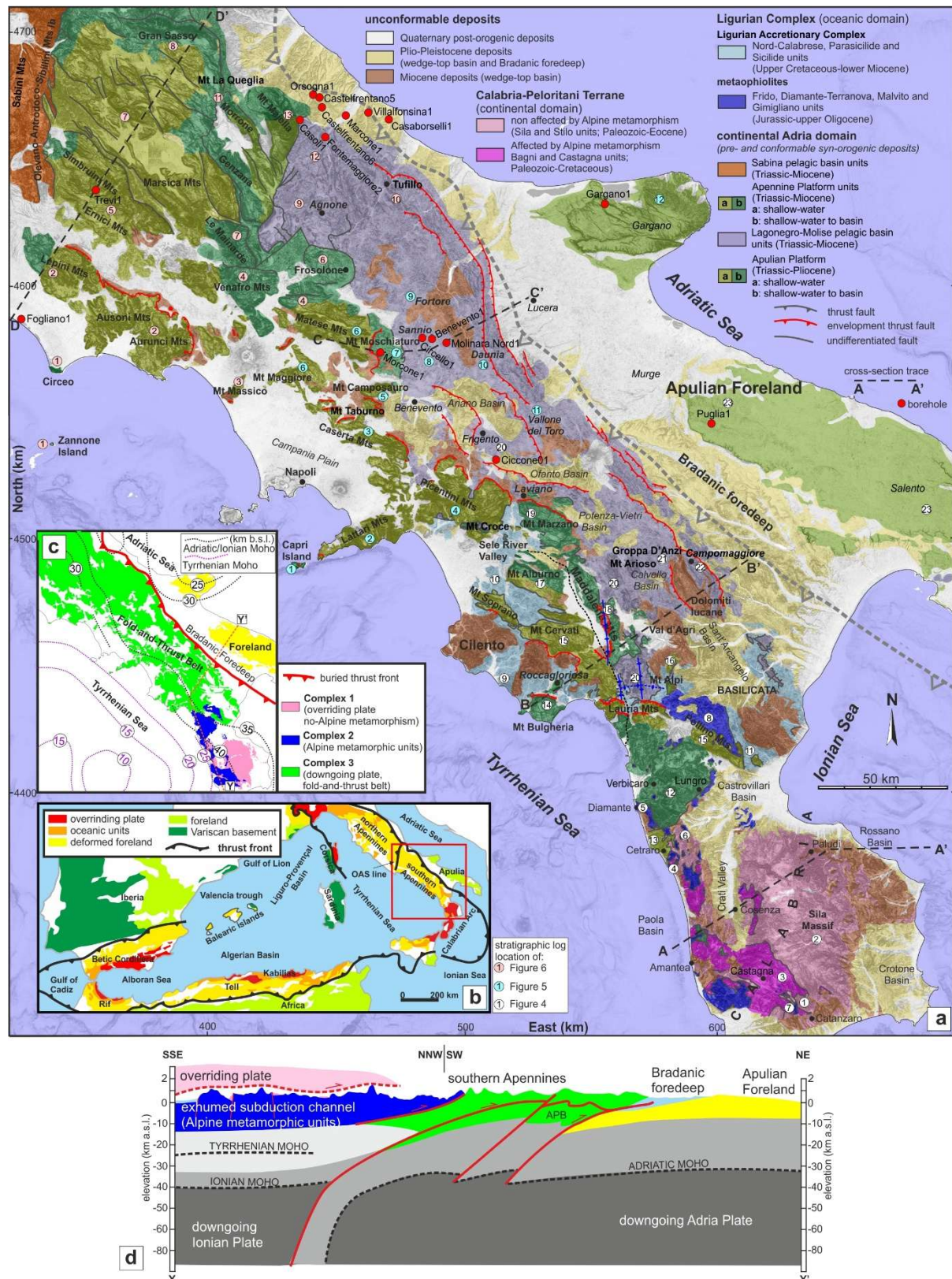


Figure 2-1: Geological map of the northern Calabria-southern Apennines (NCSA) System; (b) Tectonic sketch of western-central Mediterranean Alpine belts. OAS: Olevano-Antrondoco-Sibillini line. (c) Map of the orogenic complexes. Figure from Ciarcia and Vitale (2025)

The NCSA system (Figure 2-1 a) resulted from the subduction of the Neo-Tethys (Ligurian) oceanic lithosphere under the Europe/Calabria-Peloritani Terrane (CPT) in the Paleocene-Eocene time (Schenk, 1980; Rossetti et al., 2001; Rossetti et al., 2004; Lustrino et al., 2009; Vignaroli et al., 2012; Vitale et al., 2019a) and subsequently from the deformation of the continental Adria slab during the Oligocene-Middle Pleistocene. It is considered a retreating thrust belt where the subduction rate exceeds the convergence rate (Royden, 1993), triggering the formation of back-arc basins of the Ligurian-Provençal and Tyrrhenian seas and the stretching of the overriding plate, including the Sardinia-Corsica and CPT blocks (Faccenna et al., 1996; Gueguen et al., 1998; Martin, 2006; Carminati & Doglioni, 2012).

During the Oligocene-Middle Pleistocene, the thrust front-foredeep-forebulge system mainly migrated toward E/SE (e.g., Vitale et al., 2013b; Sabbatino et al., 2020; Sabbatino et al., 2021), and the foredeep sedimentation was generally preceded by emersion and erosion of the shallow-water domains caused by the flexure of the downgoing Adria plate (forebulge stage; (Sabbatino et al., 2020). The youngest foredeep basin sediments constrain the lower timing limit of the thrust front migration. In contrast, the oldest overlying unconformable wedge-top basin deposits, locally sealing contacts between different thrust sheets, fix the upper temporal limit of the thrusting pulse (e.g., Cosentino et al., 2003; Bonardi et al., 2009; Vitale & Ciarcia, 2013b; Vitale et al., 2018).

From a paleogeographic point of view, the NCSA system (Figure 2-1 a) includes tectonic units derived from (i) the overriding plate, including Paleozoic crystalline basements (CPT; (Bonardi et al., 2001), (ii) oceanic to ocean-continent-transition (OCT) basin, and (iii) the continental Adria Plate (e.g., Mazzoli et al. (2008).

The orogenic structure includes the superposition of three orogenic complexes (Figure 2-1 c). Complex 1, located on the top of the chain, comprises remnants of the overriding plate (CPT), covering an Alpine metamorphic belt (Complex 2) corresponding to the exhumed subduction channel (SC), formed by CPT units, metamorphic ophiolitic units, and some metamorphic Meso-Cenozoic thrust sheets of the continental Adria plate (downgoing plate). Finally, this orogenic prism tectonically covers a thrust-and-fold belt (Complex 3) defined by a series of stacked thrust sheets, including successions of different paleoenvironments, from deep basin to shallow-water platform (e.g., Merlini & Mostardini, 1986; Casero et al., 1988; Doglioni et al., 1996; Ciarapica & Passeri, 1998; Noguera & Rea, 2000; Faccenna et al., 2001a; Faccenna et al., 2001b; Catalano et al., 2004; Shiner et al., 2004; Guerrera et al., 2005; Patacca & Scandone, 2007; Mazzoli et al., 2008; Bonardi et al., 2009; Cosentino et al., 2010; Scrocca, 2010; Vezzani et al., 2010; Roure et al., 2012; Vitale & Ciarcia, 2013b; Vitale et al., 2018;

Vitale & Ciarcia, 2018; Vitale et al., 2020a; Vitale et al., 2020b; Vitale & Ciarcia, 2022). The youngest thrust front is buried below the Bradanic foredeep basin deposits (e.g., Tropeano et al. (2002). The exposed carbonates of the Puglia region and the Adriatic Sea represent the current foreland (Figure 2-1 c). Figure 2-1 (c) also shows the isobaths of the Adriatic/Ionian and Tyin. Figure 2-1 (d) shows the current lithospheric architecture along a transect from northern Calabria to the Puglia region (trace in Figure 2-1 c), with the subduction of the Ionian oceanic lithosphere.

Complex 1 (Figure 2-1 c) is formed by a Variscan Paleozoic continental crust and relative sedimentary Mesozoic cover (Sila and Stilo units). The Complex 2 (Figure 2-1 c) includes all successions affected by Alpine metamorphism: (i) the CPT Units of Castagna and Bagni (Amodio-Morelli, 1976; Bonardi et al., 2001); (ii) metaophiolites of Diamante-Terranova, Malvito, Gimigliano and Frido Units (Figure 2-1 a, Figure 2-2; (Amodio-Morelli, 1976; Bonardi et al., 2001; Rossetti et al., 2001; Rossetti et al., 2004; Liberi et al., 2006; Liberi & Piluso, 2009; Vitale et al., 2019a); (iii) continental Adria Units of Lungro-Verbicaro and Cetraro (Iannace et al., 2007; S. Vitale et al., 2020b). The A-A' cross-section (Figure 2-2), showing a transect across northern Calabria, highlights a western sector formed by the superposition of different HP-LT metamorphic units, including CPT Units above the Ligurian metaophiolites, in turn overlying the Apennine Platform Units. The Sila Unit tectonically covers all these thrust sheets. In the eastern sector, the latter unit is widespread and crosscut by NE-verging thrust faults that deform the wedge-top basin deposits of Paludi Fm (PLU), which are both unconformably topped by the Rossano Basin deposits (ROS).

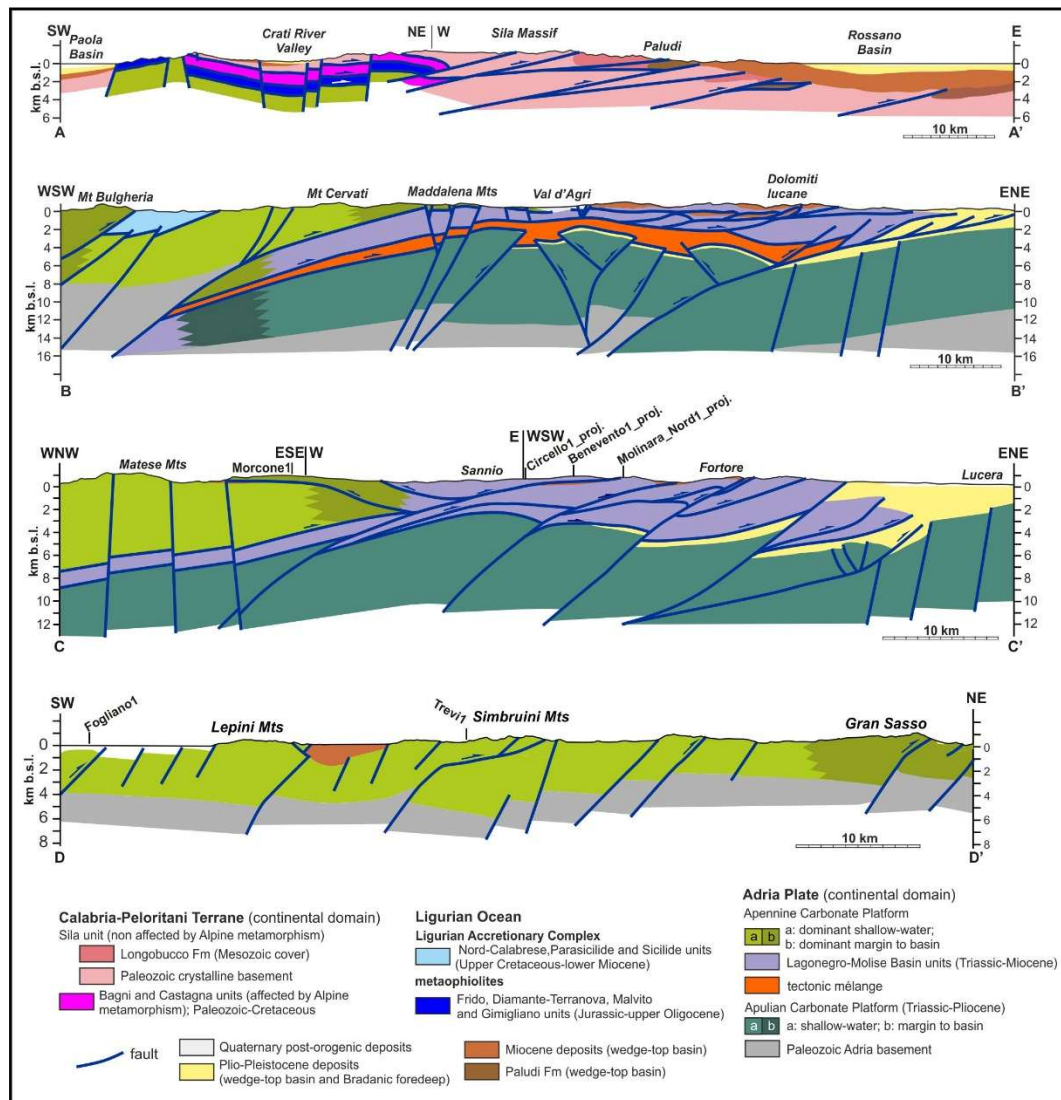


Figure 2-2: Geological cross sections. Figure from Ciarcia and Vitale (2025).

Complex 3 (Figure 2-1 c) is made of sedimentary units of (i) Nord-Calabrese, Parasicilide and Sicilide (Ligurian Accretionary Complex, LAC; Figure 2-1 a; Selli, 1962; Ogniben, 1969; Bonardi et al., 1988a; Ciarcia et al., 2009a; Vitale et al., 2010; Ciarcia et al., 2012; Vitale et al., 2019a); (ii) continental Adria Units of Meso-Cenozoic Apennine and Apulian platforms and the Lagonegro-Molise Basin (D'Argenio et al., 1973; Patacca & Scandone, 1989; Bigi et al., 1992; Bonardi et al., 1992; Cosentino et al., 2003; Patacca & Scandone, 2007; Bonardi et al., 2009) detached from their pre-Triassic basement in the central and external sectors of the chain (C-C' cross-section of Figure 2-2; e.g., Casero et al., 1988; Noguera & Rea, 2000; Shiner et al., 2004; Cippitelli, 2007).

The shallow crustal structure of the southern part of the southern Apennines (Figure 2-2, B-B' and C-C' cross-sections) is dominated by low-angle thrusts between (i) oceanic units (on the top; Ligurian Complex; Figure 2-1 a), (ii) carbonate platform/slope successions of the

Apennine Platform and (iii) thrust sheets of the pelagic Lagonegro-Molise Basin successions (Merlini & Mostardini, 1986). On the contrary, at depth, the crustal structure is characterized by high-angle faults affecting the buried Apulian Platform (APB) and the allochthonous wedge (Figure 2-2, B-B' and C-C' cross-sections). The northern sector of the southern Apennines (D-D' cross-section; Figure 2-2) is less shortened compared to the southern sector and consists of a tectonic pile characterized by high-angle thrusts rooted in the Paleozoic basement still attached to the Meso-Cenozoic cover (e.g., Tozer et al. (2002).

According to several paleogeographic reconstructions (e.g., Rossetti et al., 2004; Patacca & Scandone, 2007; Vezzani et al., 2010; Patacca & Scandone, 2011; Schettino & Turco, 2011), the pre-orogenic palaeogeography of the western-central Mediterranean area (latest Cretaceous-Paleocene; Figure 2-3 a, (modified after Vitale and Ciarcia (2013b) shows the Ligurian Ocean separating the continental sector of the European Plate, which includes the Calabria-Peloritani Terrane and the Sardinia-Corsica block, from the Africa-Adria Plate. In this period, the Alps were forming with the subduction of the oceanic part of the European plate (Piemont Ocean) beneath the continental part of the Adria plate (Schmid et al., 1996; Handy et al., 2010). In contrast, the southern part of the continental Adria plate and the northern sector of the Africa plate were undergoing a crustal stretching interpreted as the result of a rifting stage (Vitale & Ciarcia, 2022) and references therein) or to a far-field stress state associated with the orogenic processes that involved the northern edge of the Adria (Agosta et al., 2021). In this reconstruction (Figure 2-3 a), the Ligurian and Ionian oceans were separated by a strip joining the continental parts of the Adria and Africa plates. In the Cretaceous time, this continental bridge between the two plates allowed Dinosaurs to migrate between the two sectors (Zarcone et al., 2010). However, in the latest Cretaceous, this emerged/shallow-water area partially drowned, with the complete separation of the Apennine (Adria domain) and Panormide (Africa domain) platforms (Vitale et al., 2013a; Vitale & Ciarcia, 2022), despite that the two parts (Apennine and Panormide platforms) still belonged to a single plate (Channell, 1996).

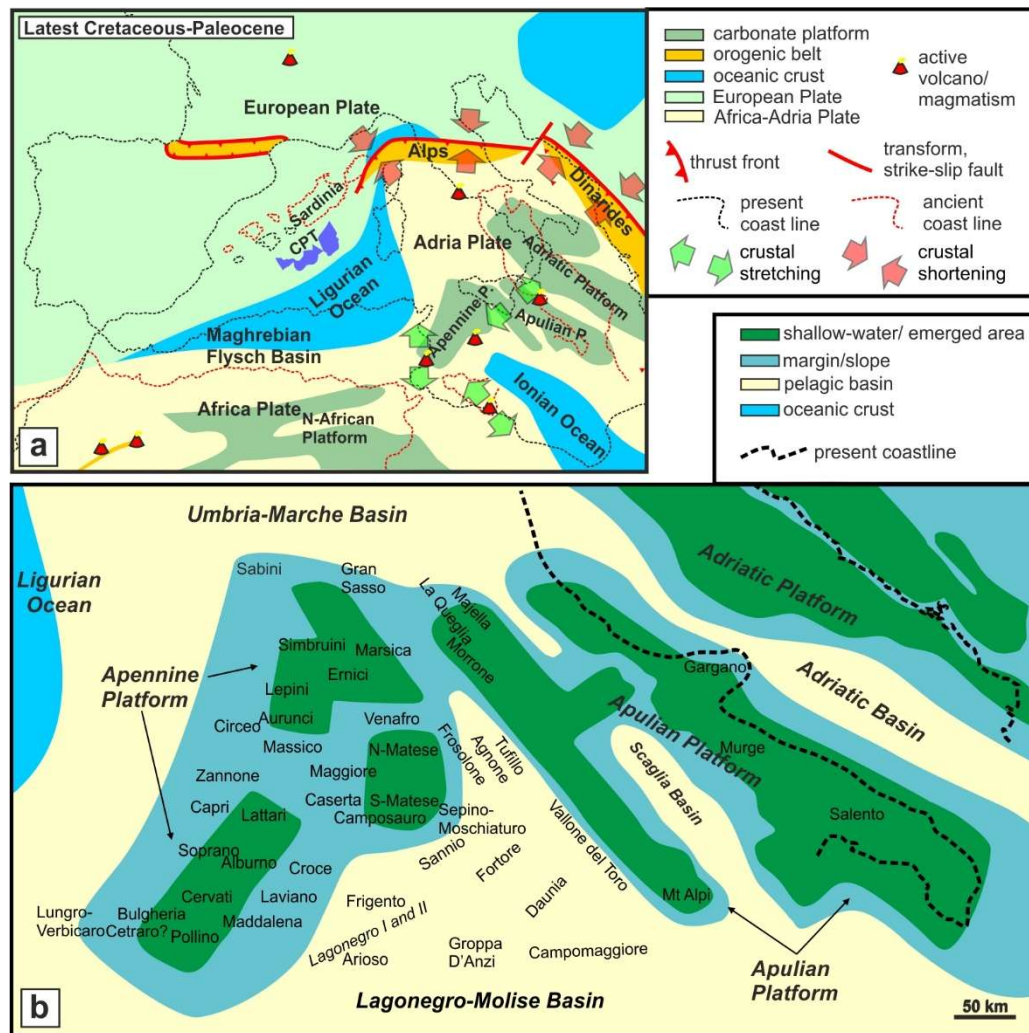


Figure 2-3: Latest Cretaceous-Paleocene paleogeographic maps: (a) western-central Mediterranean area; (b) zoom of the Apennine Platform/Lagonegro-Molise Basin/Apulian Platform area. Figure from Ciarcia and Vitale (2025).

The zoomed map of the Apennine Platform/Lagonegro-Molise Basin/Apulian Platform area (Figure 2-3; Vitale and Ciarcia (2022)) shows the articulated pre-orogenic palaeogeography of the southern continental sector of the Adria plate. To reconstruct this palaeogeography, we started from the available maps (Patacca & Scandone, 2007; Vezzani et al., 2010; Vitale et al., 2013b; Vitale & Ciarcia, 2022). We considered the paleoenvironments reconstructed for the major successions, including the uppermost Cretaceous-Paleocene shallow-water, margin to slope-basin deposits of the Apennine and Apulian platforms, and the pelagic basin successions of the Lagonegro-Molise Basin domain. In this reconstruction, the Apennine-Apulian platform system was isolated, surrounded by pelagic basins, including the Lagonegro-Molise Basin to the south, the Ligurian Ocean to the west, the Umbria-Marche Basin to the north and finally, the Adriatic Basin to the east (Figure 2-3 b). It is worth noting that the Apulian Platform hosts

an NW-SE elongated basin where pelagic sediments were deposited, such as resulting from the borehole reconstructions (Patacca & Scandone, 2013).

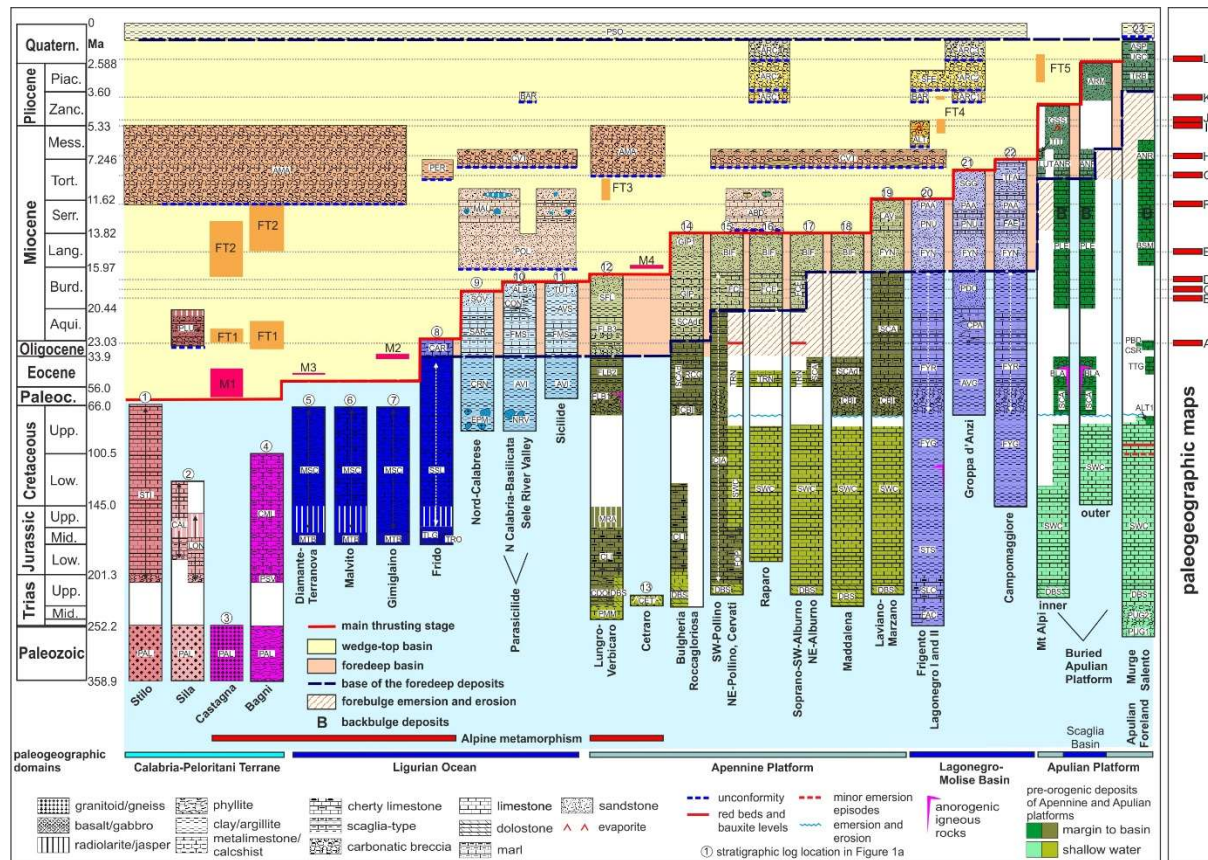


Figure 2-4: Schematic stratigraphic logs of the northern Calabria-southern Apennines (southern sector). Calabria-Peloritani Terrane: CAL: Caloveto and Longobucco groups; CML: cherty meta-limestone. PAL: Paleozoic basement; PSV: Pseudo-Verrucano Fm. Ligurian Ocean: ALB: Albanella Fm; AVI: Argille Varicolori Inferiori Fm; AVS: Argille Varicolori Superiori Fm; CAR: Mt Caramola Fm; CON: Contursi Fm; CRN: Crete Nere Fm; EPM: Episcopia mélangé; FMS: Mt Sant'Arcangelo Fm; MSC: undifferentiated meta-sedimentary cover; MTB: undifferentiated metabasalts; NRV: Noce River Valley mélangé; POS: Postiglione Fm; SAR: Saraceno Fm; SOV: Sovereto Fm; SSL: S. Severino Lucano Fm; TLG: Tempa La Guardia Fm; TRO: Tempa Rotalupo Fm; TUT: Tufiti di Tusa-Arenarie di Corleto Fms. Apennine Platform: BIF: Bifurto Fm; CBI: Calcarei Cristallini Fm; CDO: cherty dolomites; CIA: Slope limestones of Pollino-Ciagola unit; CLI: cherty limestone; DBS: Dolomia Superiore Fm; FCE: Cerchiara Fm; FLB1: Alberosa Fm (Breccie Poligeniche mb); FLB2: Alberosa Fm (Calcarei a Nummiliti mb); FLB3: Alberosa Fm (Serra Costantino mb); FOP: Dolomie e Calcarei di Monte Foraporta Fm; FYN: Numidian Flysch Fm; GIP: Argilliti e Calcarei di S. Giovanni a Piro Fm; LAV: Laviano marls, calcarenites and sandstones. MRA: metaradiolarites; PMM: phyllites, metalimestones and metarenites; RCG: Calcarei e Argilliti di Roccalgorosa Fm; SCA: Scaglia-type deposits; SCAd: Scaglia Detritica-type deposits; FL: Scisti del F. Lao Fm; CET: Cetraro Unit; SWC: Jurassic-Cretaceous shallow-water carbonates; TRN: Trentinara Fm. Lagonegro-Molise Basin: AVG: Argille Variegate Fm; CPA: Corleto Perticara Fm; FAC: Monte Facito Fm; FAE: Faeto Fm; FYG: Flysch Galestrino Fm; FYN: Numidian Flysch Fm; FYR: Flysch Rosso Fm; PAA: Serra Palazzo Fm; PDO: Paola Doce Fm; PNU: Post-Numidian succession; SGG: San Giorgio Fm. SLC: Calcarei con Selce Fm; STS: Scisti Silicei Fm; TFA: Torrente Flaga Fm. Apulian Platform: ALT1: Calcare del Ciolo mb (Calcare di Altamura Fm); ANR: Calcareniti di Andrano Fm; ARM: Clay and marl of buried outer Apulian Platform; BLA: Breccie di Lavello Fm; BSM: Calcarenite coralligena di Serra del Mito Fm; CSR: Calcare di Castro Fm; DBS: Dolomia Superiore Fm; GES: Gessoso-Solfifera Fm; LAP: Lama dei Peligni Flysch; LUT: Calcareniti di Cava Iannazzo; PBD: Calcareniti di Porto Badisco Fm; PLE: Pietra Leccese Fm; PUG1: sandstones; PUG2: breccias; SCA: Scaglia-type deposits; SWC: Jurassic-Cretaceous shallow-water carbonates (Calcarei di Bari and Calcarei di Altamura Fms); TIT: Conglomerati e arenarie di Cima S. Croce; TTG: Calcare di Torre Tiggiano; TRB: Trubi Fm; UGC: Uggiano La Chiesa Fm; ASP: Argille Subappennine and Montemarano

sands Fms. Wedge-top deposits: ABD: Albidona Fm (CIG); ALTG: Altavilla Group; AMA: Amantea; ARC: S. Arcangelo Basin deposits; BAR: Baronía Fm; CVTG: Castelvete Group; MAU: S. Mauro Fm (CIG); PLU: Paludi Fm; PER: Perosa Fm; POL: Pollica Fm (CIG); ROS: Rossano basin deposits; SFE: Sferracavallo Fm. Post-orogenic successions: PSO: Upper Pleistocene to Present deposits. Metamorphic ages: M1: Castagna Unit (Schenk, 1980); M2: Gimigliano Unit (Rossetti et al., 2001; Rossetti et al., 2004); M3: Diamante-Terranova Unit (Shimabukuro et al., 2012); M4: Cetraro Unit (Iannace et al., 2007). Thermochronologic ages: a) Castagna and Bagni Units: FT1 zircon fission tracks (Thomson, 1998); FT2 apatite fission tracks (Thomson, 1998); b) Lungro-Verbicaro Unit: FT3 apatite fission tracks (Iannace et al., 2007); c) M. Alpi: FT4, (Mazzoli et al., 2006). Figure from Ciarcia and Vitale (2025).

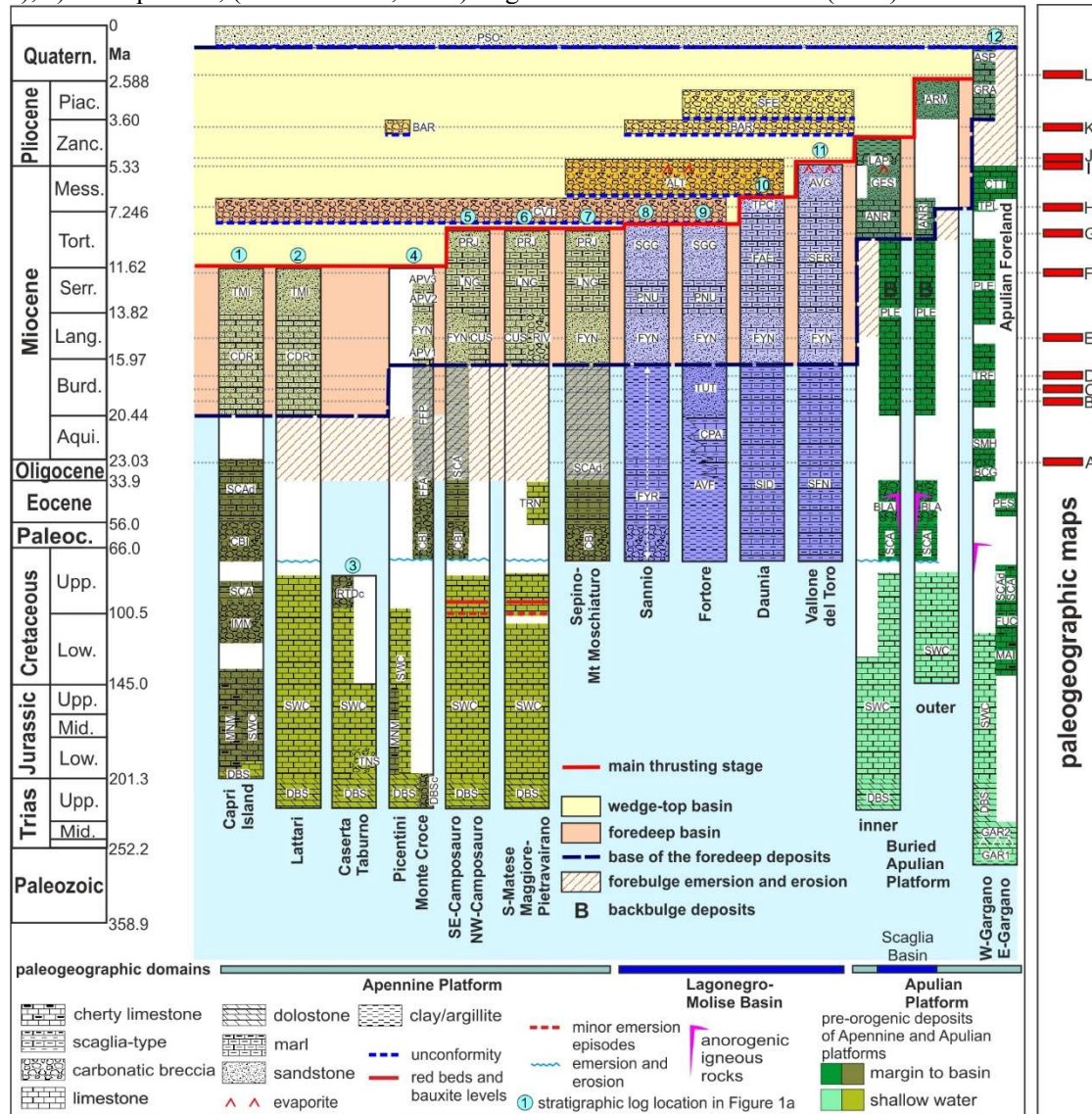


Figure 2-5: Schematic stratigraphic logs of the northern Calabria-southern Apennines (central sector). Apennine Platform: APV1: Serra della Manca Mb (Flysch della Vallimala Fm); APV2: Serralonga Mb (Flysch della Vallimala Fm); APV3: La Foresta Mb (Flysch della Vallimala Fm); BDSc: reworked Dolomia Superiore Fm; CBI: Calcari Cristallini Fm; CDR: Calcareniti di Recommone Fm; CUS: Cusano Fm; DBS: Dolomia Superiore Fm; FFA: Frasci Fm; FFP: Fontana Porcellara; FPJ: Pietraraja Fm; FYN: Numidian Flysch Fm; IMM: Calciruditi e Calcareniti del Limmo Fm; LNG: Longano Fm; MNM: Calcari e Dolomie con selce dei Monti Mai Fm; REC: Calcareniti di Recommone Fm; RIV: Mt Crivari Fm. RTDc: reworked Calcari a Radioliditi Fm; SCA: Scaglia-type deposits; SCAd: Scaglia Detritica-type deposits; SWC: Jurassic-Cretaceous shallow-water carbonates; TMI: Arenarie di Termini Fm; TNS: Conglomerati calcarei di Tuoro Maleconsiglio; TRN: Trentinara Fm. Lagonegro-Molise Basin: AVF: Argille varicolori del Fortore Fm; AVG: Argille Varicolori with gypsum. CPA: Corleto Perticara Fm; FAE: Faeto Fm; FYN: Numidian Flysch Fm; FYR: Flysch Rosso Fm; PNU: Post-Numidian succession; SER: Serroni Fm; SFN: Serra Funaro Fm;

SGG: San Giorgio Fm; SID: Mt Sidone Fm; TPC: Toppo Capuana Fm; TUT: Tufiti di Tusa Fm. Apulian Platform: ANR: Calcareni di Andrano Fm; ARM: Clay and marl of buried outer Apulian carbonate Platform; ASP: Argille subappennine and Montemarano sands Fms. BCG: Breccia di Cagnano Fm; CTT: Calcari di Trombetta, Breccie e conglomerati di Stazione S. Giacomo and Calcari di Fiumicello Fms; DBS: Dolomia Superiore Fm; FUC: Marne a fucoidi Fm; GAR1: shales, dolostones and limestones; GAR2: sandstones, dolostones and limestones; GRA: Calcareni di Gravina Fm; GES: Gessoso-Solfifera Fm; LAP: Lama dei Peligni Flysch; LAV: Breccie di Lavello Fm; MAI: Maiolica Fm; PES: Peschici Fm; PLE: Pietra Leccese Fm; SCA: Scaglia-type deposits; SCA: Scaglia-type Fm; CAD: Scaglia Detritica-type deposits; SMH: Calcari di Grotta San Michele Fm; SWC: Jurassic-Cretaceous shallow-water carbonates (Calcari di Bari and Calcari di Altamura Fms); TPL: Tripoli Fm; TRF: Torre Rossa Fm. Wedge-top deposits: CVTG: Castelvetere Group; ALTG: Altavilla Group; BAR: Baronica Fm; SFE: Sferracavallo Fm. Post-orogenic successions: PSO: Upper Pleistocene to Present deposits. Figure from Ciarcia and Vitale (2025).

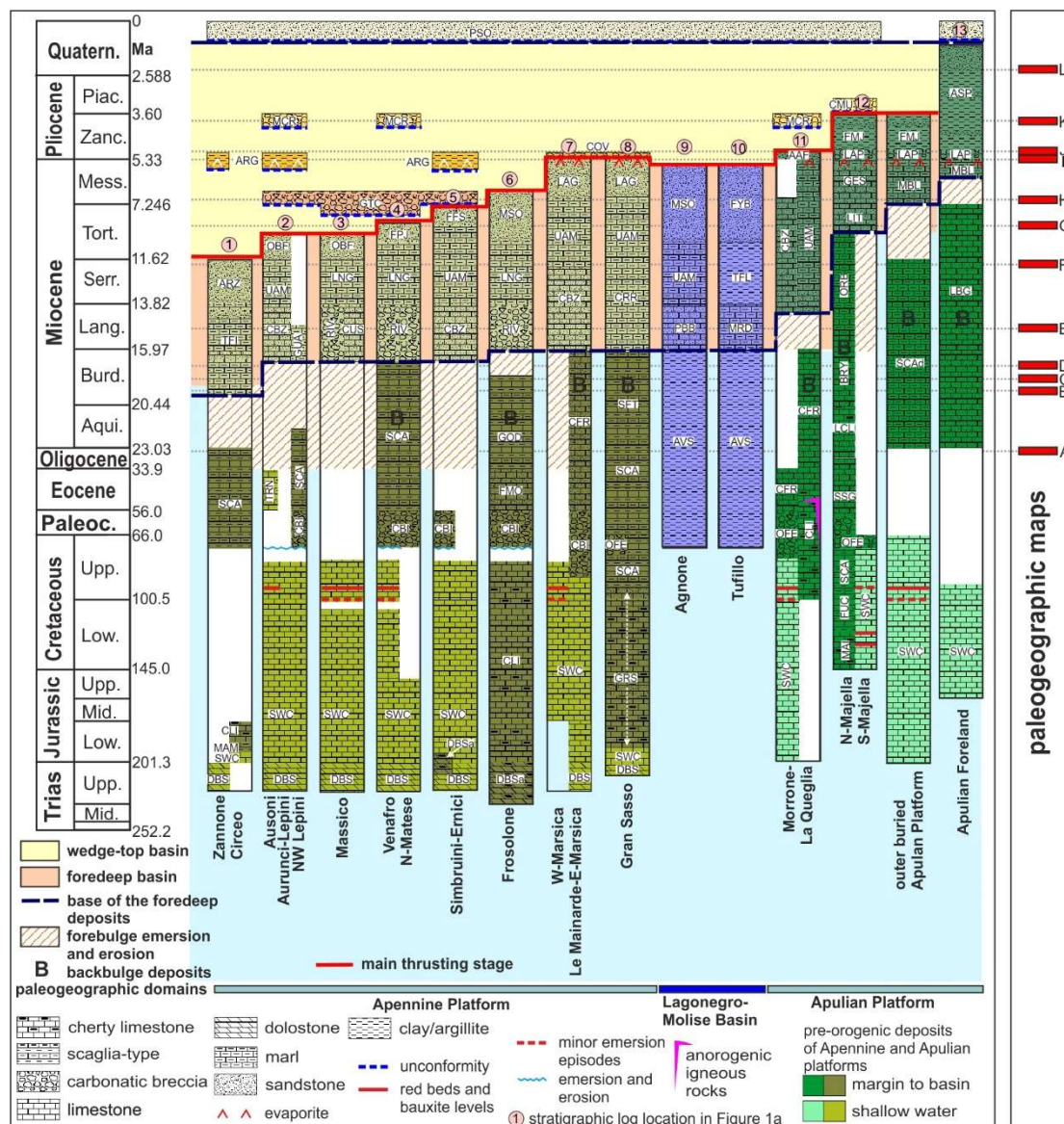


Figure 2-6: Schematic stratigraphic logs of the northern Calabria-southern Apennines (central sector). Apennine Platform: AAF: Anversa degli Abruzzi Fm. ARZ: Arenarie di Zannone; CBI: Calcari Cristallini Fm; CBZ: Calcari a Briozoi a Litotamni Fm; CFR: Calcareni a Macroforaminiferi Fm; CLI: cherty limestones; CRR: Marne con Cerrognia Fm; CUS: Cusano Fm; DBS: Dolomia Superiore Fm; DBSa: cherty dolomites; FFS: Frosinone Fm; FMO: Monaci Fm; FPJ: Pietraraja Fm; FYB: Flysch di S. Biase Fm; GOD: Macchiagodena Fm; GRS: Corniola, Verde Ammonitico e Calcari e Marne a Posidonia, Calcari bioclastici inferiori, Calcari diasprigni, Maiolica detritica, Calcari bioclastici superiori and Calcareni e Calciruditi a

Fucoidi Fms; GUA: Guadagnolo Mb; LAG: Laga Fm; LNG: Longano Fm; MAM: marls with ammonites; MSO: Flysch del Molise Fm; OBF: Occhio del Bue succession; OFE: Orfento Fm; RIV: Mt Crivari Fm; SCA: Scaglia-type deposits; SFT: Unità Spongolitica Fm; TFI: Torre Fico succession; SWC: Jurassic-Cretaceous shallow-water carbonates; TFL: Tuffillo Fm; UAM: Marne a Orbuline, and Argille a Orbulina Fms. Lagonegro-Molise Basin: AVS: Argille Varicolori Superiori Fm; MRD: Marne Rosse e Verdi di Colle Dolce Fm; MSO: Flysch del Molise Fm; PBB: Calcareni e Calciruditi di Pietrabbondante Fm; UAM: Argille a Orbulina Fm. Apulia Platform: ASP: Argille subappennine and Montemarano sands Fms. BRY: Bryozoan Limestone Fm; FMJ: Flysch della Majella Fm; FUC: Marne a Fucoidi Fm; GES: Gessoso-Solfifera Fm; LAP: Lama dei Peligni Flysch; LBG: Limestone with Bryozoa and glauconite; LCL: Lepidocyclina and Cerratina cherty limestones Fms; LIT: Lhithotamnium Limestone Fm; MAI: Maiolica Fm; MBL: marly limestone with reworked Bryozoan Unit; OFE: Orfento Fm; ORB: Orbulina Limestone Fm; SCA: Scaglia-type deposits; SSG: Santo Spirito Group. SWC: Jurassic-Cretaceous shallow-water carbonates (Calcari di Bari and Calcari di Altamura Fms). Wedge-top deposits: ALTG: Altavilla Group; CMU: Castilenti and Mutignano Fms; GTC: Gavignana-Gorga, and Torrice Fms; MCR: Minturno conglomerates, Campobasso conglomerates and Rigopiano Fm. Post-orogenic successions: PSO: Upper Pleistocene to Present deposits. Figure from Ciarcia and Vitale (2025).

2.1 Stratigraphy of pre-orogenic successions

Figure 2-4, Figure 2-5, Figure 2-6 show the stratigraphic logs of the major tectonic units and representative stratigraphic successions, grouped in agreement with their paleogeographic location (Calabria-Peloritani Terrane, Ligurian Ocean, Apennine Platform, Lagonegro-Molise Basin, and Apulian Platform) and separated into (i) pre-orogenic, (ii) conformable synorogenic (foredeep and backbulge basin deposits), (iii) unconformable synorogenic (wedge-top basin deposits) and (iv) post-orogenic rocks. From left to right, the order of the stratigraphic logs follows the ages of (i) the foredeep deposits or (ii) the youngest ages of the pre-orogenic rocks. The synorogenic succession will be described in paragraph 5 within the orogenic evolution of the NCSA.

In the following descriptions, the pre-orogenic successions are all the deposits formed before the forebulge and foredeep stages (Figure 2-4, Figure 2-5 and Figure 2-6). Calabria-Peloritani Terrane and metaophiolites of Diamante-Terranova, Malvito, and Gimigliano successions are all characterized by pre-orogenic rocks with ages older than the Paleocene-Eocene, namely the time proposed for the onset of the oceanic subduction (Schenk, 1980; Rossetti et al., 2001; Lustrino et al., 2009; Vignaroli et al., 2012; Vitale et al., 2019a). On the contrary, Frido, LAC, and the continental Adria successions are all formed by pre- and syn-orogenic successions, whose ages vary according to their paleogeographic positions. The first orogenic event recorded in the continental part of the Adria Plate was the Oligocene forebulge stage that produced uplift and erosion of the Apennine Platform and sedimentation in the surrounding basins. Subsequently, the forebulge ridge migrated toward E/NE (Sabbatino et al., 2020). In our reconstruction, the pre-orogenic successions are all deposits older than (i) Eocene for the Gimigliano, Malvito and Diamante-Terranova Units, (ii) Oligocene for the Frido, Nord-

Calabrese, Sicilide and Parasicilide Units, and the western side of the Apennine Platform (Lungro-Verbicaro Unit), (ii) Aquitanian-Langhian for central and eastern sectors of the Apennine Platform and the Lagonegro-Molise Basin, (iii) Serravallian-Tortonian for the buried Apulian Platform, and finally (iv) Messinian-Zanclean for the exposed Apulian Platform.

The continental sector of the Adria plate experienced different pre-orogenic tectonic episodes of rifting. This area recorded a Late Triassic-Early Jurassic extensional event that profoundly changed the shallow-water domain that characterized the Late Triassic palaeogeography (Bosellini & Hsü, 1973). This rifting stage produced an articulated palaeogeography characterized by two main carbonate platforms (Apennine and Apulian) defined by shallow-water areas passing to surrounding pelagic basins through margin and slope environments. The Apennine Platform passed southward to the Panormide Platform (Sicily), which was connected to the Constantine Platform (Tunisia) (e.g., Randazzo et al. (2021). The Lagonegro-Molise Basin was a pelagic realm between the Apennine and Apulian platforms (Merlini & Mostardini, 1986). This part of Adria suffered a new extensional stage in the Late Cretaceous-Eocene (Vitale & Ciarcia, 2022), that further dismembered the original shallow-water domains, creating new slope-to-basin areas, including the Scaglia Basin within the Apulian Platform (Figure 2-3 b; Merlini & Mostardini, 1986; Patacca & Scandone, 2013). This event also caused the separation of the Apennine Platform from the Panormide Platform (Vitale et al., 2013a). We refer to the Latest Cretaceous-Paleocene palaeogeography shown in Figure 2-3, as the starting point for the Eocene-Recent stratigraphic and tectonic reconstructions.

In the following paragraphs, I focus on the stratigraphy of the Ligurian and Adria units, the object of my research.

2.1.1 Ligurian Ocean

In northern Calabria, oceanic successions are exposed beneath the Calabria-Peloritani Terrane units, whereas in the southern Apennines they are located at the top of the orogenic wedge. The Ligurian successions (Figure 2-4) include the metaophiolites of Diamante-Terranova, Malvito, Gimigliano and Frido Units (Liberi et al., 2006; Liberi & Piluso, 2009; Vitale et al., 2019a) and the sedimentary successions of Nord-Calabrese, Parasicilide and Sicilide Units (LAC; Selli, 1962; Bonardi et al., 1988a; Guerrera et al., 2005; Bonardi et al., 2009; Ciarcia et al., 2009a; Ciarcia et al., 2012; Vitale et al., 2013a; Critelli, 2018; Vitale et al., 2019a, and reference therein). The Diamante-Terranova, Malvito, and Gimigliano Units are all characterized by Alpine HP/LT metamorphism. They are exposed in northern Calabria (Figure 2-1) and include

Middle Jurassic metabasalts (MTB) at the base, covered by Dogger metaradiolarites and Cretaceous cherty metalimestones and metapelites (MSC). Unlike the other metaophiolites, the Alpine HP/LT, OCT Frido Unit includes continental crustal masses (granitoids and gneisses; TRO) alternating with oceanic crust (TLG) at the base. The pre-orogenic succession evolves to Dogger metaradiolarites, Cretaceous-Oligocene phyllites, calcschists and quartzites (SSL). The LAC successions are detached from their basement and are probably made of OCT crust (Vitale et al., 2013a). They mainly crop out in the Cilento area, Sele River Valley, and Basilicata region (Figure 2-1 a). The pre-orogenic Nord-Calabrese succession is formed by Upper Cretaceous-Eocene black shales (CRN) embedding the Episcopia tectonic mélange, on the other hand, the Parasicilide and Sicilide deposits consist of Upper Cretaceous-Eocene deep-water varicoloured argillites (AVI) at places including basaltic clastics.

2.1.2 Apennine Platform

The pre-orogenic successions referred to this paleogeographic domain include deposits sedimented along the western margin, platform s.s. and eastern margin (e.g., D'Argenio et al., 1973; Pescatore & Sgroso, 1973; Parotto & Pratlun, 1975; Cosentino et al., 2003; Patacca & Scandone, 2007; Bonardi et al., 2009; Vitale & Ciarcia, 2013b, 2018; Cardello et al., 2021). From south to north, the western margin is represented by the Alpine metamorphic units of Lungro-Verbicaro and Cetraro (Figure 2-4) and the sedimentary successions of SW-Pollino Mts, Mt Bulgheria, Roccagloriosa, Capri Island, Zannone Island, Mt Circeo and some sectors of Lepini and Simbruini Mts (Scandone et al., 1964; Pantosti & Velonà, 1986; ISPRA, 1997, 2005a; Iannace et al., 2007; Patacca & Scandone, 2007; Bonardi et al., 2009; Vitale & Ciarcia, 2013b, 2018, 2022). The platform s.s. includes the successions of NE-Pollino, Cervati, Raparo, SW-Alburno, Caserta, Taburno, Picentini, NW-Camposauo, Matese, Maggiore, Pietravairano, Volsci Range (Aurunci, Ausoni, and Lepini), Massico, Simbruini, Ernici and W-Marsica Mts. The eastern margin encompasses the successions of Maddalena, NE-Alburni, Marzano, Laviano, Croce, SE-Camposauo, Sepino-Moschiatturo, Venafrò, N-Matese, Frosolone, Gran Sasso, and E-Marsica Mts (Corrado et al., 1998; Bernoulli, 2001; Cosentino et al., 2003; Patacca & Scandone, 2007; Bonardi et al., 2009; Cosentino et al., 2010; Vitale & Ciarcia, 2013b, 2018, 2022).

2.1.3 Apennine Platform western margin

The Lungro–Verbicaro succession (Iannace et al., 2007) consists of Anisian-Carnian phyllites, metalimestones, metarenites and dolomites (PMM), Norian dolomites (DBS) and cherty dolomites (CDO), Jurassic cherty limestones (CLI) and Dogger metaradiolarites (MRA), Maastrichtian–Paleocene calcareous breccias and calcarenites (FLB1) with anorogenic basaltic intrusions (Limburgites, (Quitow, 1935), and Eocene–Aquitania Scaglia-type deposits (FLB2).

Carnian phyllites, metalimestones, dolomites, and evaporites (CET; Iannace et al., 2007) characterize the Cetraro Unit (Dietrich, 1976). The SW-Pollino Mts (Figure 2-4) consist of Norian shallow-water dolomites (DBS) that pass upward into a Jurassic-Burdigalian margin succession formed by reworked limestones (CIA). The Mt Bulgheria succession (Figure 2-4; (ISPRA, 2016a, 2016b) is composed of shallow-water Norian dolomites (DBS) upward evolving to Jurassic-Lower Cretaceous cherty limestones (CLI), passing to Maastrichtian calcarenites and conglomerates with fragments of rudists (CBI) and Paleocene-Eocene Scaglia detritica deposits (SCAd). The Roccagloriosa deposits (ISPRA, 2016a, 2016b) are similar to the aforementioned Maastrichtian-Eocene succession (RCG). The Capri Island succession (Figure 2-5, ISPRA, 2011a) includes lowermost Jurassic shallow-water dolomites (DBS), evolving to Jurassic-Lower Cretaceous shallow-water (SWC) and margin-slope (MNM) limestones, Cretaceous calcareous breccias (IMM) and Scaglia-type deposits (SCA), calcareous breccia (CBI), and Eocene Scaglia Detritica deposits (SCAd).

At Zannone Island (Figure 2-6; Pantosti & Velonà, 1986), Norian dolomite (DBS) is covered by Maastrichtian-Eocene Scaglia-type limestones and marls with planktonic foraminifers (SCA; Segre, 1953). On the other hand, the Circeo succession (Pantosti & Velonà, 1986) comprises Lower Jurassic shallow-water limestones (SWC) passing to deep basin marls with ammonites (MAM) and cherty limestones (CLI), capped by Maastrichtian-Eocene Scaglia type deposits (Fogliano 001 well; (ViDEPI-Project, 1956a); (Lombardi, 1968). NW-Lepini and locally Simbruini-Ernicci Mts (ISPRA, 1975) include above the Upper Triassic-Cretaceous shallow-water carbonates (DBS and SWC), calcareous breccia (CBI) and, in the NW-Lepini Mts (ISPRA, 2021), the Eocene-Aquitania p.p. Scaglia-type deposits.

2.1.4 Apennine Platform central sector

The platform s.s. facies (Figure 2-4, Figure 2-5, Figure 2-6) include Norian to Upper Cretaceous dominantly shallow-water carbonates (DBS and SWC), exposed from the Pollino

Mts to the Volsci-Simbruini and L'Aquila Mts (D'Argenio et al., 1973; Damiani et al., 1991; Cipollari & Cosentino, 1995; Cosentino et al., 2002; Cosentino et al., 2003; Bonardi et al., 2009; Cosentino et al., 2010; Vitale & Ciarcia, 2013b), and references therein). In the Simbruini Mts (Filettino area), Rhaetian-Toarcian slope-to-basin cherty dolostones occur (Damiani et al., 1991).

In addition, some Triassic-Jurassic ephemeral intraplateau basin deposits are exposed in the Picentini Mts (MNM; Iannace et al., 2005) and the Cervati Mts (FOP; (De Alfieri et al., 1986) made of cherty dolostones and limestones, conglomerates, and marls. The northern sector of the Apennine Platform (Camposauro, Maggiore, Matese, Ausoni, and W-Marsica Mts) presents Albian and Cenomanian bauxite levels (Boni et al., 2012), on the other hand, in the central sector (Mt Soprano), only residual clays and terrestrial plant-rich fossils occur (Bartirromo et al., 2019). Generally, the shallow-water successions show a wide hiatus in the Late Cretaceous-Paleogene; however, in the southern sector (Cervati, SW-Alburni, and Soprano Mts) and locally in the northern part (Maggiore and Ausoni Mts) Eocene shallow-water limestones, conglomerates, and green marls are locally present (TRN; Selli, 1957; Vecchio, 2006).

2.1.5 Apennine Platform eastern margin

The eastern margin of the Apennine Platform includes the successions of Maddalena, NE-Alburni, Marzano, Laviano, Croce, SE-Camposauro, Sepino-Moschiatturo, N-Matese, Venafrò, Mainarde, Frosolone, E-Marsica and Gran Sasso Mts (Figure 2-1 a).

The Maddalena Mts (Figure 2-4; ISPRA, 2010a, 2014a; Vitale & Ciarcia, 2022), Mt Marzano and Laviano successions (ISPRA, 2016c; Vitale & Ciarcia, 2018) show shallow-water Norian-Upper Cretaceous successions (DBS and SWC) upward passing to slope Maastrichtian-Paleocene calcareous breccia (CBI) and Eocene Nummulitids and Alveolinids calcarenites (SCA and SCAd). The Mt Croce deposits (Figure 2-4), exposed in the Campagna tectonic window (Vitale et al., 2020b) and references therein), are characterized by Upper Triassic dolomite conglomerates (DBSc) covered by calcareous breccia (CBI) and Eocene-Aquitainian breccias (FFA), passing upward to Aquitainian-Burdigalian Scaglia-type rocks (FFP). The SE-Mt Camposauro succession (Figure 2-5; ISPRA, 2010b; Vitale & Ciarcia, 2022) and the Sepino- Moschiatturo deposits (Figure 2-5; Sgrosso, 1998; Vitale & Ciarcia, 2018, 2022), the N-Matese, Venafrò, and Mainarde Mts (Figure 2-6; Cosentino et al., 2010; Vitale & Ciarcia, 2022) include shallow-water Norian-Upper Cretaceous carbonates (DBS and SWC) covered by calcareous breccia (CBI) and Eocene Scaglia-type deposits. The Frosolone succession

(Figure 2-6; ISPRA, 2015) includes Upper Triassic cherty dolomites (DBSa; Frosolone 001 well; (ViDEPI-Project, 1961a) passing to Jurassic-Cretaceous cherty limestones (CLI), covered by calcareous breccia (CBI), Eocene Nummulitids, and Alveolinids calcarenites (FMO).

Upper Triassic-Cretaceous shallow-water carbonates (DBS and SWC), evolving to slope Upper Cretaceous-Paleocene calcareous breccia (CBI) and finally Eocene calcarenites (CFR) form Mainarde Mts and E-Mt Marsica (ISPRA, 2010c). Gran Sasso Mt succession (ISPRA, 2010d) includes at base Upper Triassic-Lower Jurassic shallow-water carbonates (DBS and SWC), passing upward to Lower Jurassic-Upper Cretaceous slope-to-basin bio- and litho-clastic limestones, marls, and jaspers (GRS). At Mt Prena-Vradda (Gran Sasso Mt), Norian-Raethian bituminous dolostones occur, formed in euxinic intra-platform troughs (Cosentino et al., 2010). The succession continues with Upper Cretaceous-Aquitania p.p. Scaglia-type deposits (SCA) locally and laterally passing to Campanian p.p.-Maastrichtian p.p. calcarenites (OFE; (ISPRA, 2010d).

2.1.6 Lagonegro-Molise Basin

The deep-sea sedimentary successions mainly exposed along the axial and external zones of the chain, from southern Abruzzo to northern Calabria, belong to the vast pelagic realm of the Lagonegro-Molise Basin (Merlini & Mostardini, 1986). These successions form the following tectonic units, from south to north: Lagonegro I, Lagonegro II, Groppa d'Anzi, Campomaggiore, Sannio, Fortore, Daunia, Vallone del Toro, Tufillo, and Agnone (Dazzaro et al., 1988; Pescatore et al., 1988; Pescatore et al., 2000; Basso et al., 2002; Patacca & Scandone, 2007; Bonardi et al., 2009; Vitale & Ciarcia, 2013b; Milli et al., 2021; Vitale & Ciarcia, 2022). The Lagonegro I and II Units (sensu (Scandone, 1967, 1972; Figure 2-4), otherwise named Mt Arioso Unit (Di Nocera et al., 2011; ISPRA, 2012) in Basilicata region, or Frigento Unit (Basso et al., 2002; ISPRA, 2011c) in Irpinia sector, consist, at the base, of Ladinian–Carnian slope to basin calcilutites and fine-grained sandstones with intercalations of reefal bodies (FAC), upward passing to Carnian–Norian cherty limestones (SLC), Rhaetian–Jurassic radiolarites and reddish, greenish silicified argillites (STS) and, finally, Lower Cretaceous dark silicified argillites with calcilutites and marls (FYG). Then, the succession evolves to slope-to-basin Upper Cretaceous-Burdigalian calciclastic coarse-to-fine sediments, marls, and varicoloured clays (FYR). The Groppa d'Anzi Unit (Figure 2-4; Pescatore et al., 1988) is formed by Upper Cretaceous (?) – Aquitania varicoloured clays and marls with calcilutites and calcarenites (AVG; ISPRA, 2007a, 2017a), laterally passing, in the upper part, to the middle Eocene-

Aquitanian marly limestones (CPA; (Selli, 1962; Pescatore et al., 2000), and Lower Miocene volcanoclastics-bearing sandstones (PDO). The Lower Cretaceous-Miocene Campomaggiore Unit (Pescatore et al., 1988) is exposed in the southern external zones of the chain. It comprises the Vaglio and San Chirico Units (ISPRA, 2017a), which are made of FYG and FYR.

The Sannio Unit (Figure 2-5; Pescatore et al., 1988), otherwise named Coltre Sannitica (Selli, 1962), is composed of Upper Cretaceous–upper Burdigalian reworked carbonate sediments (FYR), including, from bottom to the top, jaspers calcareous breccias and calcarenites with macroforaminifera, marls, and varicoloured clays (Basso et al., 2002; Vitale & Ciarcia, 2022). The lower part of the pre-orogenic succession of the Fortore Unit (Pescatore et al., 2000) is the same as that of the Groppa d’Anzi Unit. However, starting in Burdigalian, the succession evolved into volcanoclastic deposits (TUT; Ogniben, 1960; Pescatore et al., 1999; Pescatore et al., 2000). A pre-orogenic dominantly calciclastic succession characterizes the Daunia Unit (Dazzaro et al., 1988), showing at the base upper Oligocene–middle Burdigalian clays, varicoloured marls, and whitish calcilutites and calcarenites (SID; Dazzaro et al., 1988; Santo & Senatore, 1988). The Vallone del Toro Unit (Basso et al., 2002; ISPRA, 2011b) is formed by upper Oligocene–middle Burdigalian calcarenites, reddish and greenish argillites and calcareous breccias (SFN; Crostella & Vezzani, 1964). The pre-orogenic rocks of Agnone and Tufillo Units (Selli, 1962; ISPRA, 2011d) consist of Upper Cretaceous-upper Burdigalian varicoloured clays (AVS).

2.1.7 Apulian Platform

The Apulian Platform domain encompasses, in the central and southern sector, a buried imbricate succession intensively drilled for hydrocarbon exploration (e.g., Butler et al., 2004; Sciamanna et al., 2004; Shiner et al., 2004; ViDEPI-Project, 2009–2015). However, only a small part is exposed (Mt Alpi; Mazzoli & Butler, 2006). The southern sector (Figure 2-5) is characterized by upper Permian sandstone (PUG1, Puglia 001 well, Butler et al., 2004); Figure 2-1 a), Lower Triassic polygenic breccias (PUG2), Upper Triassic shallow-water dolostones (DBS), Jurassic-Cretaceous carbonates (SWC). The inner and outer buried Apulian Platform continues locally with Maastrichtian-Paleocene Scaglia-type deposits (SCA) and Eocene breccias (BLA; Patacca & Scandone, 2014) covered by Aquitanian-Langhian calcarenites (PLE). In the Murge and Salento areas (foreland; (Ciaranfi et al., 1988; ISPRA, 2013), Maastrichtian carbonates pass locally and laterally to slope bioclastic calcarenites (ALT1). The succession continues with middle-upper Eocene calcilutites and calcarenites (TTG), upper Oligocene limestones (CSR, PBD), and Langhian calcarenites (BSM).

The central sector (Figure 2-6) is similar for the inner and outer buried successions, whereas, for the foreland, the Gargano promontory is characterized by Paleozoic shales, dolostones and limestones (GAR1, Gargano 001 well, ViDEPI-Project, 1982; Patacca & Scandone, 2007); Figure 2-1 a), followed by Lower Triassic sandstones, limestones and dolostones (GAR2). The stratigraphic succession (Morsilli et al., 2017; ISPRA, 2017b) continues with two heteropic sequences with the western side made of Upper Triassic-Lower Cretaceous shallow-water carbonates (DBS and SWC) with evaporites at the base (Patacca & Scandone, 2007), covered by Oligocene breccias (BCG), Chattian-Aquitania p.p. limestones (SMH), Burdigalian-lower Langhian limestones (TRF), upper Langhian-upper Tortonian calcarenites (PLE), lower Messinian marls (TPL). The eastern side of Gargano consists of Lower Cretaceous deep-basin limestones (MAI), marls (FUC), Upper Cretaceous Scaglia-type deposits (SCAd and SCA), Eocene calcarenites with nummulitids (PES).

In the northern sector, imbricate successions of the inner and outer Apulian Platform largely outcrop, including Morrone, La Queglia, and Majella Mts (Cosentino et al., 2003; Patacca & Scandone, 2007; Vezzani et al., 2010). Morrone succession (ISPRA, 2005b) consists of Lower Jurassic-Upper Cretaceous shallow-water carbonates (SWC) evolving upward to calcareous breccia (OFE) and Oligocene-Lower Miocene ramp calcarenites with macroforaminifers (CFR).

The La Queglia succession (Figure 2-6; Bigi & Di Bucci, 1987; Patacca et al., 1992a; ISPRA, 2005c; Patacca & Scandone, 2007) is made of Upper Cretaceous-Eocene p.p. cherty limestones (CLI), hosting, in the upper part, an anorogenic lamprophyre, passing upward to upper Oligocene?-Aquitania distal ramp bioclastic limestones (CFR). The Mt Majella includes two heteropic successions (Patacca & Scandone, 2007; Patacca & Scandone, 2021). In the southern sector, the Cretaceous pre-orogenic succession includes shallow-water limestones (SWC), which host at least three levels of bauxite and are marked by subaerial exposure. In contrast, on the northern side, the base is made by a pelagic succession, including pelagic limestones (MAI) at the base, passing upward to marls (FUC) and Scaglia-type deposits (SCA). Both successions are covered by the Campanian-Maastrichtian calcarenites and calcirudites (OFE; Eberli et al., 2019). This deposit passes to Paleocene-Oligocene deep water breccias (SSG), upper Oligocene *Lepidocyclina* limestones (LCL), and Burdigalian Bryozoan limestones (BRY).

The eastern part is buried under the allochthonous wedge in the northern sector. It is made of Lower Jurassic-Cretaceous shallow-water carbonates (SWC) locally with residual clays passing upward to Lower Miocene marls and bioclastic limestones (SCAd) (Casoli 001 well,

(ViDEPI-Project, 1955); Castelfrentano 005 well, (ViDEPI-Project, 1960a); Castelfrentano 006 well, (ViDEPI-Project, 1961b); Orsogna 001 well, (ViDEPI-Project, 1981); Marcone 001 well, (ViDEPI-Project, 1960b); Figure 2-1 a). The foreland succession is formed by Upper Jurassic-Lower Cretaceous shallow-water carbonates passing upward to Oligocene-Lower Miocene bioclastic limestones (LBG) (Casaborselli 001, (ViDEPI-Project, 1956b); Villalfonsina 001, (ViDEPI-Project, 1958).

2.2 Metamorphism and tectonic exhumation

Palaeozoic rocks of the Sila and Stilo Units are affected by variable Variscan metamorphism, ranging from anchizone to greenschist, amphibolite, and granulite facies (Bonardi et al., 2001; Langone et al., 2010; Ortolano et al., 2020). Castagna Unit (Bonardi et al., 2001; Tursi et al., 2020) is characterized by Variscan amphibolite facies with an Alpine metamorphic overprint defined by a lower amphibolite-facies (480-560 °C and 0.7-0.9 GPa; Brandt & Schenk, 2021). The Paleozoic basement of the Bagni Unit (Bonardi et al., 2001), is affected by a pre-Alpine greenschist metamorphism, overprinted by Alpine lower amphibolite-facies conditions (545-565 °C and 0.71-0.77 GPa; (Brandt & Schenk, 2021). Castagna and Bagni Units show similar clockwise P–T paths for prograde Alpine metamorphism (Figure 2-7). The age of the Alpine metamorphism of the Castagna Unit ranges between 64.4 ± 3.9 Ma (Ar–Ar biotite dating, (Rossetti et al., 2001) and 43 ± 1 Ma (Rb–Sr biotite dating; (Schenk, 1980).

Ligurian successions of Frido, Diamante-Terranova, Gimigliano and Malvito Units are overprinted by an Alpine HP/LT metamorphism (Figure 2-7). Phyllites of the Frido Unit host carpholite in quartz veins (Vitale & Ciarcia, 2013b), aragonite in the calcschists (Spadea, 1976), and blue-amphibole in metadolerites (Sansone et al., 2012) with peak conditions of 330-360 °C and 1.2-1.4 GPa (Vitale & Ciarcia, 2013b). The Gimigliano Unit also hosts HP/LT carpholite/chloritoid assemblages in the metabasalt cover (Rossetti et al., 2004).

Blueschist metabasalts of the Diamante-Terranova Unit show an assemblage of blue Na-amphibole, lawsonite, chlorite and pumpellyite (Liberi et al., 2006; Liberi & Piluso, 2009; Fedele et al., 2018) that furnishes a peak condition of 350-390 °C and 1.1-0.9 GPa (Figure 2-7). However, Tursi et al. (2020) reported in the blueschists the occurrence of zoned clinopyroxene crystals, showing core–rim compositional variation from diopside to omphacite that provide peak conditions of 2.0-2.1 GPa and 475-490 °C (Figure 2-7) and decompression and cooling path characterized by ~ 380 °C and ~ 1.1 GPa and finally a mylonitic deformation under 290-315 °C and ~ 0.7 GPa conditions. The metamorphic peak age is 46.7 ± 0.4 Ma

(Shimabukuro et al., 2012). Rossetti et al., 2001; Rossetti et al., 2004, reported a late Eocene-Oligocene (ca. 33-38 Ma) Ar-Ar phengite age data for the HP/LT Gimigliano Unit. Due to the LT environment of peak Alpine metamorphism and microtextural evidence, these Ar ages should be regarded as a minimum estimate of the crystallization age of the high-pressure phengites during subduction metamorphism in the oceanic-derived units in Calabria.

Metabasalts, metapelites, and quartzites of Gimigliano Unit show assemblages including crossite and phengite, suggesting a metamorphic peak of 350 °C and 1.0 GPa (Liberi et al., 2006), followed by a nearly isothermal decompression to greenschist facies conditions of 380 °C and 0.6 GPa, and then 300 °C and 0.3 GPa (Liberi et al., 2006). Finally, Malvito metabasalts with lawsonite and phengite indicate a metamorphic climax at 350 °C and 0.8–0.9 GPa (Figure 2-7).

The only successions of the Apennine Platform affected by metamorphism are the Lungro-Verbicaro and Cetraro Units. A high pressure-low temperature metamorphism defines the former unit characterized by assemblages of Fe-Mg-carpholite and phengite in the Lower Miocene metapelites (SFL) and chloritoid and phengite in the Anisian–Ladinian phyllites (PMM, Iannace et al., 2007). The peak thermobaric conditions for the upper and lower parts of the succession are 320 °C and 1.3 GPa, and 390 °C and 1.4 GPa, respectively. Subsequently, during the decompression path, an increase in temperature characterized both upper and lower parts, reaching 340 and 400 °C, respectively (Figure 2-7). The retrograde decompression is recorded by the growth of chlorite and phengitic white mica as alteration products of both carpholite and chloritoid. Finally, the Cetraro Unit was affected by greenschist-facies metamorphism defined by P-T conditions of 380 °C and 0.4 GPa (Iannace et al., 2007).

Regarding the tectonic exhumation following the orogenic burial, temporal constraints come from the rock thermochronology studies. Data have been collected using the Zircon (Zr) and Apatite (Ap) fission-track methods on CPT and continental Adria units. The Zr ages of exhumation for the Castagna and Bagni Units (Thomson, 1998) range between 21 and 29 Ma (late Oligocene-early Aquitanian), whereas the Ap ages fall in the interval 11-15 Ma (late Burdigalian-early Serravallian). Apatite fission-track data indicate for the Lungro-Verbicaro Unit a Tortonian age (Iannace et al., 2007), for the Lagonegro-Molise Basin Units (Mazzoli et al., 2008), two main age clusters, Messinian-Zanclean (5.5-5.0 Ma) and late Zanclean (3.9-3.8 Ma). Finally, for the Mt Alpi (Apulian Platform), ages between 3.2 and 2.0 Ma were estimated (Mazzoli & Butler, 2006).

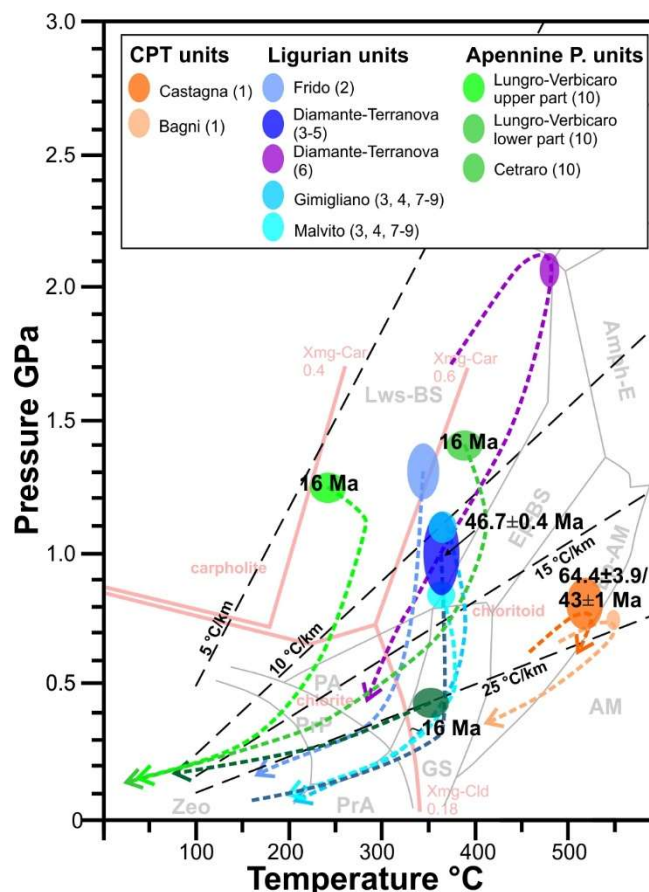


Figure 2-7: P-T diagram of the Alpine metamorphic paths of the CPT, Ligurian and Apennine Units. (1) (Vitale et al., 2013a; Vitale et al., 2013b; Brandt & Schenk, 2021); (3) (Liberi & Piluso, 2009); (4) (Liberi et al., 2006); (5) (Fedele et al., 2018); (6) (Tursi et al., 2020); (7-9) (Rossetti et al., 2001; Rossetti et al., 2004); (10) (Iannace et al., 2007). P-T ranges for the metamorphic facies are modified after (Rossetti et al., 2004) and (Tsuji-mori & Ernst, 2014). AM = amphibolite; Amph-E = amphibole-eclogite; Ep-AM = Ep-amphibolite; Ep-BS = Ep-blueschist; Ep-E = Ep-eclogite; HP-G = HP granulite; GS = greenschist; LPG = LP granulite; Lws-BS = Lws-blueschist; Lws-E = Lws-eclogite; PA = pumpellyite-actinolite; PrA = prehnite-actinolite; PrP = prehnite-pumpellyite; Zeo = zeolite. Figure from Ciarcia and Vitale, 2025.

2.3 Tectonic vergences

Figure 2-8 a, shows the geologic map of the northern Calabria-southern Apennines, reporting the tectonic transport directions in present-day coordinates deduced from the literature (Cerisola & Montone, 1992; Naso & Tallini, 1993; Hippolyte et al., 1994; Naso et al., 1995; Scrocca et al., 1995; De Corso et al., 1998; Scrocca & Tozzi, 1999; Sani et al., 2004; ISPRA, 2015); (Langone et al., 2010; Vezzani et al., 2010; Vitale et al., 2013a; Vitale & Ciarcia, 2013b; Vitale et al., 2019a). The dataset includes tectonic transport directions inferred from different kinematic indicators. In the case of dominant brittle deformation, they include fold asymmetry, S-C structures, Riedel shears, and shear veins associated with thrust faults. On the other hand, in the case of ductile strains such as those occurring in mylonites, we used to stretch lineations, rotated porphyroclasts, S-C structures and synthetic and antithetic S-C structures. According to the ages inferred by the authors, we grouped the tectonic vergences in three intervals: (i)

Eocene-Langhian, (ii) Serravallian-Pliocene, and (iii) Early Middle Pleistocene. The first interval has been selected because it corresponds to the tectonic units transported during the first stage of orogenic evolution. They include the non-metamorphic Ligurian Units tectonically verging mainly to E/SE and the metamorphic units of Calabria-Peloritani Terrane with tectonic transport vectors ranging between SE and NE (Carrara & Zuffa, 1976; Faure, 1980; Dietrich, 1988; Langone et al., 2010; Ortolano et al., 2020) (ii) Ligurian meta-ophiolites showing vergences to E-SE (Vitale et al., 2013a; Vitale et al., 2019a); (iii) metamorphic Apennine Platform units with tectonic transports to E-SE (Vitale et al., 2020). For the metamorphic units, we also added the tectonic transport (hanging wall movement) associated with extensional shear during the tectonic exhumation under syn- or post-greenschist metamorphism (Rossetti et al., 2001).

The Serravallian-Pliocene tectonic transport vectors refer to the movement of the thrust sheets in the southern Apennines (ranging between NW to E; Cerisola & Montone, 1992; Scrocca et al., 1995; De Corso et al., 1998; Sani et al., 2004; Vezzani et al., 2010; Giannandrea et al., 2014; ISPRA, 2015). This large variability of tectonic vergences in this period is related to the out-of-sequence thrusts that frequently show two about orthogonal vergences (e.g. Vitale et al. (2020a) and to the arcuate nature of the thrust systems (Ghisetti & Vezzani, 1988). The Pleistocene vergences regard only the thick-skinned thrusts involving Pleistocene deposits primarily in the external sectors of the southern Apennines with dominant E-NE-vergence (e.g., Hippolyte et al. (1994).

Figure 2-8 b-e, shows the rose diagrams of the directions of the tectonic vergences grouped for the three temporal intervals. For the Eocene-Langhian data, there is a prevalence of the E-W (Calabria-Peloritani Terrane; Figure 2-8 b) and NW-SE (southern Apennines; Figure 2-8 c) directions; for the Serravallian-Pliocene time, there are two orthogonal directions, the first more developed ENE-WSW and the secondary NNW-SSE (Figure 2-8 d). Finally, the Pleistocene directions are constantly ENE-WSW (Figure 2-8 e). However, all these data are referred to the present-day coordinates. Taking into account the different rotations of the various sectors of the chain that occurred during the orogenic evolution in the Eocene-Miocene interval, a CCW rotation of 90° has to be considered for the Calabria-Peloritani Terrane (Siravo et al., 2022) that leads to an original N-S direction (Figure 2-8 f) and a CCW rotation of about 60° for the Cilento/Basilicata region (Mattei et al., 2007) and references therein) with again an original N-S direction (Figure 2-8 g). For the Serravallian-Pliocene time, a CCW rotation of about 30° occurred for the southern Apennines (Mattei et al., 2007 and references therein) with original E-W and the secondary N-S directions (Figure 2-8 h). Finally, during the Pleistocene,

CCW rotations of 20° to 0° occurred in the southern Apennines (Mattei et al., 2007), with a restored E-W direction (Figure 2-8 i). It is worth noting that at the passage between Langhian and Serravallian, the direction of the restored tectonic vergences changes by about 90° , from ca. N-S to ca. E-W.

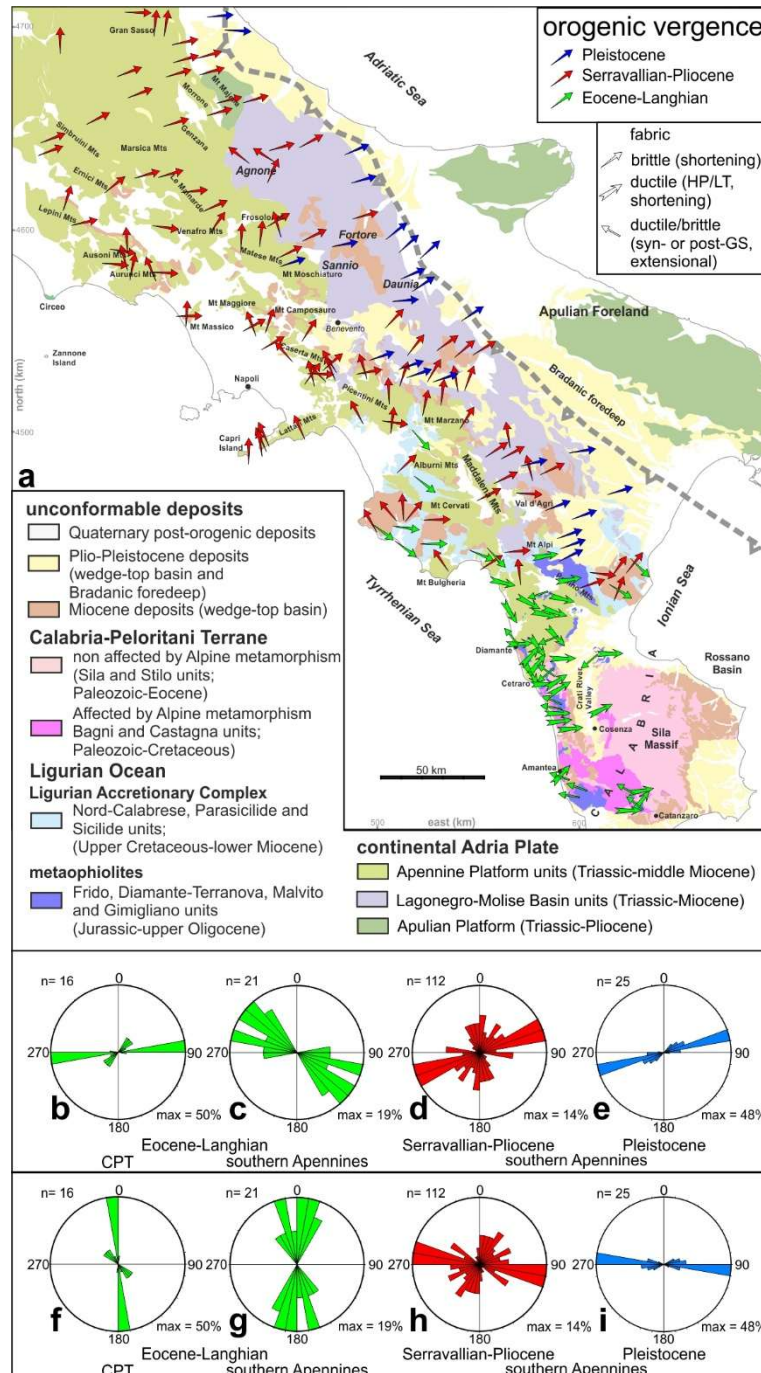


Figure 2-8: Tectonic transport map. HP/LT: high pressure-low temperature metamorphism; GS: greenschist metamorphism. Figure from Ciarcia and Vitale (2025).

2.4 Orogenic evolution

The proposed orogenic reconstruction is in line with the model of "Ancient Apennines" (Alvarez, 1991), and those of other circum-Mediterranean orogens, such as the Betic-Rif chain in the western sector, where a similar tectonic structure is reported (Chalouan & Michard, 1990; Michard et al., 2006; Vitale et al., 2015), and new data on the Alpine tectono-metamorphic evolution (Marrone et al., 2021) age of the subduction zone metamorphism (38 Ma late Eocene-Oligocene; (Bessière et al., 2022) and Variscan inheritance (e.g., Rossetti et al. (2020) are made recently available from the continental-derived units that form the Alpine tectonic edifice.

The orogenic evolution can be summarized as follows (Rossetti et al., 2001; Iannace et al., 2007; Liberi & Piluso, 2009; Fedele et al., 2018; Vitale et al., 2019a). In the Palaeocene-Eocene time, the subduction of the Ligurian lithosphere under the European Plate, including the Calabria-Peloritani Terrane and Sardinia-Corsica block, carried at depth some crust slices of both downgoing (Ligurian meta-ophiolites) and overriding (Castagna and Bagni Units) plates. The complete closure of the oceanic realm in the Early Miocene culminated in the frontal accretion of the eastern sector of the ocean and the formation of the Ligurian Accretionary Complex. After that, the subduction zone involved the western margin of the Apennine Platform with the subduction of some successions (Lungro-Verbicaro and Cetraro Units). From the Middle Miocene to the Middle Pleistocene, a thrust-and-fold belt formed as the continental Adria plate was overthrust by different successions, with both thin- and thick-skinned tectonics. From the Oligocene to the Tortonian, the subducted tectonic units of the Calabria-Peloritani Terrane, the Ligurian Ocean, and the continental Adria plate were exhumed and accreted into the orogenic chain. Therefore, different geodynamic mechanisms have shaped the orogenic evolution, acting synchronously or diachronously. We can describe the evolution of this collisional margin through the migration of the subduction channel boundaries (i.e., roof and basal décollements). Different geodynamic scenarios are associated with the movement of the main subduction channel décollement (Vannucchi et al., 2012). This surface can move up or down towards the foreland or hinterland. It follows that the subduction channel can be accreted or eroded, or these mechanisms can act in the frontal sector. To better describe this process, we illustrate the geodynamic evolution of the northern Calabria-southern Apennines (southern sector) in the cartoon of Figure 2-9.

Following the start of the subduction of the Ligurian Ocean lithosphere, the roof décollement migrated upward, inserting slices of the Calabria-Peloritani Terrane in the subduction channel (upward basal erosion; Figure 2-9 a). Subsequently, the roof décollement migrated downward, adding oceanic slices to the subduction channel (downward accretion; Figure 2-9 b, c). These

three steps occurred in the Paleocene-Eocene interval, chronologically constrained by the HP/LT metamorphic ages of Castagna (43 ± 1 Ma, (Schenk, 1980); 64.4 ± 3.9 Ma, (Rossetti et al., 2001) and ophiolite units ($33\text{--}38$ Ma; (Rossetti et al., 2001; Rossetti et al., 2004); 46.7 ± 0.4 Ma; (Shimabukuro et al., 2012). Synchronously, orogenic volcanism occurred in Sardinia, with the oldest dated at 38.3 ± 0.3 Ma (Lustrino et al., 2009), consistent with the onset of oceanic subduction within the interval $49\text{--}42$ Ma (Ypresian-Lutetian; Lustrino et al., 2009).

In the Oligocene-Aquitainian (Figure 2-9 d), the easternmost ophiolite domain (Frido Unit) was subducted, reaching HP/LT metamorphic conditions; meanwhile, downward accretion formed an orogenic stack that triggered extension on the top of the overriding plate. This event was witnessed by the age of the zircon fission tracks in the subducted units ($23\text{--}28$ Ma; (Thomson, 1998) and by forming the wedge-top basin deposits of Paludi Fm (PLU). In the early Burdigalian (Figure 2-9 e), the roof decollement moved toward the foreland with the frontal accretion of the Nord-Calabrese (NCA), Parasicilide (PAS), and Sicilide (SIC) Units (LAC). In the Burdigalian (Figure 2-9 f), the oceanic accretionary wedge (LAC) overthickened following the buttressing of the western margin of the Apennine Platform (Lungro-Verbicaro Unit, LVU) and collapsed (Figure 2-9 g), forming an extensional wedge-top basin where the Cilento Group (CIG) deposited since the Burdigalian-Langhian boundary (Vitale et al., 2010; Corrado et al., 2019; Mehmood et al., 2023). In the Burdigalian-Langhian interval, the LVU subducted (Figure 2-9 g). Apatite fission track ages in Calabria-Peloritani Terrane rocks (Thomson, 1998) indicate that the tectonic exhumation of the subducted units continued in this period. In the Langhian-Serravallian time (Figure 2-9 h), the Apennine Platform remnant was overthrust by the LAC concurrently, and the LVU was accreted in the subduction channel and started the upward migration (tectonic exhumation) with the other metamorphic thrust sheets. The fast exhumation within the subduction channel was enhanced by the synchronous rifting of the Tyrrhenian Sea (e.g., Sartori et al., 2001), which led to the exhumation of the entire subduction channel primarily through upper-crustal extension. This process resulted in the exposure of metamorphic successions and in the deposition of clastic supply within the surrounding structural lows, presently forming the Serravallian-Tortonian successions of the Amantea and Rossano Basin deposits in northern Calabria.

According to Faccenna et al. (2001a,b), during the Serravallian, the subducted oceanic slab reached the 660 km upper/lower mantle boundary, lying down along this buoyancy barrier. This event slowed down the rollback migration rate and made the downgoing slab vertical. The renewed slab retreat in Tortonian time (Figure 2-9 i) occurred with the subduction of the thicker continental Adria lithosphere underlying the Apennine Platform, which became the new

backstop of the subduction system and triggered a forelandward jump of the main subduction detachment and thick-skinned thrusting. In this stage, the Apennine Platform Units and the Paleozoic basement tectonically covered the Lagonegro-Molise Basin successions that, subsequently (Figure 2-9 j, k), were accreted through the forelandward offscraping mechanism and dominant thin-skinned tectonics (Mattioni et al., 2006). In the Serravallian-Tortonian time, the wedge-top basins located in various places of the orogenic wedge were fed by calci-, silici- and volcanoclastic sediments, including Cilento (CIG), Castelvete (CVTG) Groups and Rossano Basin deposits. In the late Messinian, evaporite-bearing calciclastic sedimentation occurred in the Amantea and Rossano basins in northern Calabria, some wedge-top basins in the southern Apennines (ALTG), and the foredeep of the Vallone del Toro Unit.

In the Pliocene (Figure 2-9 i), the involvement of the thicker continental Adria lithosphere, located under the Apulian Platform, newly created a backstop for the subducting system, causing a new thick-skinned thrusting (e.g., Mazzoli et al., 2013). This important orogenic event was probably associated with progressive slab-break off and slab-tear propagation since ca. 4 Ma (Ascione et al., 2012) that ruled the formation and SE-migration of the wedge-top basin depocenters and, in general, an isostatic rebound and uplift. Synchronously with the thrust front E-migration, extension at the rear occurred, affecting more areas of the orogenic wedge. Finally, in the southern Apennines, the crustal shortening stopped in the Middle Pleistocene, as recorded by the synfolding deposits sedimented in the wedge-top basin of Sant'Arcangelo (Sabato et al., 2005), and a widespread extension in the chain occurred up to the present-day (Macchiavelli et al., 2012).

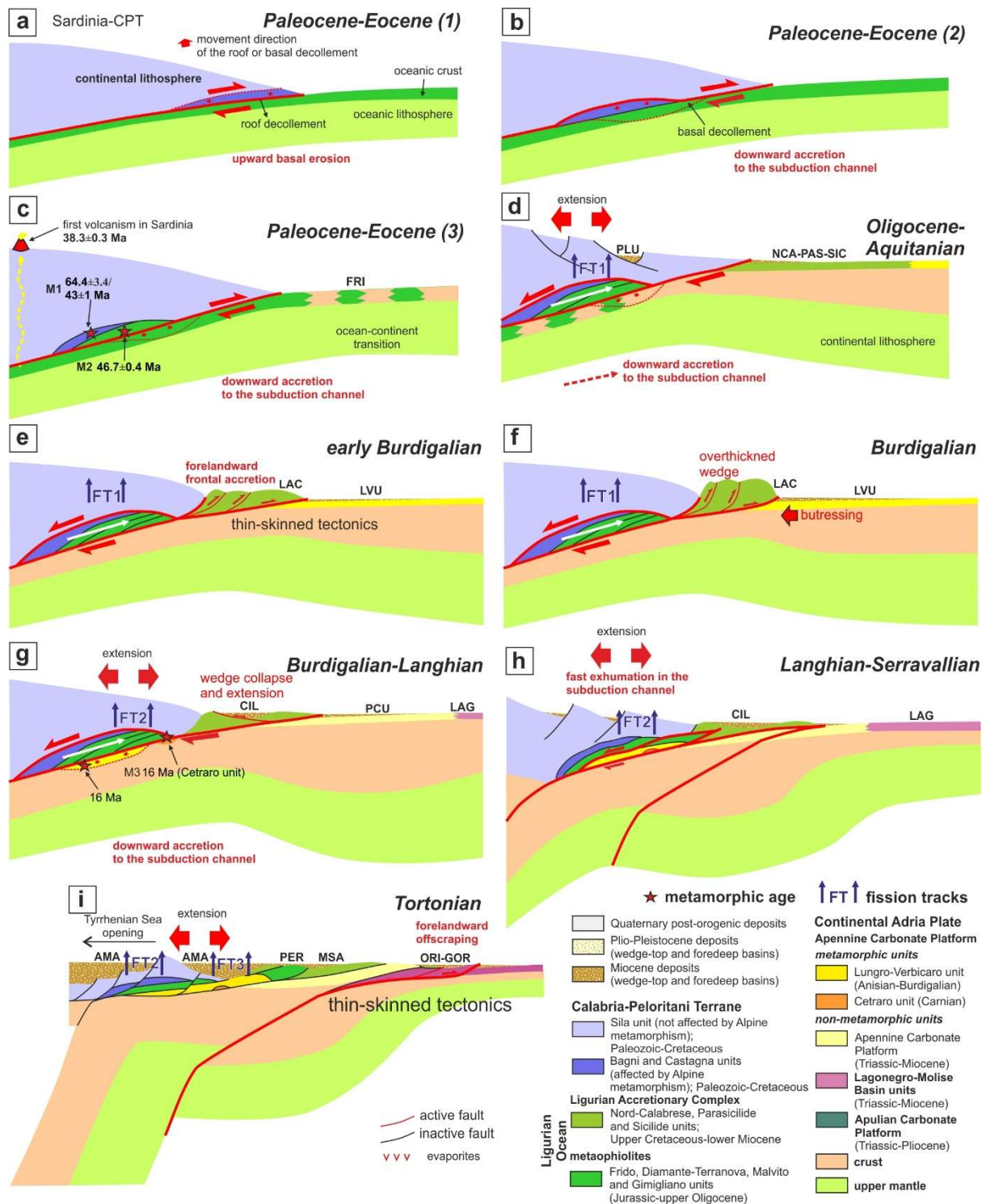


Figure 2-9: Cartoon showing the tectonic evolution of the northern Calabria-southern Apennines system from Paleocene to Recent. Not to scale. AMA: Amantea; CIG: Cilento Group; CVTG: Castelvetero Group; FRI: Frido Unit; LAC: Ligurian Accretionary Complex; LAG: Lagonegro-Molise Units; LVU: Lungro-Verbicaro Unit; NCA: Nord-Calabrese Unit; PAS: Parasicilide Unit; PCU: Pollino-Ciagola Unit; SIC: Sicilide Unit. Figure from Ciarcia and Vitale (2025).

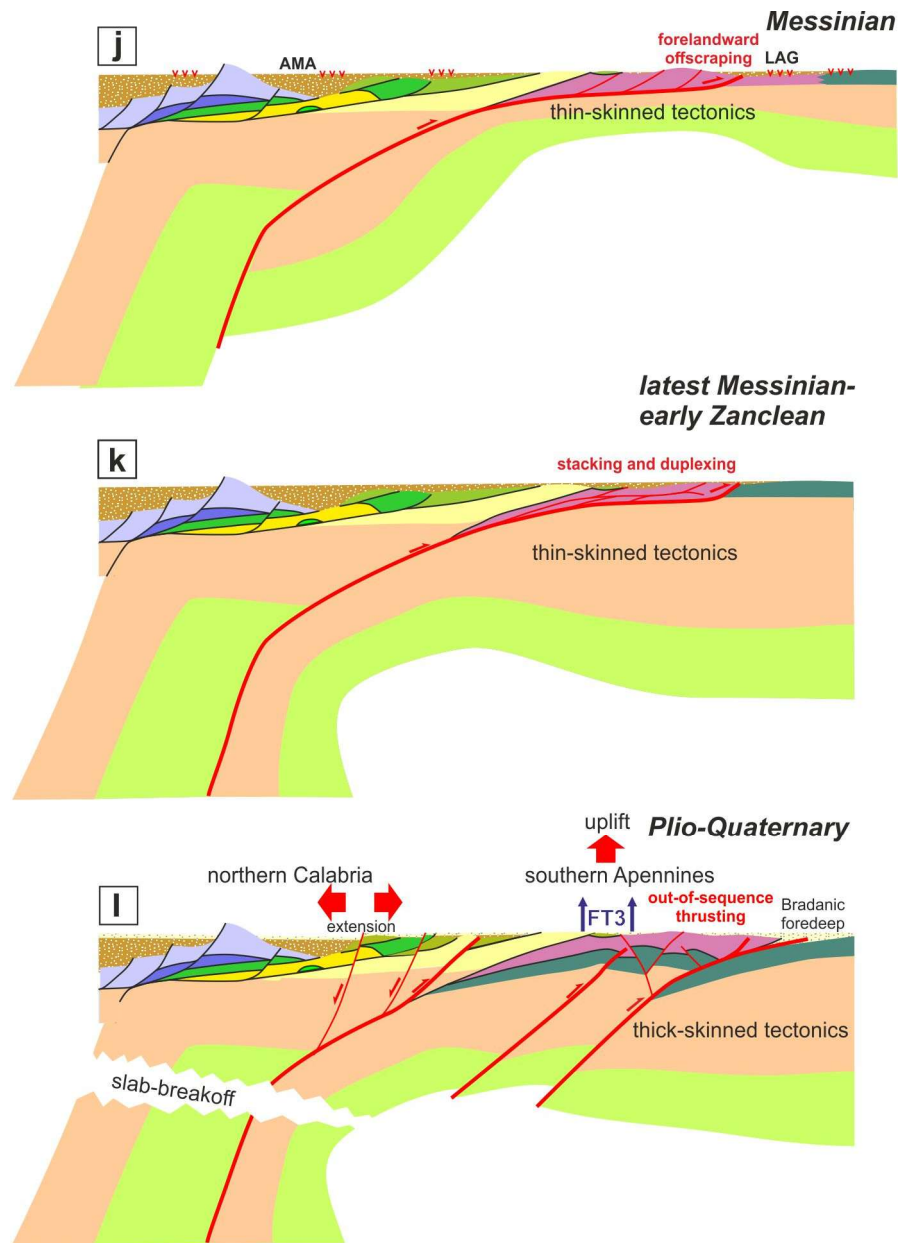


Figure 9: Continued.

2.4.1 Orogenic evolution of NCSA system within the western-central Mediterranean orogenic framework

Finally, to insert the NCSA within the framework of the orogenic evolution of the western-central peri-Mediterranean Alpine chains, we illustrate some paleogeographic maps starting from the middle Eocene, when the western side of the Ligurian oceanic lithosphere was downgoing under the European plate (Figure 2-10 a). Due to the relatively low convergence rate between African and European plates from Late Cretaceous to Recent (Dewey et al., 1989; Faccenna et al., 2001a; Faccenna et al., 2001b), the subduction of the oceanic lithosphere was

probably driven by its negative buoyancy, resulting in roll-back migration of the foredeep-thrust front system (Malinverno & Ryan, 1986; Faccenna et al., 1996; van Hinsbergen et al., 2014).

In the early Oligocene (Figure 2-10 b), the subduction of the oceanic lithosphere under the European plate produced the first manifestations of orogenic volcanism in Sardinia (Savelli, 2002; Carminati & Doglioni, 2012), with the maximum intensity in the Aquitanian (Figure 2-10 c). From this time, the Sardinia-Corsica block anticlockwise rotated, with the southern edge moving along a transform/strike-slip fault. In the Langhian (Figure 2-10 d), the Ligurian Ocean was completely closed, whereas the rotation of the Sardinia-Corsica block caused the opening of the Ligurian-Provençal basin. In the Early Pliocene (Figure 2-10 e), the Apennine Platform was incorporated in the orogenic wedge, and the Calabria-Peloritani Terrane (CPT), previously detached from the Sardinia-Corsica block, migrated SE-ward with the opening of the back-arc basin of the southern Tyrrhenian Sea. At this time, the oceanic lithosphere of the Ionian Sea was subducting under the CPT. Strike-slip faults allowed the CPT to move, including lithospheric structures separating the movements in northern and southern Calabria. The SE migration of the CPT caused the stretching of the lithosphere in the rear and the formation of the Vavilov Basin (Savelli, 2002). The Pleistocene palaeogeography (Figure 2-10 f) depicts the last stage of the orogenic evolution characterized by (i) the CPT ending its orogenic migration with the Marsili Basin formation at the rear (Savelli, 2002) and creating a mountain bridge between the southern Apennines and the Sicilian Maghrebids, (ii) the subduction of the Ionian lithosphere under Calabria; and (iii) the formation of the orogenic volcanic arc of the Eolian Islands. In contrast, the crustal shortening is no longer active in the southern Apennines. Currently, the extension affects this sector through volcanic activity, especially along the eastern side of the Tyrrhenian Sea.

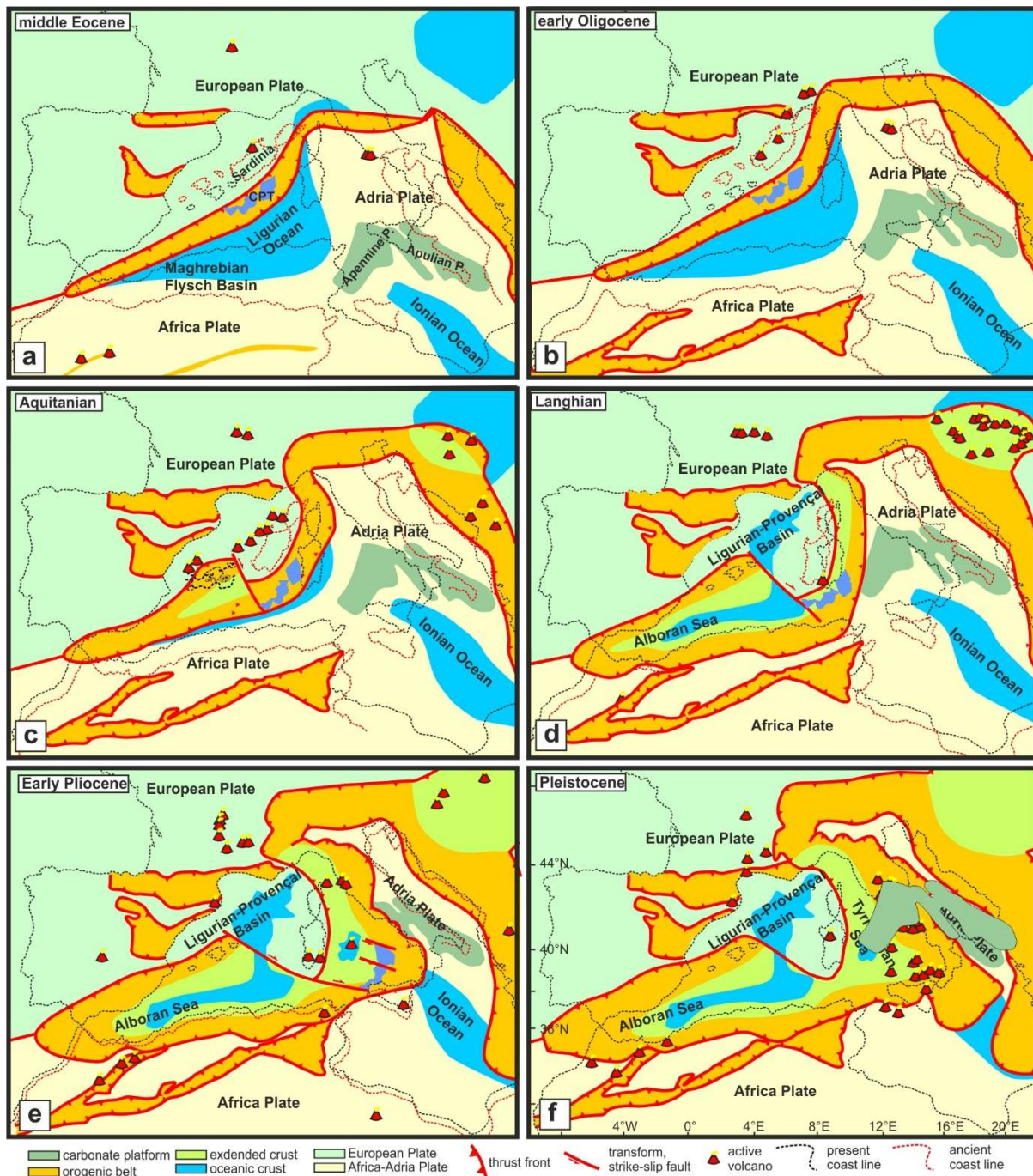


Figure 2-10: (a-f) Cartoons showing paleogeographic maps of the western-central Mediterranean area from the middle Eocene to Pleistocene. Figure from Ciarcia and Vitale (2025).

2.5 Cilento

2.5.1 Ligurian Accretionary Complex

The Ligurian Accretionary Complex (LAC) is a thrust-sheet pile formed by deep-basin successions, locally including oceanic-to-continental lithospheric rocks, cropping out from northern Calabria to the northern Apennines (Figure 2-11).

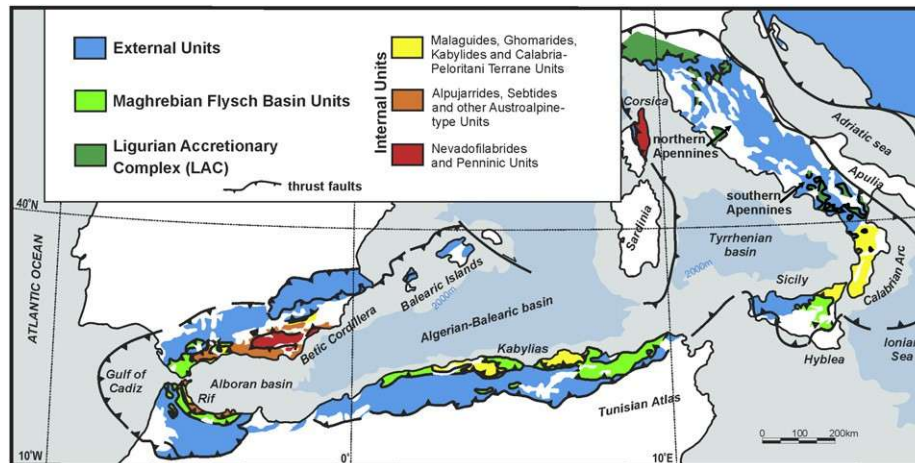


Figure 2-11: Schematic tectonic map of the circum-Mediterranean orogenic belts with exposure of the LAC (Vitale et al., 2013b).

In the southern Apennines, it occupies the highest tectonic position encompassing: (i) units detached from their pre-Cretaceous successions (Nord-Calabrese, Parasiclide and Sicilide Units, (Monaco et al., 1991; Ciarcia et al., 2009a; Ciarcia et al., 2012; Vitale & Ciarcia, 2013b) and references therein) and (ii) the HP/VLT Frido Unit characterized by an OCT (Ocean Continent Transition) basement and a metasedimentary basin succession (Knott, 1994; Vitale & Ciarcia, 2013b). The former units were piled up by means of frontal accretion mechanisms during the Burdigalian time (Ciarcia et al., 2012), whereas the Frido Unit subducted in the Late Oligocene and successively exhumed and intruded into the tectonic prism before the middle Tortonian (Vitale & Ciarcia, 2013b). In turn, LAC overlays an orogenic wedge formed by Mesozoic-Tertiary successions, more or less detached from their Paleozoic substrate, encompassing both platform carbonates (Apulian and Apennine Platforms, (Merlini & Mostardini, 1986; Vitale & Ciarcia, 2013b) and basin successions (Lagonegro and Molise Basin Units, M (Merlini & Mostardini, 1986), here named External Units (Figure 2-5) in analogy with the tectonic units occupying the same structural position in other circum-Mediterranean chains (e.g. Guerrera et al. (2005).

My research is focused on the Nord-Calabrese unit of the LAC. The Nord-Calabrese Unit includes the formations of Crete Nere and Saraceno (Figure 2-12; Bonardi et al., 1988a). The Crete Nere succession consists, in the lower part, of oceanic and continental masses (Vitale & Ciarcia, 2013b), the former formed by ophiolites locally preserving a coherent succession (“Timpa delle Murgie” ophiolites, Bonardi et al., 1988a), including gabbros, dolerites, pillow lavas, pillow breccias, and a deep basin cover formed by argillites, quartz-arenites, jaspers and allodapic limestones. The latter, named also “Calcarei di Mezzana” (Bousquet, 1973),

corresponds to scaglia-type deposits, directly covering pillow lavas and pillow breccias. Continental bodies consist of subcontinental serpentinites, gneisses and amphibolites cropping out only in the “Timpa di Pietrasasso” locality. The succession continues at the base with dark-brownish argillites alternating with grey-greenish quartzarenites, followed by a thick succession of black shales with intercalations, in the upper part, of arenites and calcareous beds. The age of the middle-upper part of this formation is Middle Eocene; however, the age of the lower part could reach the Upper Cretaceous, as suggested by Bonardi et al. (1988a). The Crete Nere Fm. gradually passes upward to the Saraceno Fm. consisting of four members (Ciarcia et al., 2012), from bottom to top: (i) calciclastic, locally silicified, arenitic turbidites with dark chert lenses and rare lithic sandstones (Punta Telegrafo member); (ii) thin layers of calciclastic, pelitic and arenitic turbidites with lenses and nodules of dark chert and subordinately arkosic-lithic sandstones (Terranova di Pollino member); (iii) alternance of marly, silty and arenitic beds, occasionally with dark chert nodules and layers of microbreccia at the top (Carpineta member); finally (iv) thinly layered immature sandstones (Sovereto Member; Bonardi et al., 2009) (Figure 2-12).

The age of the Saraceno Fm. is Upper Eocene-lowermost Burdigalian (?) (Bonardi et al., 2009; Vitale & Ciarcia, 2013b). The thickness of the entire succession exceeds 1200 m (Bonardi et al., 1988a). The Parasicilide Unit (Figure 2-12) is characterized by four formations (Ciarcia et al., 2009a), from bottom to top: (i) clays and slates of the Postiglione Fm.; (ii) marls and limestones of Monte Sant’Arcangelo Fm.; (iii) whitish marls and marly limestones of Contursi Fm.; and (iv) foredeep deposits of the “Arenarie di Albanella” Fm. (Donzelli & Crescenti, 1962). The thickness of the whole succession exceeds 800 m. The age of these deposits ranges from the Middle Eocene to the Burdigalian, although the lower undated deposits could reach the Upper Cretaceous, in analogy with the Crete Nere Fm. (Bonardi et al., 1988a; Guerrera et al., 2005). The Sicilide Unit (Figure 2-12), analogously to the Parasicilide Unit, is divided in four formations (Guerrera et al., 2005), from bottom to top: (i) clays and slates of Argille Scagliose Fm; (ii) marls and limestones of Monte Sant’Arcangelo Fm.; (iii) clays and marls, somewhere calcarenites rich in foraminifera, of Argille Varicolori Fm; and (iv) foredeep deposits of the “Arenarie di Corleto” or “Tufiti di Tusa” Fms. (Figure 2-12). The thickness is about 1000 m (APAT, 2005). The age of these deposits ranges between the Upper Cretaceous (?) and the Burdigalian. The Frido Unit (Figure 2-12) is characterized by a metamorphic and highly deformed succession (Vitale et al., 2013b) consisting of oceanic crust (metagabbros, metadolerites and metapillows; “Timpa della Guardia” Fm.) and continental lithosphere

(serpentinized peridotites, metagranitoids, gneisses, amphibolites, granofels and metacarbonates; Timpa Rotalupo Fm.) covered by a deep basin metasedimentary succession formed by metaradiolarites, calcschists, phyllites, quartzites and metapelites (San Severino Fm.) and finally by Upper Oligocene calcschists (Monte Caramola Fm.; Vitale & Ciarcia, 2013b). The Frido Unit tectonically covers the Nord-Calabrese Unit in the SE sector (“Timpa delle Murgie” area), whereas it is placed below the Nord-Calabrese and Parasicilide Units in the NW sector (Seluci area). The Frido Unit experienced an HP/VLT metamorphism characterized by a pressure peak of ca. 1.2 GPa with temperature of ca. 350 °C (Vitale & Ciarcia, 2013b). The LAC is unconformably covered by a Langhian-lowermost Tortonian succession named Cilento Group (Amore et al., 1988; Russo et al., 1995), including arenitic and marly deposits of Pollica and San Mauro Fms (Vitale & Ciarcia, 2013b), and corresponding to the clastic undifferentiated succession of the Albidona Fm. in Lucania region (Selli, 1962). In the study area, further wedge-top basin successions, unconformable on the previous units, and characterized by dominantly coarse-grained clastic deposits, include the Upper Miocene Perosa (sensu Vezzani et al. (2010) and Oriolo (Selli, 1962) Fms. (Figure 2-12).

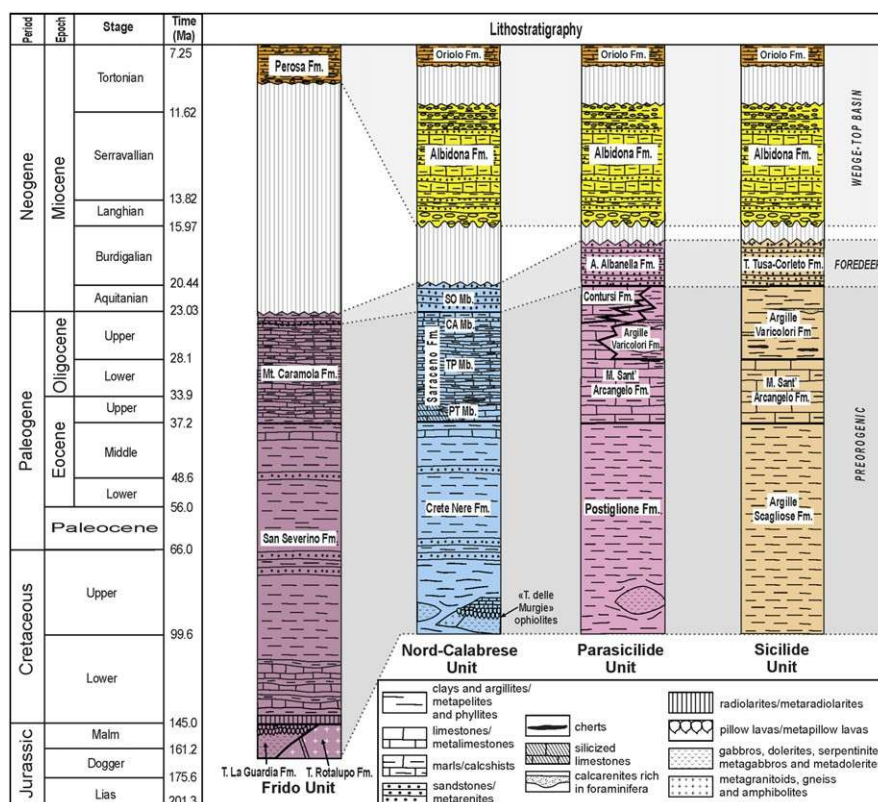


Figure 2-12: Schematic stratigraphic logs of LAC successions and Miocene wedge-top basin deposits. PT: Punta Telegrafo; TP: Terranova di Pollino; CA: Carpineta; SO: Sovereto (after Vitale et al. (2013b)).

2.6 The Cilento Group

The Cilento Group represents a large-scale, upward coarsening turbiditic megasequence (Figure 2-13). In the Mt. Stella area, its lower portion corresponding to the Pollica Formation records sedimentation within a deep-sea fan system (Critelli & Le Pera, 1990; Corrado et al., 2019).

The lower part of the Pollica Formation consists of over 300 meters of fine-grained sandy-shaly turbidites, forming the Cannicchio Member (Figure 2-13). This unit is notably intensely folded. In contrast, the upper part of the Pollica Formation is about 500 meters thick (ISPRA, 2005a, 2017b), generally undeformed or only mildly deformed, and dominated by coarse-grained sandstones, occasionally interbedded with conglomerates associated with channelized and interchannel turbidite facies (Figure 2-13).

Overlying the Pollica Fm. is the San Mauro Formation, with a thickness ranging from approximately 1,500 meters (Critelli & Le Pera, 1994) to over 2,000 meters (ISPRA, 2005a). Based on composite sections (Critelli & Le Pera, 1994), its lower segment (300–500 m) consists of thin-bedded sandstones intercalated with thick mudstones and marls, representing fan-fringe and basin-plain turbidites (Corrado et al., 2019). This interval also contains calciclastic turbidite beds and a coarse volcanoclastic layer.

The middle part of the San Mauro Fm. (~700 m) includes thick sandstone-lobe turbidites alongside thin fan-fringe facies, grading upward into channel-mouth and channelized sandstone and conglomerate turbidites. Notably, this section contains several intrabasinal carbonate turbidites, as well as two significant carbonates containing clastic megaturbidite beds (known locally as “Fogliarina”; (Ciarcia et al., 2011), measuring 65 m and 35 m thick, respectively (Critelli, 1999; Critelli et al., 2013). These megaturbidites exhibit impressive lateral continuity, extending over 300 km, from the Tyrrhenian to the Ionian coast (Colella & Zuffa, 1988).

The upper part of the San Mauro Formation is characterized by a 400–500 m thick succession of coarse-grained sandstone and conglomerate turbidites deposited in channel-mouth and channelized facies environments.

In the Mt. Sacro area, the Cilento Group includes the Torrente Bruca Formation (Amore et al., 1988), a >1500 m thick sequence composed of sandstone and conglomerate turbidites. This formation also includes siliciclastic olistostromes (20–100 m thick; (Cocco & Pescatore, 1968; Amore et al., 1988; Critelli, 1999) and carbonaticlastic megaturbidite beds similar to those in

the San Mauro Fm. Notably, these olistostromes contain large blocks ranging from outcrop to large size of meta-igneous rocks of inferred oceanic origin, including pillow basalts, granite, gneiss, and a condensed suite of bedded pelagic shale, limestone, chert, interpreted as disrupted fragments of deep-ocean sequences (Critelli & Le Pera, 1994; Critelli, 1999). Paleocurrent data show dominant sediment transport from the north-northwest and west for siliciclastic turbidites, with subordinate trends from the south-southwest. In contrast, calci-clastic turbidites reflect transport from the south-southeast (Ciarcia et al., 2011).

The Cilento Group is unconformably overlain by the Monte Sacro Conglomerate Formation (Tortonian) in the Cilento area (Critelli & Le Pera, 1990; Critelli & Le Pera, 1994), and by the Gorgoglione Formation in Basilicata. The Monte Sacro Conglomerate Fm. is a ~500 m thick channel-fill turbidite sequence composed of sandstones and conglomerates, overlying the Torrente Bruca Fm., with consistent south-southwest paleocurrent trends (De Pippo & Valente, 1991; Critelli & Le Pera, 1994).

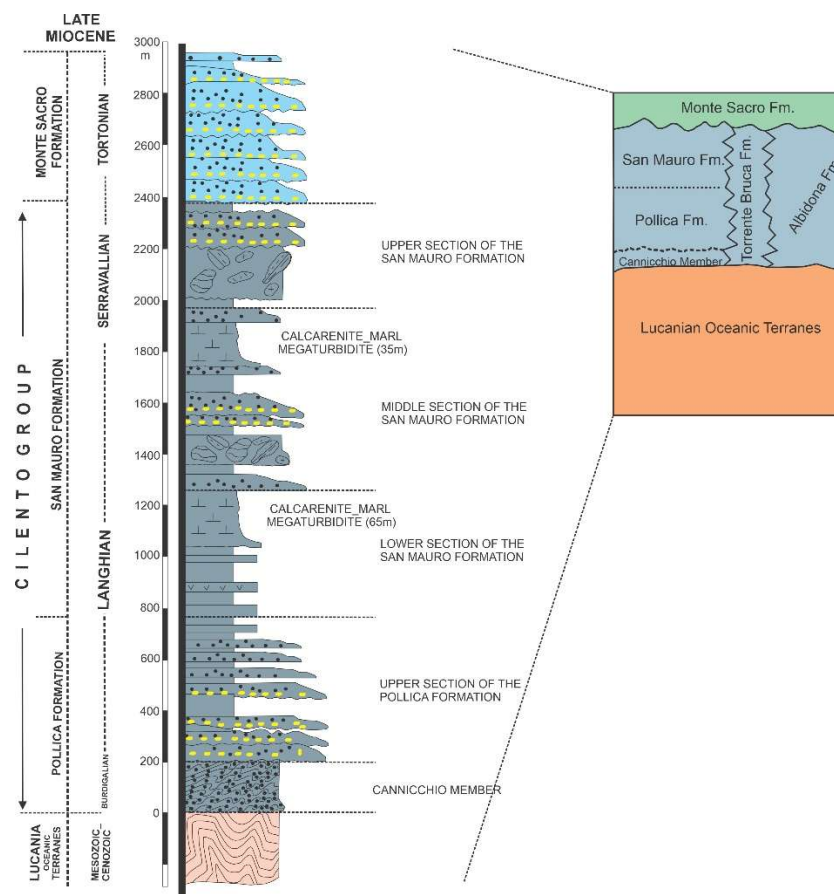


Figure 2-13: Schematic stratigraphic section of the Cilento Group, modified after Corrado et al. (2019).

2.7 Mt. Sirino

The Tempa Roccarossa area is located in the Basilicata region of southern Italy, specifically along the northeastern flank of Mt. Sirino, a prominent structural high in the southern Apennine fold-and-thrust belt (Figure 2-14). Mt. Sirino forms a large NE-verging anticline, whose internal architecture offers critical insights into the stratigraphic and tectonic evolution of the Southern Apennines. This anticline is primarily composed of units belonging to the Lagonegro Basin succession (often abbreviated as LMB), structurally underlain by elements of the Apenninic Platform (ApeP) and the Ligurian Accretionary Complex (LAC). These units are in contact along low-angle extensional detachments, reflecting post-thrusting extensional deformation associated with the late stages of orogenic collapse (Mazzoli et al., 2006). The Tempa Roccarossa area has been the subject of detailed geological mapping, most notably by the (SGI, 1969) and more recently by (ISPRA, 2014a). The structural configuration of the area is defined by the superposition of two major thrust sheets derived from the Lagonegro Basin. These thrust sheets represent distinct stratigraphic units: the lower Lagonegro unit (Lagonegro I) and the upper Lagonegro unit (Lagonegro II), as originally classified by (Scandone, 1967). Each of these units is characterized by its own distinct stratigraphic succession, reflecting variations in depositional environment and sediment supply during Mesozoic basin evolution.

2.7.1 Lagonegro I Unit (Lower LMB Unit)

The Lagonegro I Unit forms the structural core of the Mt. Sirino anticline and represents the deeper part of the basin-fill stratigraphy (Figure 2-14). It is predominantly composed of deep-marine sedimentary deposits accumulated in a pelagic basin setting during the Jurassic to Early Cretaceous (Vitale et al., 2025). The unit includes:

- ***Calcarei con Selce Formation***: This formation consists of Jurassic-age cherty limestones, representing hemipelagic to pelagic carbonate sedimentation. The presence of black chert nodules and intercalated radiolarian-rich layers indicates deposition in a relatively deep, low-energy marine basin with high biological productivity.
- ***Scisti Silicei Formation***: Overlying the *Calcarei con Selce*, this unit is made up of radiolarites and finely laminated silicified limestones interbedded with argillaceous sediments. These siliceous rocks are rich in radiolarians and reflect open-marine conditions with significant biogenic silica input, likely deposited under slow sedimentation rates in a distal basin environment (Scandone, 1967).

- ***Flysch Galestrino* Formation:** This formation represents a transition to more siliciclastic-dominated sedimentation. It comprises well-laminated silicified argillites and occasional fine-grained turbiditic sandstones. This unit signals a deepening or increased subsidence phase, possibly associated with the progressive closure of the Lagonegro Basin during convergence (Mazzoli et al., 2006).

These successions exhibit characteristics of a pelagic and hemipelagic depositional environment, consistent with long-term, deep-marine sedimentation in a slowly subsiding basin. The overall stratigraphy of the Lagonegro I Unit is consistent with its paleogeographic position within the internal domain of the southern Tethys during the Mesozoic.

2.7.2 Lagonegro II Unit (Upper LMB Unit)

The Lagonegro II Unit, which tectonically overlies the Lagonegro I Unit, includes older sedimentary sequences, mainly from the Triassic to Early Jurassic. At its base lies the Monte Facito Formation, one of the most significant and diagnostic units in this succession. This formation is notable for its complex lithological composition and is interpreted as a transition from shallow carbonate platform to deeper basin environments (Vitale et al., 2025).

- ***Monte Facito* Formation:** This unit includes a heterogeneous mix of Triassic recifal carbonates (recifal bodies) embedded within argillaceous, siliceous, and volcanoclastic sediments. The presence of radiolarites, shales, and sandstones interbedded with the carbonate buildups indicates a slope-to-basin setting adjacent to carbonate platforms. The *Monte Facito* Formation represents a tectonically dissected slope and basin system, possibly reflecting synsedimentary faulting and mass-wasting processes (Scandone, 1967; Mazzoli et al., 2006).

The inclusion of recifal carbonates suggests proximal influence from carbonate platforms such as the Apulian or Apennine domains. This stratigraphic unit is particularly important as it marks the beginning of the pelagic sedimentation that characterizes much of the Lagonegro succession.

The juxtaposition of the Lagonegro I and II units as thrust sheets in the Tempa Roccarossa area reflects the complex structural evolution of the Southern Apennines during the Neogene. These units were sequentially accreted to the orogenic wedge during the convergence of the African and Eurasian plates and were later affected by post-orogenic extension.

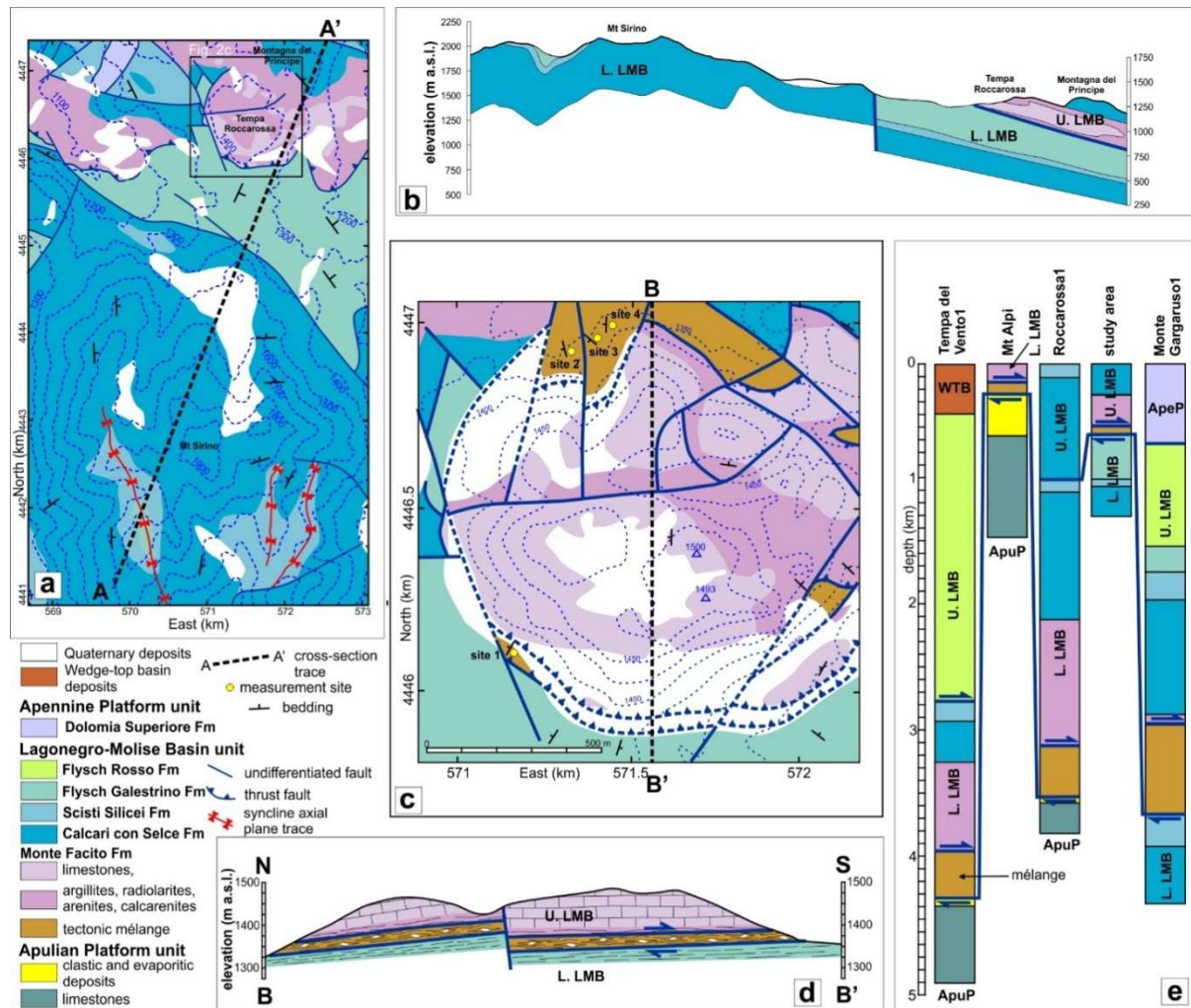


Figure 2-14: (a) Geological map of Mt Sirino (modified after ISPRA, 2014a). (b) Geological cross-section A-A'. (c) Geological map of Tempa Roccarossa (modified after Ciarapica & Passeri, 2000). (d) Geological cross-section B-B' (modified after ISPRA (2014a). (e) Schematic well logs (modified after ViDEPI-Project, 2025).

Chapter 3 Materials and Methods

This study adopts a multidisciplinary approach that integrates field-based structural analysis, petrography, geochemistry, thermobarometry, and advanced microstructural techniques to investigate the tectonic and metamorphic evolution of the Southern Apennines. The combination of these methods provides a comprehensive and robust framework for understanding the complex interactions between deformation, metamorphism, and sedimentation. Structural analyses, including the reconstruction of finite strain ellipsoids and kinematic vorticity, offer crucial insights into the deformation mechanisms and strain history of the study units. Petrographic and microstructural observations, particularly using SEM-EDS, EPMA-WDS, and Raman spectroscopy, allow for the detailed characterization of the mineralogical composition, fabric, and metamorphic conditions of the rocks, revealing how deformation and metamorphism have influenced the development of the *mélange* and LAC.

Thermobarometric analyses based on electron microprobe and multiequilibrium methods provide constraints on the pressure-temperature conditions during metamorphism, contributing to a better understanding of the burial and exhumation history of these tectonic units. Geochemical and X-ray diffraction (XRD) analyses complement these results by providing mineralogical and geochemical data that inform the interpretation of depositional and tectonic settings. A detailed description of these methods is in the following paragraphs.

3.1.1 Field Survey and Sampling

Field investigations were conducted across selected exposures of the Cilento Group in the Cilento area and the Tempa Roccarossa outcrop in the northeastern Mt. Sirino region. Stratigraphic sections were studied, and tectonic contacts were measured using standard geological mapping techniques. Systematic and oriented sampling was carried out along measured sections, targeting key lithological units, including arenaceous, pelitic, and calciclastic units, within the deformed sedimentary and metasedimentary rocks of the Tempa Roccarossa *mélange* and LAC. A total of 200 samples were collected, and selected samples were used for petrographic, geochemical, and microstructural analysis.

3.1.2 Structural Analysis

Detailed structural mapping was conducted in the Tempa Roccarossa and the LAC to investigate the geometry, internal architecture, and deformation features of the TR mélange and the LAC. The fieldwork focused on identifying and characterizing lithological boundaries, tectonic contacts, and mesoscopic structures.

To obtain information about the strain in 3D, we reconstructed the FSEs for four samples (ML1, 2, 3, 6) from the TR melange and 16 samples from the LAC. Using the R_f/ϕ method (Figure 3-2; (Ramsay, 1967; Dunnet, 1969)), strain markers were analyzed with the free software EllipseFit (Vollmer, 2018). The technique uses strain markers that can be approximated as ellipses by measuring their ellipticity (R_f) and the angle ϕ between the long axis of the ellipse and a reference line (Figure 3-2). We considered only the objects deformed by ductile strain for the finite strain estimation, that is, the calcareous clasts, and avoided the rigid porphyroclasts, that is, the quartz clasts, on the contrary characterized by brittle deformation, which would furnish an underestimation of the finite strain (Vitale et al., 2020). Another assumption is that the long axes of the original objects were randomly oriented before the ductile deformation (Dunnet, 1969). The main limitations of this method are the preparation of the slabs and thin sections, especially the choice of the XZ and YZ planes. In addition, another uncertainty arises when a preferred orientation is recorded before ductile deformation. Finally, the FSEs were classified using the Flinn-Ramsay plot (Flinn, 1962; Ramsay, 1986).

3.1.3 Structural measurements

Structural measurements were carried out in both the TR melange and the LAC units. We analysed planar and linear fabrics, such as foliations and stretching lineations, which are essential features in the study of geological structures and in understanding their tectonic history. The structural analysis included the study of tectonic foliations, fold axes, axial planes, and fold geometries to understand the deformational characteristics of folds, thereby providing insights into the tectonic stresses responsible for their formation. Shear sense indicators, such as S–C and S–C' structures, asymmetrical pressure shadows, and rotated clasts, were also analysed, as they play a key role in interpreting the sense of shear or the direction of movement in shear zones. Fault plane orientations and kinematic indicators, such as slickensides, were used to determine the motion along fault lines.

A comprehensive kinematic analysis was conducted to determine the dominant shear sense and the deformation mechanisms responsible for fabric development in the tectonic *mélange*. Shear sense determinations were made in both outcrop and thin-section scale using petrographic analysis under plane- and cross-polarized light.

3.1.4 Kinematic Vorticity Analysis

To determine the degree of coaxial strain in the TR melange and the LAC units, we estimated the kinematic vorticity W_k using two methods. The dimensionless number W_k indicates the partitioning between a coaxial and non-coaxial deformation (Means et al., 1980), varying between 0 (pure shear) and 1 (simple shear). Intermediate values indicate the coexistence of pure and simple shear (general shear, Means & Park, 1994; Tikoff & Fossen, 1999); or sub-simple shear, (Fossen & Cavalcante, 2017). The first of the methods employed for W_k determination (Figure 3-1) is based on the evidence that in a general shear, rigid objects can rotate forward or backward with respect to the sense of shear (Wallis et al., 1993; Jessup et al., 2007). However, all porphyroclasts are further partitioned into two populations: one rotates continuously without developing a preferred orientation, and the other aligns along a minimum energy position (Jessup et al., 2007). We calculated the vorticity number W_k using the Rigid Grain Net (RGN), which comprises a series of semi-hyperbolas that enable a simple estimate of this kinematic parameter. The method (Figure 3-1) uses the estimation of the angle ξ that the long axis forms with the mylonitic foliation and the factor B^* attained by the following equation:

$$B^* = \frac{(a^2 - b^2)}{(a^2 + b^2)}.$$

Where a and b are the long and short axes of the rigid object, respectively.

The second method for evaluating the W_k uses synthetic (R2) and antithetic (R1) S-C' structures (Platt & Vissers, 1980; Behrmann, 1987; Grasemann & Stüwe, 2001), which are very common features in general shear zones (Xypolias, 2010). This method (Kurz & Northrup, 2008; Gillam et al., 2013), is based on the orientation of R1 and R2 planes with respect to the shear zone boundary. Generally, R2 shear planes are oriented parallel to the acute bisector of the flow apophyses (A1 and A2, Figure 3-2) or at a lesser angle (Kurz & Northrup, 2008), whereas R1 shear planes are poorly developed and have a mean inclination parallel to the obtuse bisector of the flow apophyses. The vorticity number is calculated by the equation (Kurz & Northrup,

2008), $W_k = \cos(2\theta)$, where θ is the angle between R2 and the shear plane assumed to be parallel to the flow apophysis A2 (Kurz & Northrup, 2008; Gillam et al., 2013), hence parallel to the mylonitic foliation in highly deformed shear zones.

The two methods are independent approaches that use two different types of structures. Both depend on the amount of simple and pure shear and can, therefore, provide independent estimates. However, both methods have some limitations, including the three-dimensional problem, that is, we have studied the motion of rigid clasts in two-dimensional sections that are only projections in the plane of the thin section (e.g., Iacopini et al., 2011; Montemagni et al., 2020). As in the sample preparation for the strain analysis, the sample cuts may not be perfectly parallel to the XZ plane of the FSE and, hence, result in underestimations of the dip angles of the R1 and R2 planes and of the contribution of the pure shear. Furthermore, for the S-C' method, some R1 and R2 planes may be influenced by rigid porphyroclasts, modifying the dip values. To overcome the latter limitation and reduce uncertainty, we calculated the mean dip angles using a frequency distribution that is approximately Gaussian. For individual W_k values, we adopted an uncertainty of ± 0.1 , as indicated by (Tikoff & Fossen, 1999), which are values greater than 0.8 (Iacopini et al., 2011). In any case, using both methods allows us to identify more realistic value intervals.

We analysed the thin sections of samples ML1, 2, 3, and 6 and the polished slab of ML1 by the RGN method, obtaining 174, 202, 136, and 165 data for samples ML1, 2, 3, and 6, respectively. The R1-R2 method was applied to the same samples, collecting 282, 276, 348, and 218 data for samples ML1, 2, 3, and 6, respectively. We applied a similar method on mesoscopic scale on the LAC, we analysed the F1, F2, F3, F4, and F5 and the total number measured is 138, 214, 153, 313, and 54, respectively and microscopic level (SEM based), the samples were labelled as P41, P91, P21, P11, and P01. The total number measured for SEM based analysis are 147, 92, 68, 106 and 110 respectively.

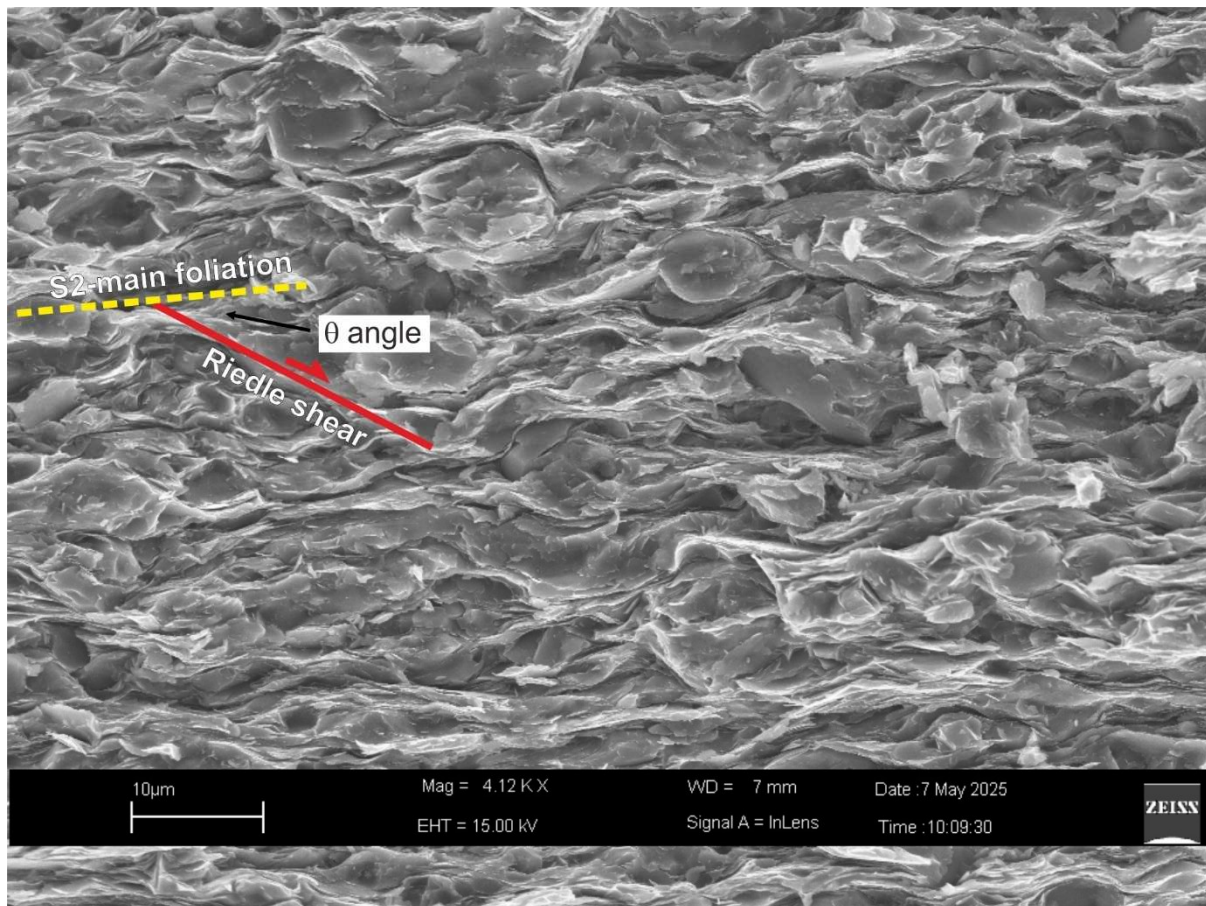


Figure 3-1: Example of vorticity analysis measurements carried out on microscale in LAC.

3.1.5 3D Finite Strain Analysis

To obtain information about the strain in 3D, we reconstructed the FSEs for oriented four samples (ML1, 2, 3, 6) for the *mélange* samples and 16 samples from the LAC units. Using the Rf/ϕ method (Figure 3-2 a, b; (Ramsay, 1967; Dunnet, 1969), we analyzed the strain markers with the free software EllipseFit ((Vollmer, 2018). The technique uses strain markers that can be approximated to ellipses by measuring their ellipticity (Rf) and the angle ϕ that the long axis of the ellipse forms with a reference line (Figure 3-2 a, b). We considered only the objects deformed by ductile strain for the finite strain estimation, that is, the calcareous clasts, and avoided the rigid porphyroclasts, that is, the quartz clasts, on the contrary characterized by brittle deformation, which would furnish an underestimation of the finite strain (Vitale et al., 2020; Vitale et al., 2025). Another assumption is that the long axes of the original objects were randomly oriented before the ductile deformation (Dunnet, 1969). The main limitations of this method are the preparation of the slabs and thin sections, especially the choice of the XZ and YZ planes. In addition, another uncertainty arises when a preferred orientation is recorded

before ductile deformation. Finally, the FSEs were classified using the Flinn-Ramsay plot (Flinn, 1962; Ramsay & Huber, 1983).

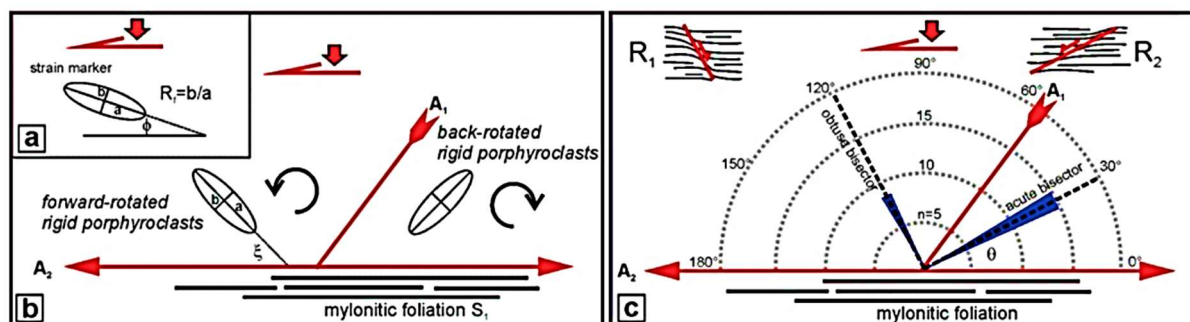


Figure 3-2: (a) R_f/ϕ method for strain analysis. Vorticity W_k estimation: (b) rigid porphyroclast method; (c) synthetic and antithetic S-C' structures (R1-R2) method.

3.1.6 Petrography and Microstructural Observations

We analysed a total of 110 samples for the TR melange and the LAC. The goal of this petrographic study was to characterize the fabric, foliation and microstructure of the samples, to understand the geological processes, deformation, veining events and tectonic activity that shaped these rocks. Samples for petrographic microstructural analysis were collected from within the Ligurian Accretionary Complex, including three different locations (Punta Telegrafo, Pioppi and Caprioli) and from the melange in Tempa Roccarossa. Each sample was chosen to provide a comprehensive view of the geological setting and to capture the variability within the complex and melange. The preparation of rock samples for petrographic analysis is a critical step to ensure accurate mineral identification and analysis. The following steps were employed:

Collected rock samples were first cut through a rock cutting machine in the petrographic laboratory in DISTAR, University of Naples Federico II. A mechanical rock cutter was used to break them into smaller, manageable pieces. After cutting, the smaller rock samples were then processed for the preparation of thin sections, which allows the samples to be used for various forms of analysis, including thin section preparation (Figure 3-3 a, b). Small portions of the ground rock samples were selected for thin section preparation. These samples were mounted on glass slides and polished to a thickness of approximately 30 microns (Figure 3-3 c). The thin section preparation process involves precise cutting and polishing to obtain a smooth, clear surface for mineralogical analysis under a microscope. Once the thin sections were prepared, they were mounted on glass slides using an epoxydic resin to ensure they remain fixed and

stable during analysis. The resin also prevents air bubbles and cracks from forming that might interfere with microscopic observation.

3.1.7 Identification of microstructural features

The prepared thin sections were studied in detail for microstructural analysis. Recrystallised or new-formed minerals or alteration products were observed. These minerals, which include white mica and chlorite, can reveal insights into the deformation or metamorphic conditions that the rock underwent after its formation. Thin sections of representative samples were analysed under polarising and reflected light microscopy to characterise mineralogical composition, grain fabric, and microstructures. Special emphasis was placed on identifying deformation-related recrystallisation, different vein stages, and foliation development. Original fabric alteration was noted to interpret the intensity and nature of anchimetamorphism in the *mélange* units and the LAC samples.

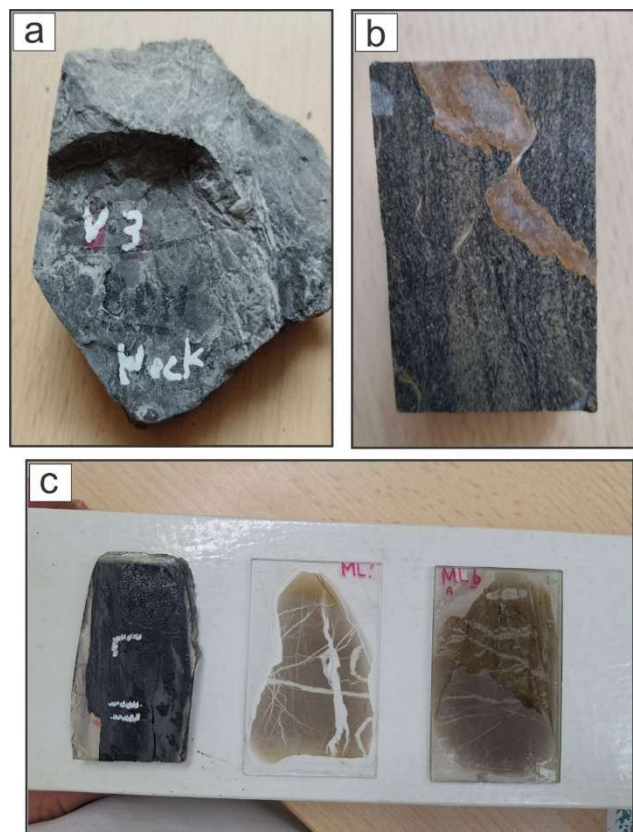


Figure 3-3: The photos show stages from the rock sample to the cut thin slab and finally a thin section of 30 microns.

3.1.8 Scan electron microscopy analysis (SEM-EDS)

3.1.8.1 Tempa Roccarossa

The SEM-EDS microanalyses were carried out at the DiSTAR using an Oxford Instruments Microanalysis Unit equipped on a Zeiss Merlin VP Compact Field Emission SEM (Figure 3-4). The technique has been shown to provide reliable microanalytical data for major oxide concentrations (see Guyett et al. (2024) and references therein). Measurements were performed using a 15-kV primary beam voltage, 60 μ A filament current, variable spot sizes and a 10-s of acquisition time. Corrections for matrix effects were performed using the XPP correction routine based on a Phi-RoZeta approach. Standards for calibration were: anorthoclase (Na, Al, Si); microcline (K); serandite (Mn); diopside (Ca); fayalite (Fe); olivine (Mg); rutile (Ti); apatite (P); fluorite (F); barite (Ba); celestine (Sr); zircon (Zr and Hf); synthetic Smithsonian orthophosphates (La, Ce, Nd, Sm, Y and Sc); pure niobium (Nb); pure vanadium (V); pure cobalt (Co); pure nickel (Ni); chromite (Cr); Corning glass (Th and U); sphalerite (Zn); pyrite (S); sodium chloride (Cl); pollucite (Cs). Accuracy and precision were tested using the rhyolitic Lipari obsidian ID3506 and basaltic Laki 1783 tephra from Kuehn et al. (2011) as secondary standards. Relative analytical uncertainty was <5% for SiO₂, Al₂O₃, K₂O, CaO, and FeO, ~10% for the other elements.



Figure 3-3: Scan electron equipment's DiSTAR, University of Naples Federico II.

3.1.8.2 Ligurian Accretionary Complex

The microstructures were studied using Scanning Electron Microscopy (SEM/EDS) using the SEM lab at DiSTAR (description section 3.1.8.1) and SEM lab of IMAA-CNR. The analyses were performed at the Scanning Electron Microscopy Laboratory of IMAA-CNR in Tito Scalo (Figure 3-5, equipped with a Zeiss Supra 40 field emission scanning electron microscopes (FE-SEM) with an Oxford INCA Energy 350 energy-dispersive X-ray (EDS) system and an INCA X-act SDD detector. The SEM observations focused on small representative portions of the samples cut perpendicular to the stratification plane and along the lineation. For analysis, the samples were first fixed onto an aluminum stub (12 mm in diameter) using graphite double-sided tape and then coated with carbon to make the surface conductive. The samples were examined using secondary electron (SE) images. The images were acquired at various magnifications with an acceleration voltage ranging from 3 to 20 kV and apertures of 20 μm and 60 μm . The EDS spectra were acquired with an acceleration voltage of 20 kV using a 30 μm aperture. The secondary electron images provided important information about the rock's microstructure. We analysed 16 samples in the SEM lab at DiSTAR for 3d strain analysis.



Figure 3-4: SEM equipment at IMAA-CNR in Tito Scalo.

3.2 Mineralogical and Geochemical Analysis

3.2.1 XRPD (X-Ray Powder Diffraction) analysis (TR Mélange)

Mineral identification and quantification were performed using XRPD. Bulk rock powders and $<2\ \mu\text{m}$ clay fractions were analyzed using a Philips X'Pert Pro diffractometer. Oriented clay mounts underwent air-drying, glycolation, and heating treatments. Quantitative phase analysis was performed using Winfit software, and semi-quantitative analysis of clay minerals followed established protocols. Paleotemperatures were estimated using the Kübler Index (KI), with reference calibration standards, and interpreted using empirical diagrams linking KI values to temperature ranges.

XRPD analysis was performed to identify and quantify mineral phases. Samples were dried at room temperature, gently crushed, homogenized, and split into two aliquots. One portion was grinded for both XRPD and XRF analyses, while the other was used for wet extraction of clay fractions ($<2\ \mu\text{m}$) following Stock's law (Moore & Reynolds, 1997). A Philips X'Pert Pro instrument, with Bragg Brentano geometry, equipped with a Co anode (setting 40 kV and 40 mA) and X'Celerator detector housed at the *Dipartimento di Geoscienze* of the *Università di Padova* was used. The data were collected using a spinner sample holder, with a 2θ angle

ranging from 3° to 85°, for both random powder bulk samples and oriented specimens of the clay fraction. The random powder of the bulk sample was prepared using the side-loading method (Środoń, 1999). Oriented specimens of the clay fraction were prepared by settling a suspension after being saturated with Mg⁺⁺ cations using 1 N MgCl₂ solution. Each specimen was analyzed, air-dried, glycolated at 60°C for 8 hr and heated at 375°C for 1 hr (Moore & Reynolds, 1997). Quantitative analysis was carried out on bulk sample XRD profiles measuring the peak area of both phyllosilicates and non-phyllosilicates (Cavalcante et al., 2007) using the Winfit software (Krumm, 1999) for the decomposition of peaks, the determination of integrated intensity, and the measurement of peak height and Full Width at Half Maximum (FWHM). Results were compared with whole-rock major oxides data using the experimental software vbAffina (Leoni et al., 2008; Cesarano et al., 2018). A semi-quantitative analysis of clay minerals amount in clay fraction was performed on glycolated and heated XRD patterns (Cavalcante et al., 2007). The paleotemperatures were determined by the Kübler Index (KI) using the international standards proposed by (Warr & Rice, 1994), recalibrated with KI values provided by M. Frey and W.B. Stern (University of Basel) (Warr & Mählmann, 2015). In particular, the temperatures were estimated using the diagram (KI vs. T) reported by (Cavalcante et al., 2012). This diagram was reconstructed considering the KI value and temperature from the literature (see (Cavalcante et al., 2012) and references therein). The average error percentage for FWHM measurements is about 6% (Warr, 2018), corresponding to a temperature range of ±10°C.

3.2.2 XRPD (X-Ray Powder Diffraction) analysis (LAC)

To determine the thermal maturity degree and perform microstructural analyses of the Crete Nere Formation of the North Calabria Unit, 14 samples of claystones and phyllites were collected from three different locations. In detail, 5 samples were collected from the Pioppi locality (PP1, PP2, PP3, PP4, PP5), 6 samples from Punta Telegrafo (PT1, PT2, PT3, PT4, PT5, PT6) and 3 samples from Caprioli (CPP1, CPP2, CPP3). The collected samples were analyzed using X-ray diffraction method.

3.2.3 Sample preparation for XRD analysis

X-ray diffraction (XRD) is widely used for the mineralogical analysis of sediments, allowing both qualitative identification and quantitative analysis of powdered samples. Mineralogical

analyses were conducted at the Institute of Methodologies for Environmental Analysis (IMAA) in Tito Scalo (PZ). Before irradiation, the samples underwent several preparatory steps, including drying, crushing, quartering, pulverization, cation saturation, glycolation, and heating. Initially, samples were dried under a hood at 30-40°C for several days (Cavalcante & Belviso, 2005), then disaggregated using a porcelain mortar and pestle. The quartering process was performed to obtain a representative sample for mineralogical composition analysis and an aliquot for clay fraction extraction ($<2\ \mu\text{m}$) (Cavalcante & Belviso, 2005), (Figure 3-6).

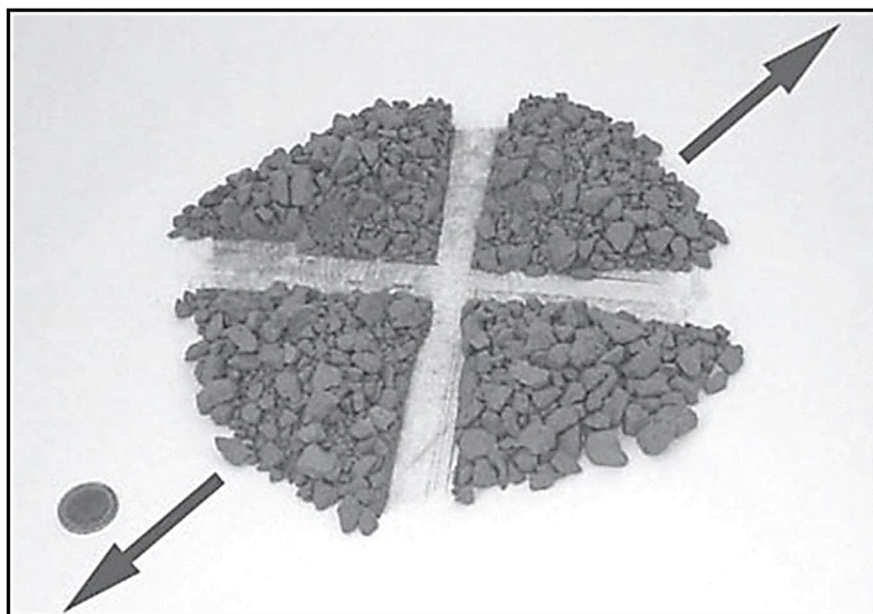


Figure 3-5: Quartering operation: opposite sectors are removed and subjected to further quartering (from Cavalcante and Belviso (2005)).

The first portion was pulverized using a mechanical pulverizer with an agate concentric disc jar and a hermetically sealed lid (Figure 3-7 a, b).

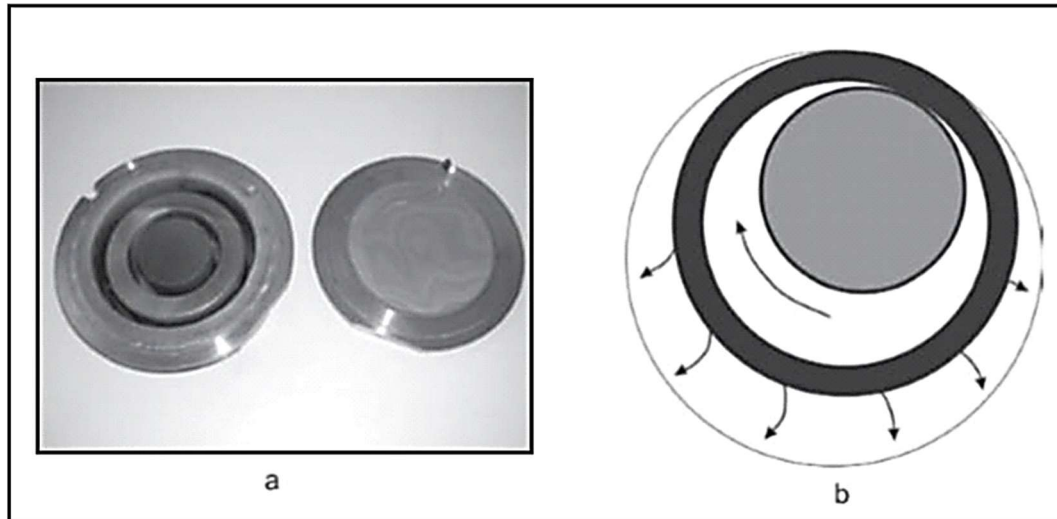


Figure 3-6: a) Agate concentric disc jar with lid and hermetically sealed closure. b) Representation of the pulverization mechanism inside a concentric disc jar (from Cavalcante and Belviso (2005)).

Thirty grams of the sample were pulverized for 7 minutes to ensure good random orientation and compaction using the lateral loading method (Cavalcante & Belviso, 2005). The clay fraction ($<2 \mu\text{m}$) was extracted via fractional sedimentation in a beaker. An aliquot of approximately 10 grams of sample was agitated mechanically and left to settle from a height of 10 cm, following Stokes' law (Cavalcante & Belviso, 2005). The sedimentation process relies on the settling velocity of a body in a fluid under gravity, as described by Stokes' law.

$$V = g(\rho_p - \rho_l) D^2 / 18\eta$$

The velocity V can also be expressed as the ratio between the height h and the time t given by:

$$t = 18\eta h / g(\rho_p - \rho_l) D^2$$

Where:

- t is the time it takes for a particle of a given diameter to settle at the bottom of the suspension of height h .
- g is the acceleration due to gravity.
- $\rho_p - \rho_l$ represents the difference in density between the particle and the liquid.
- D is the diameter of the particle in cm.
- η is the viscosity of the liquid at a given temperature **Error! Reference source not found..**

Table 3-1: Sedimentation times for particles of different diameters calculated using Stokes' law (from Cavalcante and Belviso (2005)).

Temp.	Granulometric Fractions					
T°C	φ63mm	φ32mm	φ16mm	φ8mm	φ4mm	φ2mm
30	23"	1'32"	6'06"	24'25"	1h 38'	6h 30'
29	23"	1'34"	6'16"	24'57"	1h 39'	6h 39'
28	24"	1'35"	6'28"	25'31"	1h 42'	6h 48'
27	24"	1'38"	6'31"	26'05"	1h 44'	6h 57'
26	25"	1'40"	6'40"	26'41"	1h 47'	7h 07'
25	25"	1'43"	6'50"	27'28"	1h 49'	7h 17'
24	26"	1'45"	6'59"	27'57"	1h 52'	7h 28'
23	27"	1'47"	7'10"	28'37"	1h 54'	7h 38'
22	28"	1'50"	7'20"	29'19"	1h 57'	7h 49'
21	28"	1'53"	7'31"	30'01"	2h 00'	8h 01'
20	29"	1'55"	7'41"	30'46"	2h 03'	8h 13'
19	29"	1'58"	7'53"	31'33"	2h 06'	8h 25'
18	30"	2'01"	8'05"	32'20"	2h 09'	8h 38'
17	31"	2'04"	8'17"	33'10"	2h 12'	8h 51'
16	32"	2'08"	8'31"	34'02"	2h 16'	9h 05'
15	33"	2'11"	8'44"	34'55"	2h 19'	9h 19'
14	33"	2'14"	8'58"	35'50"	2h 23'	9h 34'

The clay fraction was saturated with a 1N magnesium chloride (MgCl_2) solution (MgCl_2) solution to form a stable complex of water molecules in the interlayer region of vermiculite and smectite minerals, resulting in a basal spacing of approximately 1.4 nm (Brown & Brindley, 1980). This allows differentiation between expandable clay minerals and non-expandable ones (micas, illite), which retain a 1 nm spacing after saturation. Excess chlorides were removed by washing the sample with a distilled water and alcohol solution (1:1) (Cavalcante & Belviso, 2005). The washing was repeated until chloride ions were eliminated, as confirmed by adding silver nitrate to the suspension. Oriented powder preparations of the $<2 \mu\text{m}$ fraction were made using the sedimentation method, depositing 1 ml of the suspension on a glass support at a concentration of 4 mg/cm^2 (Cavalcante & Belviso, 2005) (Figure 3-8).

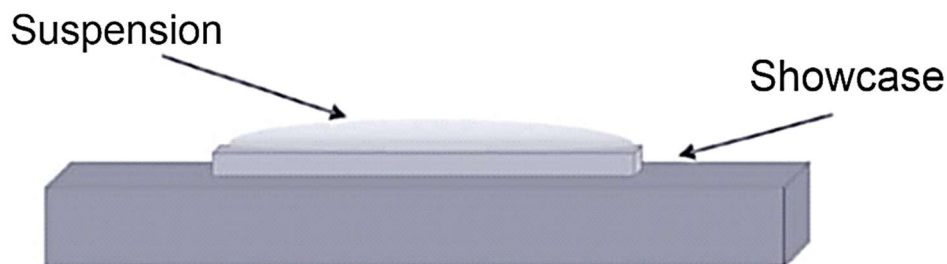


Figure 3-7: Preparation of oriented powder by sedimentation from a suspension (from Cavalcante and Belviso (2005)). To better characterize the clay phases present, the samples were subjected to glycolation using ethylene glycol (Bradley, 1945), and heating at 375 °C (Moore & Reynolds, 1997). The glycolation process was carried out by exposing the sample to the vapors of the reagent for 8 hours at 60 °C in a thermostat-controlled oven. The heating at 375 °C was performed in a muffle furnace for 2 hours (Cavalcante & Belviso, 2005).

3.2.4 Mineralogical Analyses by X-ray Diffraction

The diffraction analyses were carried out at the Department of Geosciences at the University of Padua using a Philips X'Pert Pro diffractometer, with Bragg-Brentano geometry, equipped with a Co anode (40 kV voltage and 40 mA current), and the X'Celerator detector. The data were collected with a sample holder equipped with a spinner over an angular scanning range of 3 to 85° 2θ for both randomly oriented powders of the total sample and the oriented preparations of the clay fraction. For the determination of phases in the total sample, reflections from the "data base" of the Joint Committee on Powder Diffraction Standards (JCPDS) were used.

The analysis of clay minerals was performed on the oriented preparations of the fraction dried in air, glycolated for 8 hours at a temperature of 60 °C, and heated for 1 hour at 375 °C (Moore & Reynolds, 1997). For the quantitative determination of minerals identified in both the total sample and the <2 μm fraction, methodologies from the literature were used (Cavalcante et al., 2007), and references therein). The areas of the most intense reflections (except for quartz, where the second most intense reflection was used) of individual minerals were then measured using the Winfit program (Krumm, 1996). The intensities thus obtained were normalized by multiplying them by the corresponding correction factors and then related to those of quartz:

$$P_i = A_i \cdot f_i / A_{\text{quartz}} \cdot f_{\text{quartz}}$$

Where P_i , A_i , and f_i are respectively the weight, area, and correction factor of the i -th phase, and A_{quartz} and f_{quartz} are the area and correction factor for quartz. The correction factors were defined based on the analytical methodologies and standardization procedures used in the

IMAA laboratories. Once the weights of the various phases were known, they were normalized to 100 as a percentage of abundance using the following formula:

$$\% \text{ Mineral} = P_i / \sum P_i n_i$$

Since quartz is the reference mineralogical phase, it was assigned a weight of 1. The results were refined with chemical data obtained through X-ray fluorescence of the total rock using the experimental software vbAffina (Leoni et al., 2008; Cesarano et al., 2018). The semi-quantitative analysis of the clay minerals present in the <2-micron fraction was performed on the glycolated and heated diffractograms (Cavalcante et al., 2007). Paleotemperatures were determined using the Kübler index (KI) with international standards proposed by (Warr & Rice, 1994), recalibrated to the KI values provided by M. Frey and W.B. Stern (University of Basel) (Warr & Mählmann, 2015). Specifically, temperature estimates were determined using the (KI vs T) diagram reported in (Cavalcante et al., 2012). This diagram was reconstructed considering the KI and temperature values reported in the literature (see Cavalcante et al. (2012), and references therein). The average error percentage for the FWHM measurements is about 6% (Warr, 2018), which corresponds to a temperature range of ± 10 °C. The characteristics of illite/smectite (I/S) and chlorite/smectite (C/S) interstratifications were determined following the guidelines in (Moore & Reynolds, 1997).

3.2.5 XRF (X-Ray Fluorescence) Analysis

Major and trace element compositions analysis were carried out for the TR melange, and information was obtained from powdered samples using a Panalytical Axios spectrometer. Pressed powder pellets were analyzed, and weight loss on ignition (LOI) was measured gravimetrically. Analytical uncertainties were approximately 1–2% for major oxides and 5–10% for trace elements.

For XRF analyses, powdered rock samples were used to prepare pressed powder pellets, which were analyzed for major oxides and trace elements using a Panalytical Axios instrument housed at the DiSTAR. Analytical uncertainties are in the order of 1%–2% for major oxides and 5%–10% for trace elements. Weight loss on ignition (LOI) was determined gravimetrically after heating 1 g of rock powder (pre-dried at ~ 150 °C overnight) at 900 °C for three hours.

3.3 Thermobarometric Constraints

3.4 Electron microprobe analysis- TR Melange

We carried out a thermobaric analysis to estimate metamorphic conditions, mineral assemblages in metasedimentary samples from Tempa Roccarossa were analysed using electron microprobe and scanning electron microscopy. Phase equilibrium modeling was supported by X-ray diffraction and chlorite geothermometry, following the methodology of (Lanari et al., 2014), as adopted by (Vitale et al., 2025). Pressure-temperature (P–T) conditions were reconstructed to evaluate the burial and exhumation history of the *mélange* within the Upper Miocene–Pliocene orogenic wedge and subsequently for the LAC.

A multi-equilibrium approach was used to determine the P-T conditions for the formation of chlorite and white mica. The chlorite + quartz + H₂O method of Vidal et al.(2005, 2006) was used to estimate the formation temperature of chlorite. If the maximum temperatures represent the conditions of peak metamorphism (e.g., Lanari et al., 2012; Petroccia et al., 2025) , the pressure conditions of white mica formation were determined using the mica + quartz + H₂O method of (Dubacq et al., 2010). Finally, a P-T estimate was obtained using the chlorite + mica + quartz + H₂O method of (Vidal et al., 2006). Chlorite and white mica analyses have been selected where the equilibria show good convergence in the P-T space, but this is not sufficient to prove chemical equilibrium; it only shows that the mineral composition is consistent with chemical equilibrium being achieved at these P-T conditions (Lanari & Duesterhoeft, 2019). A full description of each method and the limitations of this approach compared to others can be found in (Lanari & Duesterhoeft, 2019; Petroccia et al., 2025). Thermodynamic calculations were performed using the ChlMicaEqui 1.5 program (Lanari et al., 2012; Lanari & Duesterhoeft, 2019). The following solution models were used: (Vidal et al., 2005; Vidal et al., 2006) for chlorite (chlorite + quartz + H₂O and chlorite + mica + quartz + H₂O methods), (Parra et al., 2002) for white mica (chlorite + mica + quartz + H₂O method) and (Dubacq et al., 2010) for the determination of a P-T line along which mica is in equilibrium with quartz and H₂O (mica + quartz + H₂O method). All solution models are compatible with the equations of state and the internally consistent data set of Berman, 1988. The activity of H₂O in the fluid was set to 1.

3.4.1 Electron microprobe analysis (EPMA)- LAC

We carried out a similar analysis on the LAC samples. The X-ray maps and quantitative spot analyses were obtained with the Electron Probe Microanalyzer (EPMA) JEOL 8200 Super Probe equipped with five wavelength-dispersive spectrometers (WDS) at the High-pressure and temperature lab, National Institute of Geophysics and Volcanology of Rome (Figure 3-9). We selected 3 samples from Caprioli, Punta Telegrafo and Pioppi, respectively. Point analyses were acquired before X-ray compositional maps were acquired on the same thin section. Analytical conditions of point analysis were a 15 keV accelerating voltage, a 5 nA beam current and a beam $\varnothing$ of $\sim 1 \mu\text{m}$. Analytical conditions for X-ray map acquisition were a 15 keV accelerating voltage, a 100 nA specimen current, and a 40 ms dwell time. Ten elements (Si, Ti, Al, Fe, Mn, Mg, Na, Ca, K and S) have been acquired with two cycles of WDS-detector mapping. X-ray maps were processed using XMapTools 4.3 (Lanari et al., 2012), and intensity X-ray maps were standardized to concentration maps of oxide weight percentage using spot analyses as internal standards. X-ray calibrated maps were used to extract the local bulk composition necessary as input for thermodynamic modelling to constrain the metamorphic conditions of different fabrics, following the procedure described in (Lanari & Engi, 2017) adopted by Petroccia et al., 2025; Vitale et al., 2025.

The standardization process for WDS spot analyses was evaluated by comparing the results against reference materials with known concentrations or using stoichiometric constraints for unknown mineral phases. For the X-ray map standardization, only high-quality WDS spot analyses were used to ensure accuracy. The software XMapTools 4 (Lanari et al., 2012) was employed for both the classification and analytical standardization, which allowed the calculation of mineral structural formulas and the generation of cation abundance maps.

The chemical resolution of individual pixels within the X-ray maps is influenced by key analytical parameters such as dwell time, accelerating voltage, and beam current. As shown by Petroccia et al. (2025), the uncertainty and detection limits improve when several pixels are included in the analysis compared to relying on single-pixel statistics. By averaging over multiple pixels from either unzoned material or a single growth zone, the total counting time increases, thereby reducing the relative uncertainty and enhancing the reliability of the data.



Figure 3-8: Electron microprobe equipment at INGV, Rome.

3.5 Raman spectroscopy

Raman spectra were obtained using a Renishaw (Wotton-under-Edge, UK) In VIA Reflex microspectrometer (ENS Paris) equipped with a 514-nm Spectra Physics (20 mW) argon laser (Figure 3-11). The laser was focused on the sample using a DMLM Leica (Wetzlar, Germany) microscope with a 100 $\times$ objective (NA = 0.90), and the laser power at the sample surface was set at around 1 mW. The signal was filtered by edge filters and finally dispersed using an 1800 g mm⁻¹ grating to be analysed by a Peltier cooled REN-CAM CCD detector. Before each session, the spectrometer was calibrated with a silicon standard. As RSCM can be affected by several analytical pitfalls, we followed the analytical procedure described by Beyssac et al. (2002, 2003, 2007) closely. To check the within-sample structural heterogeneity, 10–15 spectra were recorded for each sample. We used the extended scanning mode, which allows the continuous recording of large spectral windows from 700 to 2000 cm⁻¹. Such a spectral range is the minimum required to image the whole background signal clearly and to define the baseline properly. There are several available fitting procedures in the literature for such disordered CM, such as four-band fitting using a Voigt profile (Beyssac et al., 2002, 2003; Rahl et al., 2005), two-band fitting using a combination of asymmetric Breit–Wigner–Fano and

Lorentzian profiles (Bonal et al., 2006) and a five-band combination of Lorentzian/Gaussian profiles (Sadezky et al., 2005).

Using a Voigt profile for five bands, the system does not converge towards a unique solution, for instance, in balancing the respective contributions of D1 and D4 to the D1 composite band, or simply to fit the D3 band. Using a Lorentzian profile for all bands reduces the degree of freedom of the system compared with the Voigt alternative, and a five-band solution with Lorentzian shape for all bands systematically converges rapidly towards a unique solution. This fitting procedure is represented in Figure 3-11 and is relatively similar to that given for the study of carbonaceous soots by (Sadezky et al., 2005).

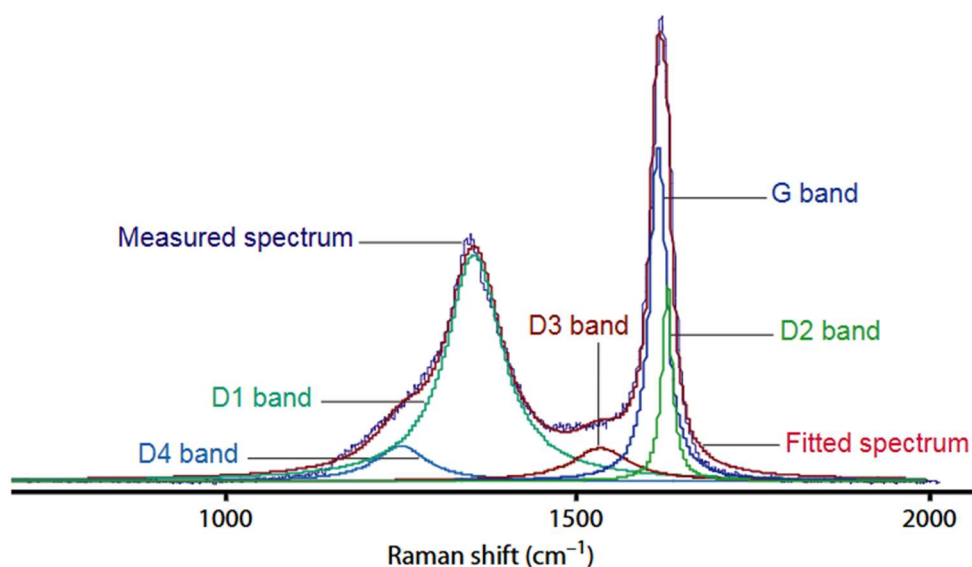


Figure 3-9: Peak-fitting of the Raman spectrum of disordered CM from low-grade metamorphic rocks. This spectrum is composed of a G band including the graphite G s. s. band and the D2 defect band, a defect D band including the D1 and D4 bands, as well as a D3 band between these two bands (see Sadezky et al. (2005) for band labelling). For peak-fitting, each of the five bands has a pure Lorentzian shape and no a priori restriction is imposed on the band position, FWHM or intensity. RSCM in low-grade metasediments.



Figure 3-10: Renishaw (Wotton-under-Edge, UK) In VIA Reflex micro spectrometer (ENS Paris).

3.6 Data Integration and Interpretation

All datasets, including structural measurements, petrographic results, geochemical data, sedimentary structures, and P–T estimates, were integrated into a regional tectonic framework. The synthesis aimed to link deformation and metamorphism at Tempa Roccarossa Mélange and LAC in Cilento to sedimentation patterns of the Cilento Group with contemporaneous events, thus contributing to a broader understanding of orogenic processes in the Southern Apennines.

3.7 Paleocurrent Analysis

For the assessment of turbidity paleocurrent directions, 74 different locations in Cilento were analyzed, spanning from the southern Pisciotta–Mt. Sacro ridge to the northern town of Agropoli, resulting in the collection of a total of 338 paleoflow directions. An average of 4–5 paleocurrent indicators per site were measured, with data distinguished according to their host rocks. As the succession was heavily deformed, especially in the lower part, for every paleocurrent orientation, the bedding attitude was recorded to allow the retro-deformation of the paleocurrent lineation, restoring the bedding in the horizontal attitude, assuming that the original paleocurrent plunge, which is less than or equal to the turbidite bed dip angle, measures

a few degrees. The retro-deformation of the paleocurrent and plotting of bedding orientation data were performed using the software TectonicsFP©1.7.9 (Reiter, 2023). Subsequently, the retro-deformed data were analyzed as rose diagrams for the final evaluation of the mean azimuth. Finally, paleocurrent lineations were plotted in a geological map of Cilento as arrows for each site, according to their original turbidity paleocurrent directions.

3.7.1 Structural Correction

Bedding attitudes were measured at each site. Measured paleocurrent indicators were rotated back to horizontal using TectonicsFP©1.7.9 software (Reiter, 2023). This correction accounts for post-depositional tilting and folding of the strata due to thrusting and nappe stacking in the Apennine orogen.

3.7.2 Data Processing and Visualization

Restored paleocurrent azimuths were input into Tectonics software to construct rose diagrams. Clustering and directional consistency were analysed to identify prevailing sediment transport trends and their temporal/spatial evolution (Figure 3-12).

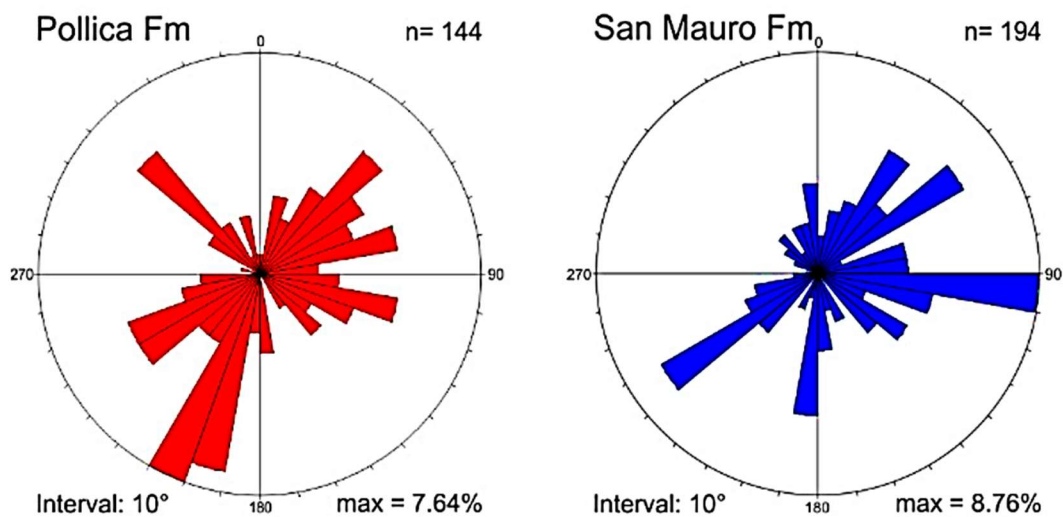


Figure 3-11: Example of data generated through TectonicsFP©1.7.9 software (Reiter, 2023).

Chapter 4 The Tempa Roccarossa Tectonic Mélange

4.1 Introduction

In orogenic belts, *mélanges* are defined as exhumed chaotic rocks characterized by a block-in-matrix texture. They are the result of sedimentary (submarine mass transport), tectonic (thrust-related shearing), and mud-diapiric (fluid/gas-driven remobilization of mud/shales) processes (Ogata et al., 2019). They are therefore classified as sedimentary, tectonic, or diapiric (Festa et al., 2019). Hsü (1968) defines a tectonic *mélange* as mappable bodies of deformed rocks incorporating mixed fragments or blocks within a pervasively sheared fine-grained matrix. In general, the texture of these rocks appears chaotic due to the disruption of the original sedimentary features, resulting in a lack of internal continuity of strata (Cloos, 1983; Evans & Brown, 1986). *Mélanges* associated with faults or shear zones are also known as “sheared *mélanges*” (*sensu* Needham, 1995) or “asymmetric *mélanges*” (*sensu* Fisher & Byrne, 1987; Hammond, 1987).

Overall, tectonic *mélanges* are common in subduction zones (Cloos & Shreve, 1988; Federico et al., 2007; Kimura et al., 2012; Zagorevski et al., 2012; Wang et al., 2013; Penniston-Dorland & Harvey, 2023) and forearc basins (e.g., Huang et al., 2006; Brueckner et al., 2009; Chang et al., 2009). They can form in association with major fault zones (e.g., Brueckner et al., 2009) in accretionary complexes (e.g., Cloos, 1984; Barnes & Korsch, 1991; Kimura et al., 2012; (Mazzoli, Barkham, et al., 2001; Richter et al., 2025). During their formation, the chaotic structure of *mélanges* can incorporate both 'exotic' (extraformational or allochthonous) and 'native' (intraformational or autochthonous) materials (Ogata et al., 2019). The formation of tectonic *mélanges* may persist for long periods, progressively incorporating younger native syn-orogenic material, including fossil content, across successive deformation events (e.g., Cardello et al. (2021). Brittle-ductile transitional *mélange* contexts develop at their seismogenic roof décollement, where *mélanges* form as plate boundary fault rocks in the subduction zone (Kitamura et al., 2005).

Mélanges exhumed from more advanced metamorphic contexts are common in Alpine settings, where the lithofacies display pervasive, general retrograde greenschist facies metamorphism, along with a mylonitic transposition fabric (e.g., Vissers et al., 2001; Roda et al., 2020). At low-pressure-temperature (P-T) conditions, their textures can still carry information on the originating processes (Festa et al., 2010; Ogawa et al., 2011; Codegone et al., 2012; Festa et

al., 2012). In these contexts, microstructural and compositional studies of *mélange* tectonites precede in situ geochronological studies as they are crucial for investigating orogenic dynamics (e.g., Byrne, 1984; Cowan, 1985; Fukui & Kano, 2007; Kimura et al., 2012; Wang et al., 2013; Cardello et al., 2019; Penniston-Dorland & Harvey, 2023). This can be achieved by integrating thermochronological data with geochronological, biostratigraphic and petrologic constraints.

To gain insights into the processes active in tectonic *mélanges*, we have developed an interdisciplinary study in a key area that, besides the general processes, also allows a better definition of the geometry and evolution of the nappe stack of the southern Apennines (Italy) (e.g., Roure et al., 1991; Mazzoli, Barkham, et al., 2001; Mazzoli, Zampetti, et al., 2001; Patacca & Scandone, 2007). This orogen is largely made of non-metamorphic terrains (e.g., Ciarcia and Vitale 2025 and references therein). However, the studied outcrop is located on the hanging wall of a large, deep segment of a *décollement* that guided thrusting within the Upper Miocene-Lower Pliocene accretionary wedge, whose P-T conditions, strain, and tectonic evolution are currently poorly constrained (e.g., Mazzoli, 1992; Catalano et al., 2004). We have investigated the kinematics and the finite strain recorded by these highly deformed rocks, which were also characterized by their mineral and whole-rock geochemical compositions, to estimate the temperature and pressure associated with maximum burial. This study aims to shed light on the mechanisms and processes at work in tectonic *mélanges*, which accommodate most of the crustal shortening during orogenic processes, using the southern Apennines as a case study.

4.2 Geological Setting

The southern Apennines (Figure 4-1 a), together with the Calabria-Peloritani Terrane (CPT), are part of the circum-Mediterranean Alpine chain (Figure 1b), formed following the subduction of the Ligurian oceanic lithosphere beneath the European plate (e.g., Schenk, 1980; Doglioni, 1991; Lustrino et al., 2009; Vitale et al., 2019a). At a regional scale, the orogenic chain is characterized by the superposition of three complexes (Figure 4-1 c; Ciarcia and Vitale (2025), and references therein): on the top, remnants of the overriding plate, consisting of a Paleozoic Variscan crystalline basement and a sedimentary Meso-Cenozoic cover (Complex 1; Sila, Stilo, Bagni and Castagna units; Figure 4-1 a; Amodio-Morelli, 1976; Schenk, 1980; Ciarcia & Vitale, 2025), superposed onto greenschist to high pressure-low temperature (HP-LT) metamorphic thrust sheets, including meta-ophiolites and slices of Adria continental cover

forming the exhumated subduction channel (Complex 2; Diamante-Terranova, Malvito, Gimigliano, Frido, Lungro-Verbicaro and Cetraro units; Figure 4-1 a; Amodio-Morelli, 1976; Vitale et al., 2013a; Vitale et al., 2019a). The latter, in turn, covers a fold and thrust belt (Complex 3) made of the imbrication of several thrust sheets consisting of oceanic cover on the top (Ligurian Accretionary Complex, LAC; e.g., Vitale & Ciarcia, 2013b; Vitale et al., 2019a) and references therein), and shallow water to pelagic successions at the bottom, including Apennine Platform (ApeP), Lagonegro-Molise Basin (LMB), and buried Apulian Platform (ApuP) units; Figure 4-1 a; Ciarcia & Vitale, 2025) and references therein). Exposed carbonates of the Puglia region and part of the Adriatic Sea represent the current foreland. The oceanic subduction started in Paleocene-Eocene times (Schenk, 1980; Vitale et al., 2019a) and references therein), with the underplating of oceanic and continental slices of both overriding and downgoing plates, subsequently exhumed from the Oligocene to the Late Miocene (e.g., Ciarcia and Vitale, 2025 and references therein). Starting in the Early Miocene, an oceanic accretionary wedge formed from off-scraped successions, which thrust over the western margin of the ApeP domain. Subsequently, the thrust front moved to SE/E/NE, including the LMB successions and the remnant of the ApeP successions within the orogenic wedge. Finally, at the end of the Miocene, after the complete closure of the LMB, the orogenic wedge thrust over the ApuP domain. The involvement of the thick continental lithosphere resulted in a buttressing effect and thick-skinned thrusting from the Early Pliocene to the Middle Pleistocene, with the formation of deep-seated thrust faults that also crosscut the allochthonous wedge, forming several out-of-sequence thrusts (Vitale et al., 2020b). In the Early Pleistocene, post-orogenic extension shifted from W to E, affecting the chain by forming several coastal and intramontane plains, accompanied by volcanism (e.g., Santangelo et al., 2017; Fernandez et al., 2025). The deep structure of the southern Apennines (Figure 4-1 d) consists of the superposition of the allochthonous wedge made of LAC, ApeP and LMB thrust sheets on a para-allochthonous wedge, corresponding to the buried and shortened ApuP succession.

4.3 M $\acute{e}$ lange

The study area of Tempa Roccarossa is in the Basilicata region along the northeastern slope of Mt. Sirino (Figure 4-1 e), a large anticline composed of LMB units underlain by ApeP and the LAC through low-angle normal faults (e.g., Mazzoli et al., 2006). The Tempa Roccarossa area has been mapped by the SGI (1969) and ISPRA (2014a). Here, the tectonic architecture is defined by the superposition of two main LMB thrust sheets. The succession forming the

backbone of the Mt. Sirino (lower LMB unit or Lagonegro I unit of Scandone et al. (1967) is composed of Jurassic-Lower Cretaceous pelagic basin sedimentary units, including cherty carbonates (*Calcari con Selce* Formation), radiolarites and silicified limestones and argillites (*Scisti Silicei* Formation), and silicified argillites (*Flysch Galestrino* Formation; Ciarrapica & Passeri, 2000; ISPRA, 2014a). This succession is tectonically covered by another LBM thrust sheet (upper LMB unit or Lagonegro II unit of Scandone (1967), (Figure 4-2 a, b, and f), which includes at the base a Triassic succession of reefal bodies embedded in argillites, silts, radiolarites and sandstones (*Monte Facito* Formation; (Ciarrapica & Passeri, 2000; ISPRA, 2014a).

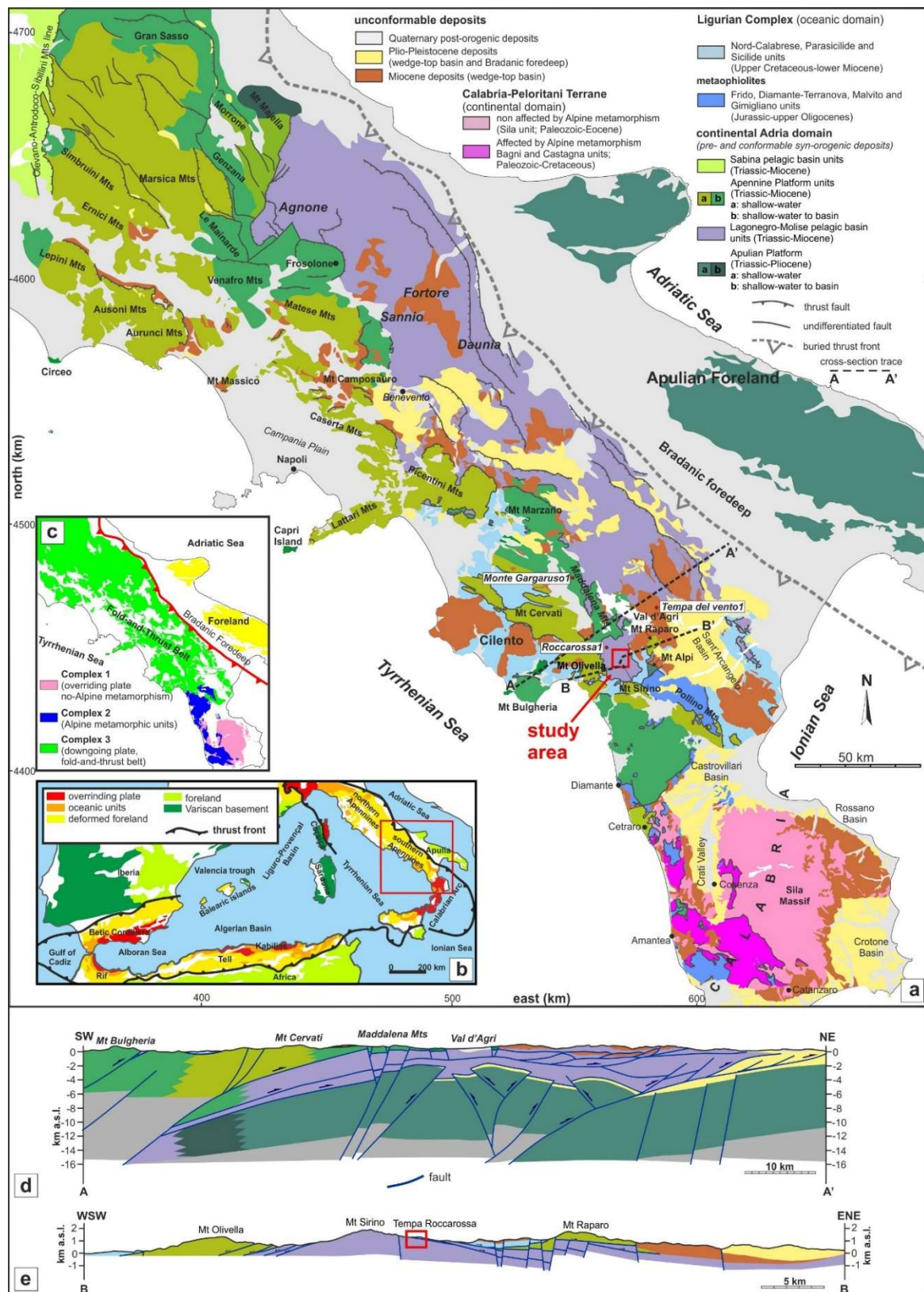


Figure 4-1: (a) Geological map of the northern Calabria-southern Apennines System (modified after (Vitale & Ciarcia, 2022)). (b) Tectonic sketch of western-central Mediterranean Alpine belts (modified after Vitale et al. (2015)). (c) Map of the orogenic complexes of the Apennines. (d) A-A' geological cross-section (modified after Mazzoli et al. (2013)). (e) B-B' geological cross-section including the studied area.

4.3.1 Tempa Roccarossa Mélange

In the southern Apennines, the term “tectonic mélange” usually refers to highly deformed rocks exposed in a few outcrops and drilled at depth for oil exploration (Corrado et al., 2002; Mazzoli et al., 2006); also called “Irpinian units” in several well-logs, (ViDEPI-Project, 2025). These rocks are set between allochthonous LMB units and between LMB and the ApuP units (Figure 1d; Mazzoli, Barkham, et al., 2001; Mazzoli, Zampetti, et al., 2001; Corrado et al., 2002; Butler et al., 2004; Catalano et al., 2004; Ascione et al., 2013; Rinaldi et al., 2020; Schirripa Spagnolo et al., 2024). Borehole data indicate a maximum thickness of a few hundred meters and a succession formed by intensely deformed shales and siltstones, with exotic fragments of sandstones and limestones. Cuttings from well logs show that the mélange results from the fragmentation of LMB and ApuP successions. The mélange is also exposed in different localities (Ciarapica & Passeri, 2000; Van Dijk et al., 2000; Corrado et al., 2002; Mazzoli et al., 2006; Catalano et al., 2007; Passeri, 2010), including the area between Mt. Alpi, Mt. Sirino (Tempa Roccarossa) and Sasso di Castalda (Figure 4-1 a).

At Mt. Alpi, the mélange is tectonically juxtaposed to the upper Messinian clastic unit (*conglomerati ed arenarie di Cima S. Croce* Formation; “Lauria” sheet; (ISPRA, 2014a) at the top of the ApuP succession and to the base of the LMB thrust sheet. The Messinian succession consists of flattened polygenic conglomerates, sandstones with carbonate cement, and shales with subordinate sandstones from the bottom to the top. The tectonic mélange is composed of a heavily deformed argillaceous silty matrix, including black-brown, red, and green pelites with blocks of calcareous sandstones, micritic limestones, radiolarian cherts, laminated black algal limestones, and calcarenites (Van Dijk et al., 2000; Alberti et al., 2001; Mazzoli et al., 2006). In other localities, the mélange is embedded between LMB thrust sheets and is typically located at the base of the Lower Triassic *Monte Facito* Formation (Ciarapica & Passeri, 2000; Passeri, 2010; Cavalcante et al., 2023).

The analyzed tectonic mélange crops out at Tempa Roccarossa, east of Mt. Sirino (Figure 4-2 a and Figure 4-2 f). The latter is a 20 km-sized, dome-shaped anticlinal structure characterized by the interference of two sets of folds and thrusts, the first verging to NE/E/SE and the second to N (Mazzoli, 1992; Mazzoli, 1995). The analyzed mélange is located at the base of the upper LBM unit, at the contact with the underlying lower LBM unit (Figure 4-2 c, d, and f). Information about the deep structure of the area comes from the available well logs of *Roccarossa1*, *Tempa del Vento1* and *Monte Gargaruso1* (Figure 4-1; ViDEPI-Project, 2025). In particular, the Roccarossa1 well is located close to the study area (Figure 4-1 a). The

reconstructed stratigraphic logs (Figure 4-2 e) reveal a structure characterized by the *mélange* located above the Apulian carbonates, and embedded between two LBM thrust sheets, the latter corresponding to the exposed section of the study area.

Mazzoli (1992) described the Tempa Roccarossa *mélange*, associated with a regional shear zone, as highly deformed chlorite claystones characterized by rotated porphyroclasts and shear bands, indicating a mean tectonic transport to NE. The Author reconstructed two deformation phases, the first (D1) responsible for forming a main foliation, and a second phase (D2) characterized by structures folding the previous foliation. Catalano et al. (2007) studied some outcrops of the LMB unit, including the Tempa Roccarossa *mélange*, and reconstructed a polyphase deformation evolution. They show the development of a main foliation (S1) associated with the formation of the tectonic *mélange* (stage D1) defined by a preferred orientation of platy minerals with associated intrafolial folds. These structures are deformed by millimeter to centimeter folds (stage D2) with NE–SW trending axes producing a crenulation cleavage (S2). In addition, D1 and D2 structures are deformed by decimetric to metric thick shear zones (stage D3) that form an S–C fabric, producing an anastomosed pattern formed by lenticular microlithons bounded by shear planes. Kinematic indicators define a top to the NE direction of shear. The latest shortening structures are NW–SE trending conjugate thrust faults with associated asymmetric folds (stage D4). The Authors also determined temperature estimates using the Kubler Index for samples from the *Monte Facito* Formation collected at several locations, including the Tempa Roccarossa locality. Their results suggest a maximum temperature range of 220–250 °C and a burial depth of 8–9 km. However, contrary to Mazzoli, (1992), the Authors associate the *mélange* with a shear zone located between LMB units and overlying ApeP units.

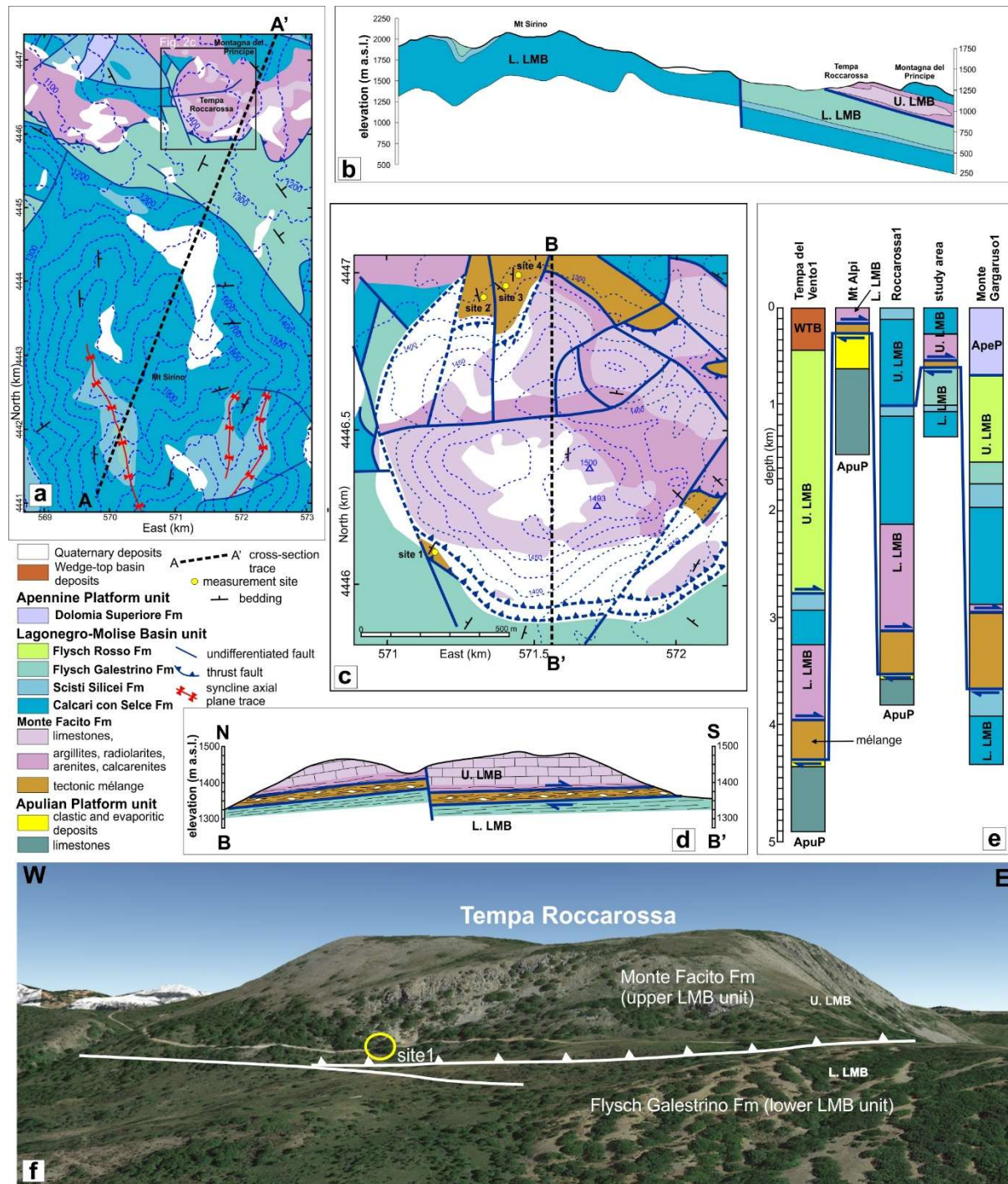


Figure 4-2: (a) Geological map of Mt. Sirino (modified after (ISPRA, 2014a). (b) Geological cross-section A-A'. (c) Geological map of Tempa Roccarossa (modified after (G. Ciarapica & L. Passeri, 2000). (d) Geological cross-section B-B' (modified after ISPRA (2014a). (e) Schematic well logs (modified after ViDEPI-Project (2025). The locations of the wells are shown in Figure 1a. UTM projection, zone 33, datum WGS84. (f) Panoramic view of Tempa Roccarossa. L. LMB and U. LMB: Lower and upper Lagonegro-Molise Basin units, respectively.

4.4 Materials and Methods

In the Tempa Roccarossa study area, we collected several samples from Site 1 (Figure 4-2 c; SIR1, ML1-6) close to the main thrust fault. They are characterized by calcareous and quartzose clasts, less than a few centimeters in size, embedded in a well-foliated argillitic matrix. In Site 2, located at a distance from the thrust, we collected two samples (ML14, 15). These rocks are defined by a few clasts and a dominant argillitic matrix with a weak foliation. Twenty subsamples were taken from the sample SIR1 and ML1-6 to obtain thin sections and polished slabs for the structural and petrographic characterization parallel to the XZ, YZ, and XY planes of the Finite Strain Ellipsoid (FSE). As explained in the following paragraphs, the analyzed samples are S-tectonites, characterized by oblate strain, and are defined by a dominant tectonic foliation, with stretching lineations being rare or absent. Therefore, to evaluate the X, Y, and Z axes of the FSE, other features such as synthetic and antithetic S-C' structures (R1 and R2, respectively) and rotated porphyroclasts have been considered. The results indicate that the XZ is defined by an NW-SE direction (shear direction) with the XY plane approximately horizontal. We then cut samples orthogonal to the main foliation (XY plane) and parallel and orthogonal to the shear direction (NW-SE). In total, we obtained 13 thin sections (ML1a, 2a, b, 3a, 4, 5, 6a, b) cut parallel to the XZ plane and six (ML1b, c, 2c, 3b, c, 6c) cut parallel to the YZ plane, and one (SIR1) cut parallel to the XY plane.

4.4.1 Mineralogical, Mineral Chemical and Whole-Rock Geochemical Analyses

Mineralogical XRPD (X-Ray Powder Diffraction), mineral chemical SEM-EDS (Scanning Electron Microscopy Energy Dispersive Spectrometry) and whole-rock geochemical XRF (X-Ray Fluorescence Spectrometry) analyses were performed on a selection of five representative samples (SIR1, ML1, 2, 14, 15). Two additional samples (ML3, 6) were analyzed only for mineral chemical SEM-EDS characterization.

4.4.2 XRPD Analysis

X-ray diffraction (XRPD) analysis was conducted to identify and quantify the mineral phases in the samples SIR 1, ML 1, ML 2, ML 14, and ML 15. Prior to analysis, the samples were dried at room temperature, gently crushed, homogenized, and split into two aliquots. One portion was used for both XRPD and X-ray fluorescence (XRF) analyses, while the other was used for wet extraction of the clay fraction ($<2\ \mu\text{m}$) following Stokes' law (Moore & Reynolds,

1997). XRPD was conducted on a Philips X'Pert Pro instrument with a Co anode (40 kV, 40 mA) and X'Celerator detector at the Università di Padova. Data were collected from 3° to 85° 2 θ for both bulk and oriented clay samples. Bulk samples were prepared by side-loading (Środoń, 1999), and clay fractions were treated with 1N MgCl₂, air-dried, glycolated, and heated (Moore & Reynolds, 1997). Quantitative analysis used Winfit software (Krumm, 1999) for peak area, intensity, and FWHM. Results were compared with whole-rock data using vbAffina software (Cavalcante et al., 2007). Semi-quantitative clay mineral analysis was done on glycolated and heated XRD patterns. Paleotemperatures were estimated using the Kübler Index (KI) method (Warr & Rice, 1994), recalibrated by Frey and Stern (Warr & Mählmann, 2015). FWHM measurements had an error of $\pm 6\%$, corresponding to $\pm 10^\circ\text{C}$ (Warr, 2018).

4.4.3 XRF Analysis

For XRF analyses, powdered rock samples were used to prepare pressed powder pellets, which were analyzed for major oxides and trace elements using a Panalytical Axios instrument housed at the DiSTAR. Analytical uncertainties are in the order of 1%–2% for major oxides and 5%–10% for trace elements. Weight loss on ignition (LOI) was determined gravimetrically after heating 1 g of rock powder (pre-dried at $\sim 150^\circ\text{C}$ overnight) at 900°C for three hours.

4.4.4 SEM-EDS Microanalysis

The SEM-EDS microanalyses were carried out at the DiSTAR using an Oxford Instruments Microanalysis Unit equipped on a Zeiss Merlin VP Compact Field Emission SEM. The technique has been demonstrated to provide reliable microanalytical data at the major oxide's concentration levels (see Guyett et al. (2024) and references therein). Measurements were performed using a 15-kV primary beam voltage, 60 μA filament current, variable spot sizes and a 10-s of acquisition time. Corrections for matrix effects were performed using the XPP correction routine based on a Phi-RoZeta approach. Standards for calibration were: anorthoclase (Na, Al, Si); microcline (K); serandite (Mn); diopside (Ca); fayalite (Fe); olivine (Mg); rutile (Ti); apatite (P); fluorite (F); barite (Ba); celestine (Sr); zircon (Zr and Hf); synthetic Smithsonian orthophosphates (La, Ce, Nd, Sm, Y and Sc); pure niobium (Nb); pure vanadium (V); pure cobalt (Co); pure nickel (Ni); chromite (Cr); Corning glass (Th and U); sphalerite (Zn); pyrite (S); sodium chloride (Cl); pollucite (Cs). Accuracy and precision were tested using the rhyolitic Lipari obsidian ID3506 and basaltic Laki 1783 tephra from Kuehn et

al. (2011) as secondary standards. Relative analytical uncertainty was <5% for SiO₂, Al₂O₃, K₂O, CaO, and FeO, ~10% for the other elements.

4.4.5 Chlorite and White Mica Multi-Equilibrium Geothermobarometry

A multi-equilibrium approach was used to determine the P and T conditions for forming chlorite and white mica. The chlorite + quartz + H₂O method of (Vidal et al., 2005; Vidal et al., 2006) was used to estimate the formation temperature of chlorite. If the maximum temperatures represent the conditions of peak metamorphism (e.g., Lanari, 2012; Petroccia et al., 2025), the pressure conditions of white mica formation were determined using the mica + quartz + H₂O method of Dubacq et al. (2010). Finally, a P-T estimate was obtained using the chlorite + mica + quartz + H₂O method of (Vidal et al., 2006). Chlorite and white mica analyses have been selected where the equilibria show good convergence in the P-T space, but this is not sufficient to prove chemical equilibrium; it only shows that the mineral composition is consistent with chemical equilibrium being achieved at these P-T conditions (Lanari & Duesterhoeft, 2019). A full description of each method and the limitations of this approach compared to others can be found in (Lanari & Duesterhoeft, 2019; Petroccia et al., 2025). Thermodynamic calculations were performed using the ChlMicaEqui 1.5 program (Lanari, 2012; Lanari & Duesterhoeft, 2019). The following solution models were used: (Vidal et al., 2005; Vidal et al., 2006) for chlorite (chlorite + quartz + H₂O and chlorite + mica + quartz + H₂O methods), (Parra et al., 2002) for white mica (chlorite + mica + quartz + H₂O method) and (Dubacq et al., 2010) for the determination of a P-T line along which mica is in equilibrium with quartz and H₂O (mica + quartz + H₂O method). All solution models are compatible with the equations of state and the internally consistent data set of (Berman, 1988). The activity of H₂O in the fluid was set to 1.

4.4.6 Strain and Structural Analysis

We obtained information about the 3D strain and the kinematic vorticity to understand the 3D strain and the co-axial strain. We analysed several representative samples. Strain markers have been analyzed with the free software EllipseFit (Vollmer, 2018). The technique uses strain markers that can be approximated to ellipses by measuring their ellipticity (R_f) and the angle ϕ that the long axis of the ellipse forms with a reference line (detail methodology see previous chapter).

4.5 Results

4.5.1 Mesostructures

We studied four outcrops in the Tempa Roccarossa *mélange* (Sites 1 to 4 in Figure 4-2 c). Site 1 hosts the most deformed rocks, located near the main thrust contact, close to the underlying Flysch Galestrino Formation located in the footwall. In contrast, the other three sites are more distant from it and generally appear as less deformed rocks. In general, the tectonic *mélange* appears as a well-foliated rock with clasts embedded in a very fine-grained matrix (Figure 4-3).

We recognized three foliations at the outcrop scale: the bedding (S0) and two superimposed tectonic cleavages, S1 and S2. The latter is the main foliation. Rare mesoscale F1 folds are present (Figure 4-3 a), deforming the S0 bedding, generally with a rounded hinge, overturned and with narrow to isoclinal geometries and NW-SE trending axes, parallel to tectonic transport. Along the limbs, an S1/S2 cleavage is observed (Figure 4-3 a). To the naked eye, clasts embedded in the main foliation vary in size from a few centimetres to <1 mm (Figure 4-3 b, g). The largest clasts often have a rectangular shape, commonly rotated, and, in places, show the bedding S0 (Figure 4-3 b, g) and, in some cases, the first tectonic foliation S1 (Figure 4-3 c, g). The latter corresponds to a stylolitic cleavage in the calcareous clasts (Figure 4-3 c) and forms a high angle with the bedding.

At Site 2 (Figure 4-3 c), the *mélange* shows a less intense deformation. In fact, locally, remnants of the protolith formed by dismembered calcareous or arenaceous layers alternate with fine-grained layers (Figure 4-3 h). The S1/S2 foliation is parallel to the original S0 bedding, as indicated by the alternating purple/green colours.

Locally, the S2 is affected by microscale folding, marked by a crenulation lineation (L3) with a subhorizontal NE-SW trend (Figure 4-3 i). Furthermore, the S2 foliation is cut by two sets of normal shear planes, set 1 (R1) forming high-angle planes with the main foliation and set 2 (R2) with a lower angle (Figure 4-3 j, k). In general, planes R1 and R2 dip to the NW and SE, respectively, as indicated by the stereographic projection (Figure 4-3 j). Sites 3 and 4 are similar, showing a lower amount of porphyroclasts and a parallel main foliation (S2) cut only by the R1 and R2 shear planes (Figure 4-3 j).

4.5.2 Microstructures

Polished slabs of *mélange* samples Figure 4-4 a) exhibit a mylonite fabric, characterized by deformed porphyroclasts enclosed within a highly foliated fine matrix. Clasts are affected by ductile strain, frequently displaying an elliptical shape. Thin sections exhibit up to three distinct foliations (Figure 4-4 b, o). S_0 is observed only in the rigid porphyroclasts (Figure 4-3 b, g).

The earliest deformation structure is a cataclasite attached to the porphyroclasts (Figure 4-5 h) made of rounded fragments of the porphyroclast or by an older cataclasite. Porphyroclasts (Figure 4-4 c, k) are made of elliptical fine-grained objects (Figure 4-4 c), large rectangular coarser quartz and calcite grains (Figure 4-4 d, e) or fragments of calcite veins. Small rigid porphyroclasts are generally made of fractured (Figure 4-4 f) or single quartz crystals (Figure 4-4 g, i). Frequently, they show a monoclinical symmetry forming σ shapes (e.g., Passchier & Trouw, 2005).

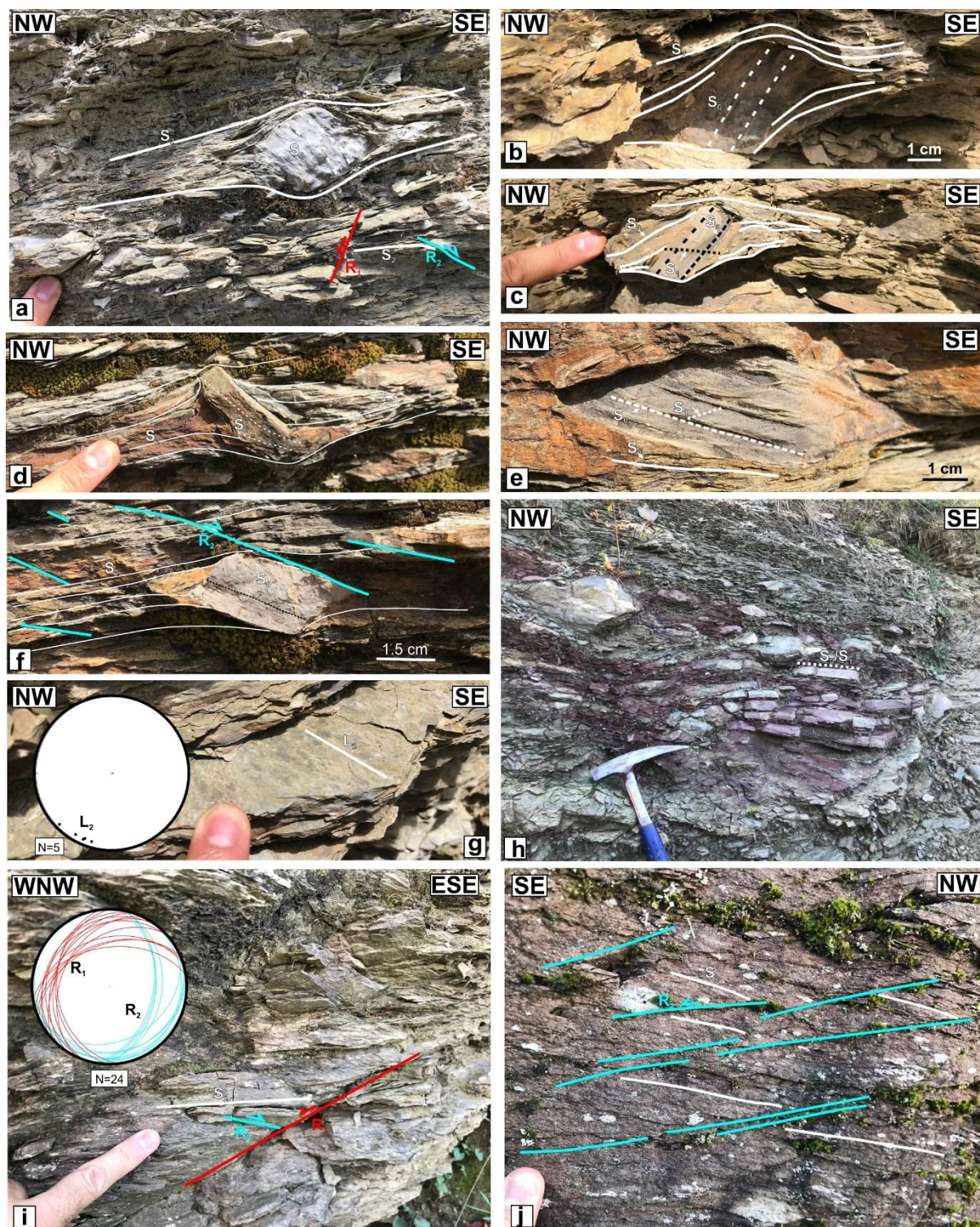


Figure 4-3: Representative field photographs of the Tempa Roccarossa study area showing (a) F1 isoclinal folds (Site 2); (b–g) rotated clasts embedded in a foliated matrix with R 1 and R2 shear planes (Site 1); (h) remnants of the original bedding (Site 2), (i) lineation L2 on the S2 foliation (Site 1); (j) R1 and R2 shear planes and related stereographic projection (Site 3), (k) R2 shear planes (Site 4). Site locations are shown on Figure 4-1c.

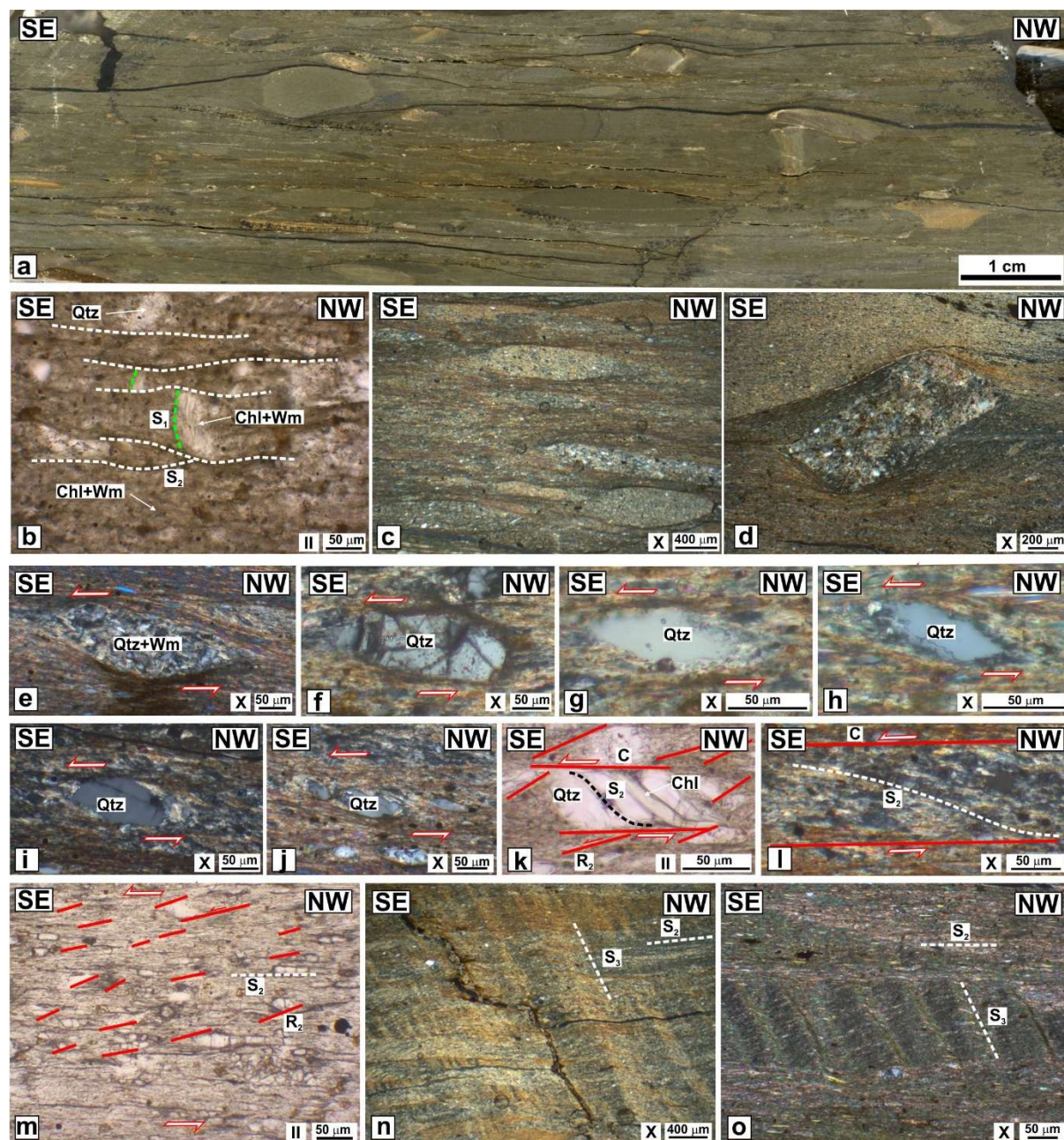


Figure 4-4: Representative microstructures for the investigated Tempa Roccarossa mélange samples, observed on polished thin slab (a) and in thin section photomicrographs (b–o, II for plane polarized, X for crossed polarizers light views). (a) Sample ML1 cut along the SE-NW direction showing rotated rigid porphyroclasts and elliptical markers affected by ductile strain. (b) S1 and S2 cleavages (sample ML2; Qtz, Chl and Wm for quartz, chlorite and white mica crystals, respectively). (c) Elliptical strain marker (sample ML2). (d) Rotated rectangular porphyroclast (sample ML2). (f–l) S-C structures and porphyroclasts with a sigma shape (sample ML6a). (m) R2 shear planes (synthetic S-C' structures) (sample ML3). (n–o) Late cleavage (S3) superposing on main foliation (S2) (sample ML2).

As depicted in Figure 4-4 b, the tectonic foliation associated with a first deformation stage (D1) is a disjunctive cleavage S1 occurring in microlithons of main foliation S2. S1 is characterized by alternations of chlorite and mica crystals. Some veins associated with stage D1 are well-preserved in the strain shadow of large porphyroclasts (Figure 4-5 b, e). These veins are made

of calcite fibers departing from the median line (antitaxial veins), with the youngest generations located to the boundaries, where quartz crystals are often present (Figure 4-5 c). The fibers show a sigmoidal shape (Figure 4-5 c) or a complex geometry (Figure 4-5 d). Locally, early antitaxial veins occur at the necks of boudinaged layers showing the same fibrous sigmoidal growth and quartz cement at the boundaries (Figure 4-5 f, g). At the boundary of the pressure shadow, the veins are deformed by ductile shear (Figure 4-5 e) and isoclinally folded with limbs parallel to the main foliation S₂ (Figure 4-5 a).

As shown in Figure 4-4 b, the latter is a crenulation cleavage that is always defined by chlorites and micas, which generally mantle the porphyroclasts (Figure 4-4 c, k). S₂ cleavage is deformed by S-C (R₁) and S-C' (R₂) structures (Figure 4-4 j, m) and, especially in the very fine-grained parts, is deformed by kink folds, whose axial planes form a cleavage S₃ (Figure 4-4 n, o). σ -shaped porphyroclasts, S-C structures invariably indicate top-to-the-E/SE shear sense.

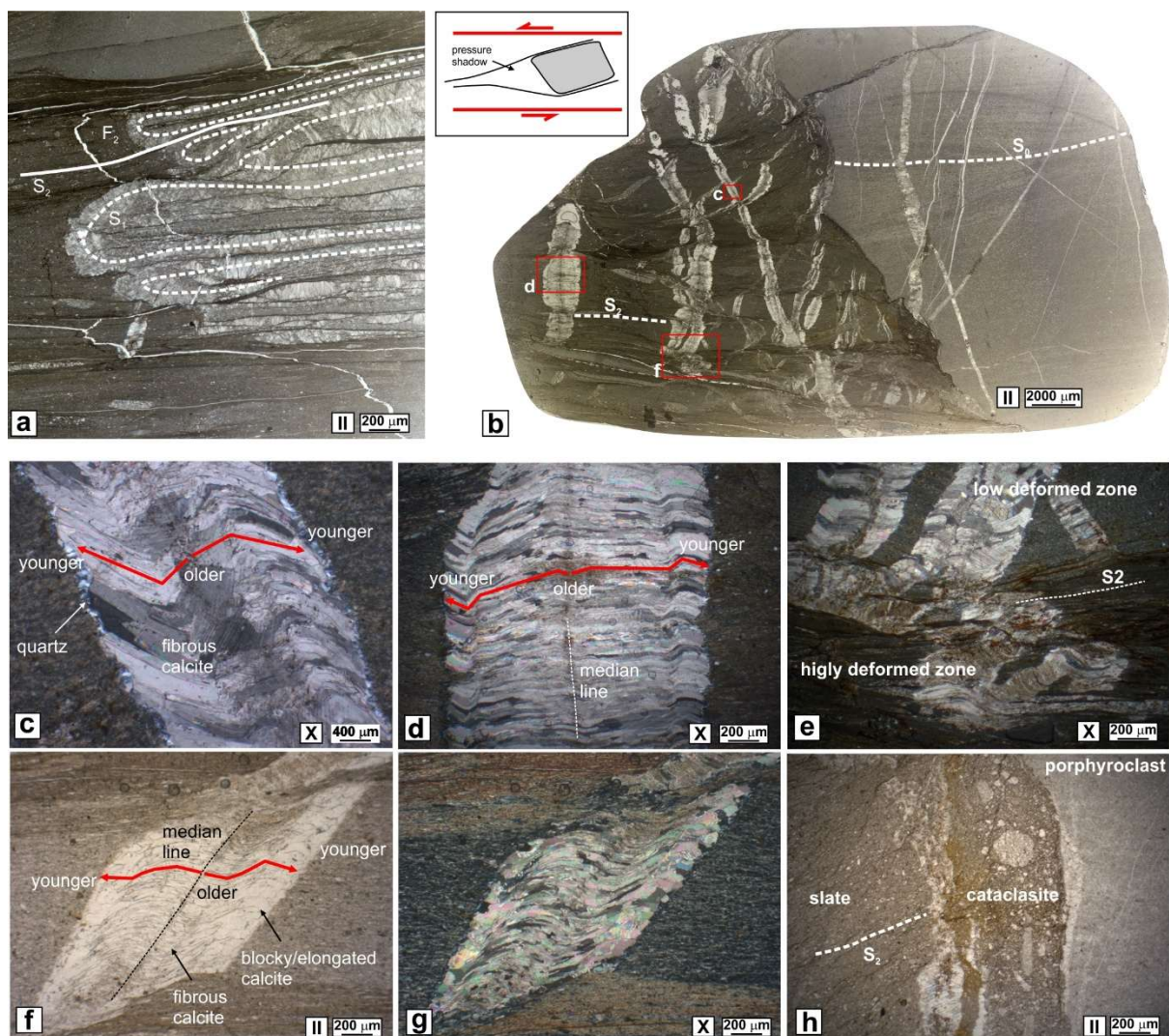


Figure 4-5: Representative microstructures for the investigated Tempa Roccarossa mélange samples, observed in thin sections (II for plane polarized, X for crossed polarized light views). (a) Folded early vein (ML6a). (b) Veins located in a pressure shadow of a rotated porphyroclast (ML6a). (c–e) Close-up view of some veins of picture (b): (c) antitaxial vein with sigmoidal growth of calcite fibers and quartz cement at the boundaries; (d) fibrous vein with a medial line; (e) deformed early vein; (f–g) Antitaxial vein in the neck of a stretched layer (ML6b). (h) Remnant of cataclasite in a pressure shadow of a porphyroclast (ML6b).

SEM investigations revealed that a disjunctive cleavage (S_1) is deformed by a main crenulation cleavage (S_2) that partially or totally transposes previous structures. In the highly deformed zones (Figure 4-6 a), only the main foliation (S_2) marked by the orientation of white mica, chlorite, elongated quartz, rutile, and apatite crystals is observed. In the lower deformed areas, relicts of S_1 foliation are preserved in microlithons (Figure 4-6 b, f). The S_1 is a disjunctive cleavage marked by the alternation of white mica and chlorite crystals, which are occasionally folded (Figure 4-6 d, e). Locally, the main foliation S_2 is crenulated (Figure 4-6 f) with a less developed new foliation S_3 . Porphyroclasts and porphyroblasts, made of chlorite, quartz, rutile and pyrite, are also observed (Figure 4-6 g). Generally, micro-shear planes R_1 and R_2 occur (Figure 4-6 h). Finally, syn-kinematic antitaxial veins show quartz as the last growth at their boundaries (Figure 4-6 i).

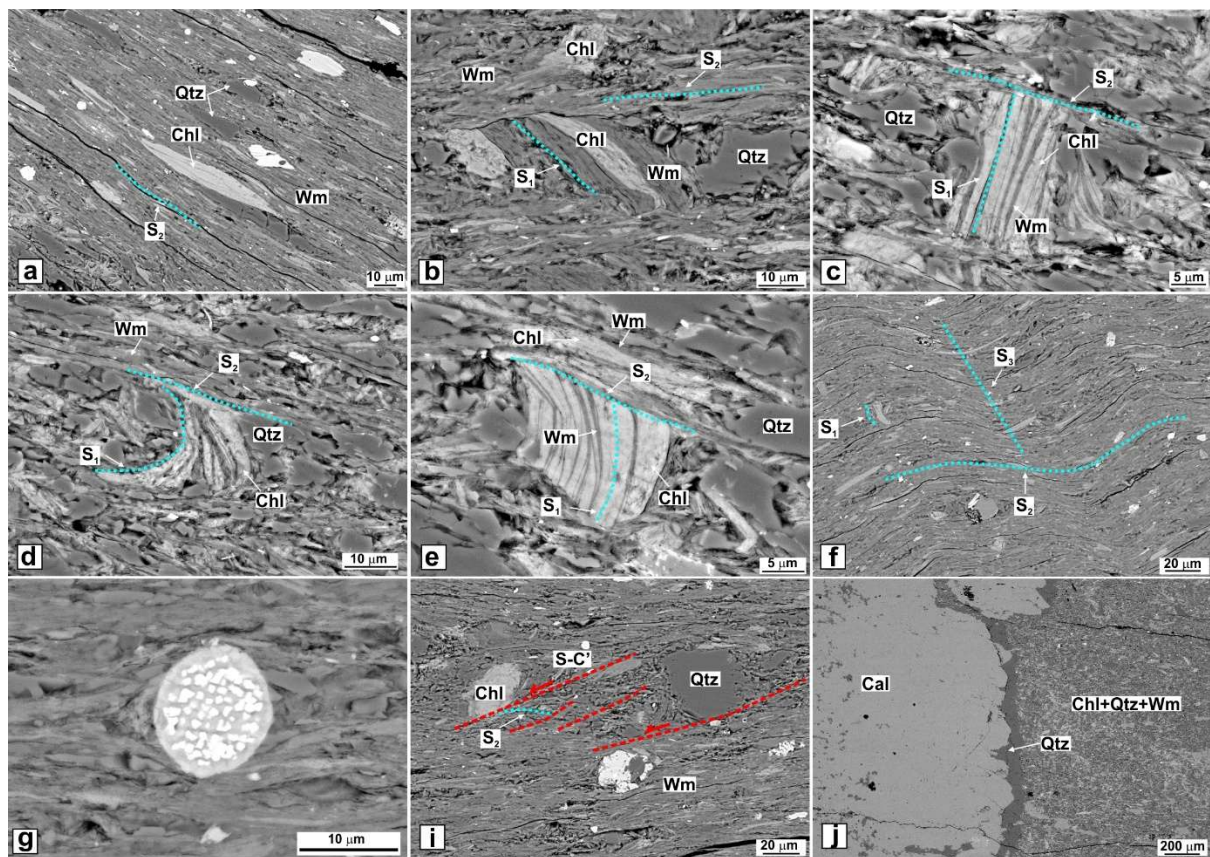


Figure 4-6: Representative back-scattered electron SEM images for the investigated Tempa Roccarossa mélange samples. (a) Highly deformed area, showing the main foliation (S_2) marked by the iso-orientation of white mica (Wm), chlorite (Chl), and elongated quartz (Qtz) (ML6). (b–e) Examples of relicts of S_1

foliation in microlithons marked by Wm and Chl crystals: (b) ML2; (c–e) ML6a. (f) Relict S1 foliation in microlithons, the main foliation S2 and a new, early S3 crenulation foliation (ML2b). (g) Pre-kinematic grain of pyrite crystals (white) in iron hydroxide (light gray) embedded by Wm (ML2). (h) R2 shear planes (ML2). (i) Boundary between the pelitic Chl + Wm + Qtz matrix and a syn-kinematic antitaxial vein of Qtz and calcite (Cal; ML6a).

4.6 Petrographic Description

The samples investigated are extremely fine-grained and display a weakly-to-developed main foliation (S2) in thin sections. They all comprise a mineral assemblage dominated by white mica (Wm) and chlorite (Chl) crystals. Sporadic porphyroclasts of quartz (Qtz), albite and calcite, accessory apatite, rutile, zircon, pyrite, and Fe- hydroxides (identified during SEM observations) also occur, as well as veins of quartz and/or calcite.

A textural distinction is tentatively made between samples showing a more evident foliation (ML1,2,3,6 and SIR1) and samples showing a very weak foliation (ML14, 15). For the sake of simplicity, in the following sections, the two sample types will be referred to as “slate” and “metapelite,” respectively.

4.7 Mineralogical Data

The quantitative mineralogical analysis shows that the bulk rock is mainly characterized by the presence of phyllosilicates ranging from 56% to 72%, subordinately quartz (from 14% to 30%) and feldspars (i.e., alkali feldspar + plagioclase) in some percentage units (from 3% to 4%). Calcite is absent in the metapelite sample ML14, whereas in the slate ML2 sample, it reaches 27%. Hematite is present in a few percentages only in metapelite sample ML15. Phyllosilicates consist of illite-muscovite (36–43 wt.%), chlorite plus mixed-layers chlorite/smectite (C/S) (17–24 wt.%). Paragonite (plus K/Na mica) is present in very low amounts (3%–6%) in SIR1, ML1 and ML2 samples (Figure 4-7).

The semi-quantitative mineralogical analysis of the clay fraction shows the main presence of illite-muscovite and chlorite. Mixed layers of chlorite/smectite (C/S) are detectable in lower amounts, while paragonite plus K/Na mica is slightly higher, especially in the slate samples (Figure 4-7). KI value varies from 0.42 in the metapelite sample ML14 to $0.31^{\circ} 2\theta$ in slate SIR 1. Intermediate values are measured in samples ML1, ML2, and ML15. This suggests that metapelite samples (Site 2) were subjected to a less pronounced metamorphic recrystallization than slate ones (Site 1). Therefore, differences in the KI between ML14 and ML15 are due to different textural and/or chemical compositions of the original protolith. In summary, the KI

indicates that the paleotemperatures vary between about 200 ± 10 °C (ML14) and 250 ± 10 °C (SIR 1). This agrees with the presence of paragonite and the very low amount of C/S (Figure 4-7) (Kisch, 1987; Leoni et al., 1996; Cavalcante et al., 2012).

4.8 Mineral Chemistry

4.8.1 Chlorite (Chl)

Chl crystals are generally larger compared to Wm ones but still very small, thus requiring some particular care in evaluating the reliability of mineral chemical data. The performed analyses ($n = 466$) were thus selected by discarding data with $(\text{CaO} + \text{Na}_2\text{O} + \text{K}_2\text{O}) > 0.5$ wt.% and with the sum of oxides < 83.0 and > 89.5 wt.%, following (Vidal & Parra, 2000; Escuder-Viruete et al., 2011). In addition, analyses with anomalously high TiO_2 (> 0.90 wt.%, where most analyses have < 0.30 wt.%) were also cautiously filtered out as likely indicative of mixed rutile-chlorite data points. Considering the above analytical difficulties, we also tested an alternative data filtering strategy and adopted a more conservative confidence interval of 86.0–89.0 wt% for the sum of oxides. Nevertheless, we did not observe any significant difference in the overall chemical features of the analyzed Chl crystals. This can be observed by comparing the Chl plots described below with their alternative versions reported in Figure 4-8 a-d.

Compositions lie between chamosite and clinocllore (Figure 4-8 a-d), with XMg values $[\text{Mg}/(\text{Mg} + \text{Fe}^{2+})]$ varying mostly from 0.23 to 0.56 (plus one single data point with 0.73), corresponding to ripidiolite brunsvigite types (Hey, 1954). The spread in chemical composition reveals Tschermak $[\text{IVAl}^{\text{VI}}\text{AlSi}_{-1}(\text{Mg,Fe})_{-1}]$ and di/trioctahedral substitutions $[(\text{Mg}, \text{Fe}^{2+})_{3\Box-1}\text{Al}_2]$, where $\Box$ = vacancy; Figure 4-8 a, b]. A relatively large compositional range is covered by Si (2.482–3.009 atoms per formula unit, apfu) and R^{2+} cations (3.323–4.622 apfu), while R^{3+} cations are mostly around 3 and < 0.5 apfu, respectively (except for a single data point with Si 2.976, R^{2+} 3.323 and slightly > 0.5 apfu; Figure 4-8 c, d). No systematic difference seems evident between Chl crystals from the analyzed metapelite and slate samples, as well as between crystals forming the S1 and S2 cleavages in the slate samples.

4.8.2 White Mica (Wm)

Wm crystals have remarkably small grain sizes (generally a few μm in thickness), which made microanalyses extremely challenging. All the acquired data points ($n = 121$) were therefore

carefully selected to exclude contaminated intergrown phyllosilicate analyses with $(\text{TiO}_2 + \text{MnO} + \text{Cl}) > 0.5 \text{ wt.}\%$, as well as unreliable analyses with an oxide sum < 92.0 and $> 97.0 \text{ wt.}\%$ (Vidal & Parra, 2000; Escuder-Viruete et al., 2011). As in the case of Chl, we also tested an alternative data filtering and adopted a more conservative confidence interval of 93.0–97.0 wt.% for the sum of oxides. We again recognized no significant difference in the overall chemical features of the analysed Wm crystals.

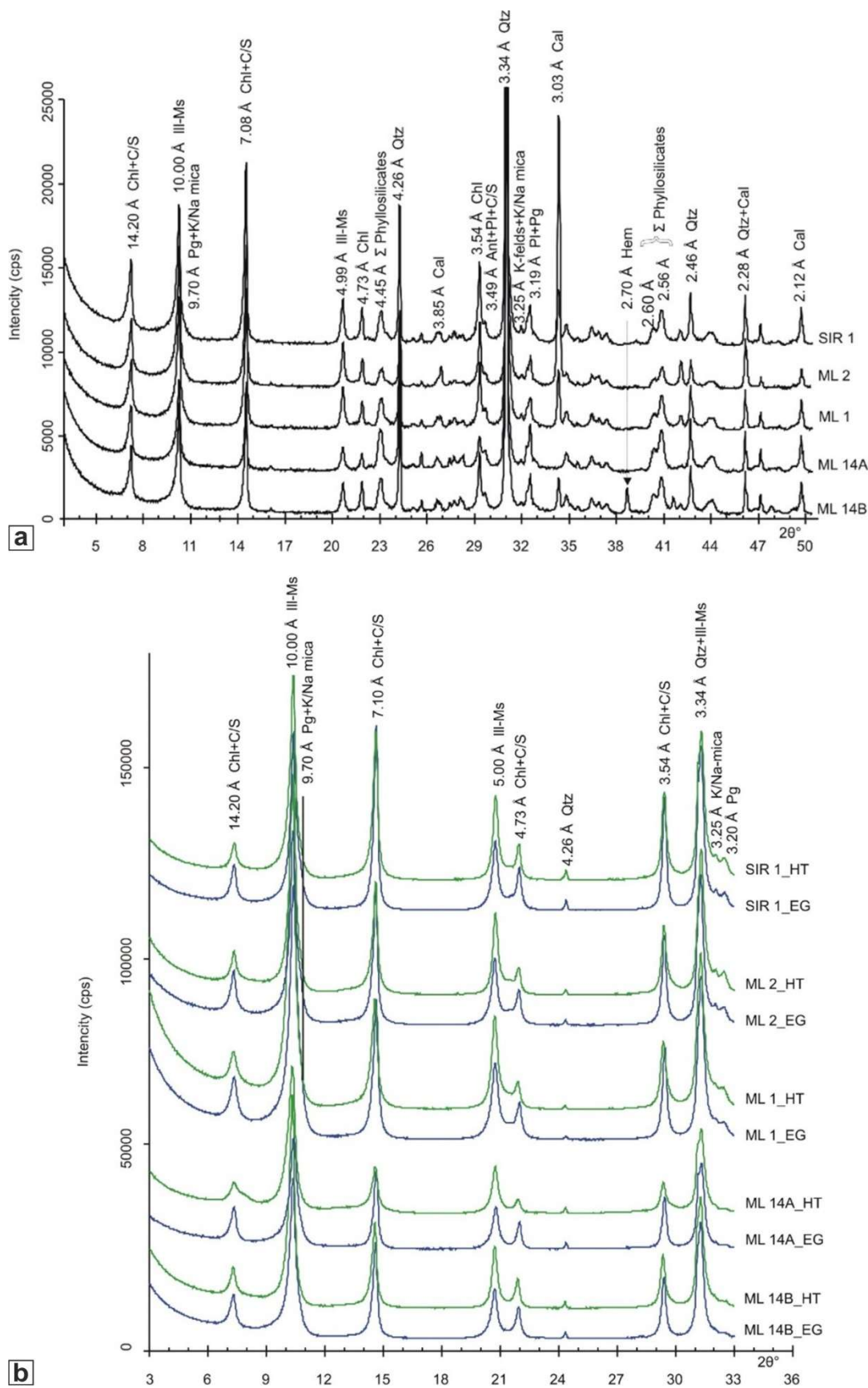


Figure 4-7: XRD patterns of (a) bulk sample and (b) clay fraction. Chl, chlorite; C/S, mixed layers chlorite/smectite; Ill+Ms, illite-muscovite; Pg, paragonite; K/Na mica, White K-Na mica Qtz, quartz; Cal, calcite.

calcite; Pl, plagioclases; K-feld, k-feldspars; Ant, anatase; Hem, hematite. EG, glycolated sample; HT, heated sample.

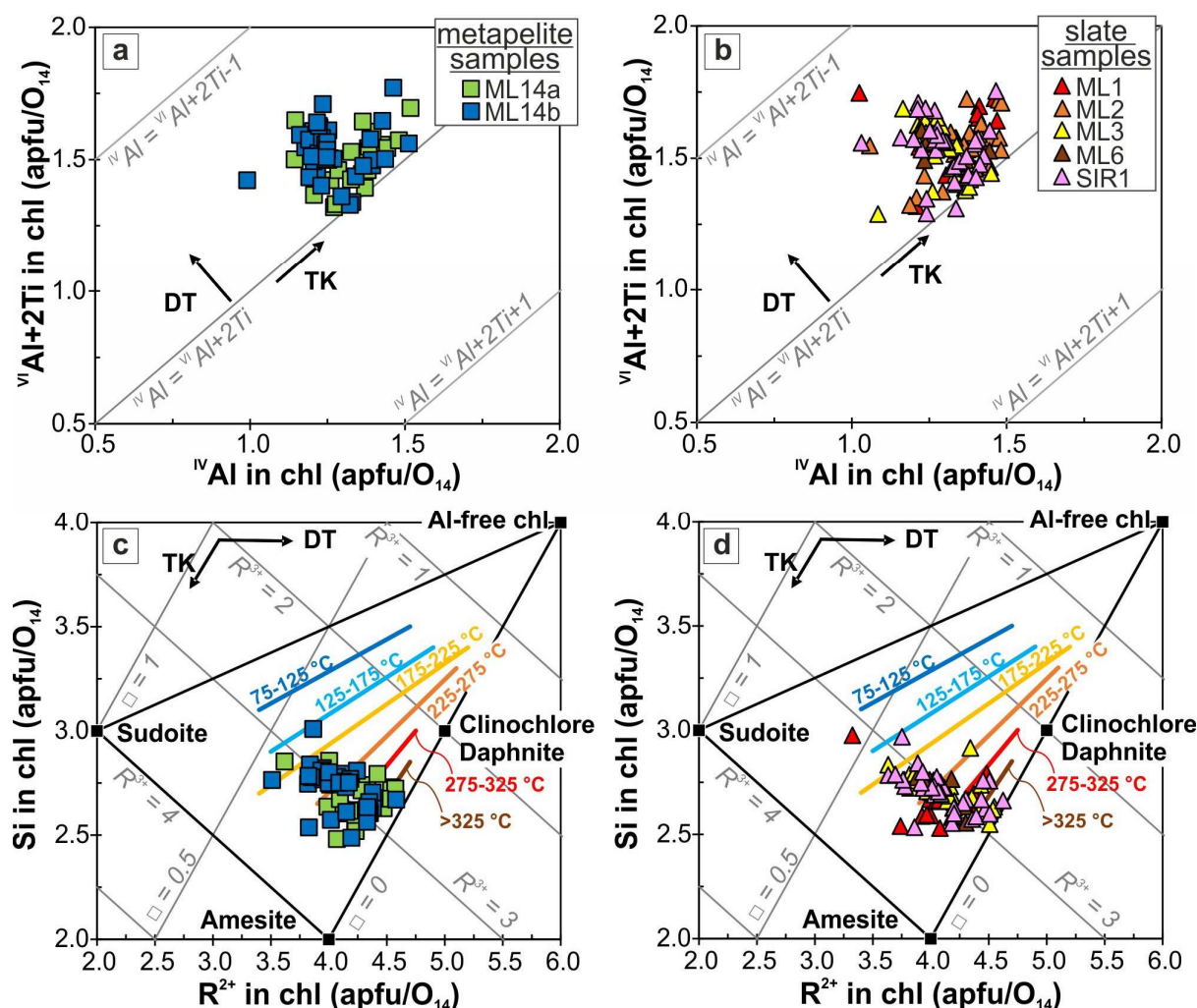


Figure 4-8: (a, b) $^{VI}Al+2Ti$ versus ^{IV}Al and (c, d) Si versus R^{2+} in chlorite crystals (chl) from the investigated Tempa Roccarossa mélange metapelite and slate samples. In (c, d), the semiquantitative isotherms after Bourdelle and Cathelineau (2015) are also reported. Arrows point to the effects of the Tschermak (TK) and di/trioctahedral (DT) mineralogical substitutions.

The samples investigated show relatively large compositional variation, suggesting some complex mineralogical substitutions. Data points spread along the muscovite-celadonite Tschermak exchange line (Figure 4-9 a) close to the Al_{tot} -rich (2.259–2.867 apfu) and Si-poor (3.021–3.382 apfu) end. Deviations point to an excess of Fe and Mg related to di/trioctahedral substitutions and some ferrimuscovite components (Figure 4-9 b). The pyrophyllitic substitution $[KAlSi_{-1}□_{-1}]$ is also significant, as indicated by the large range in the total interlayer charge (t.i.c. = $K + Na + 2Ca$; Figure 4-9 c) ranging from 0.714 to 0.956 apfu. The paragonitic substitution (KNa_{-1}) is generally limited, although some Wm crystals with high Na content were also analyzed (>0.200 and up to 0.656 apfu; Figure 4-9 d). The paucity of Wm data for

the metapelite samples (only two analyses vs. 51 for the slate samples) does not allow to confidently evaluate any systematic compositional difference between the two rock types.

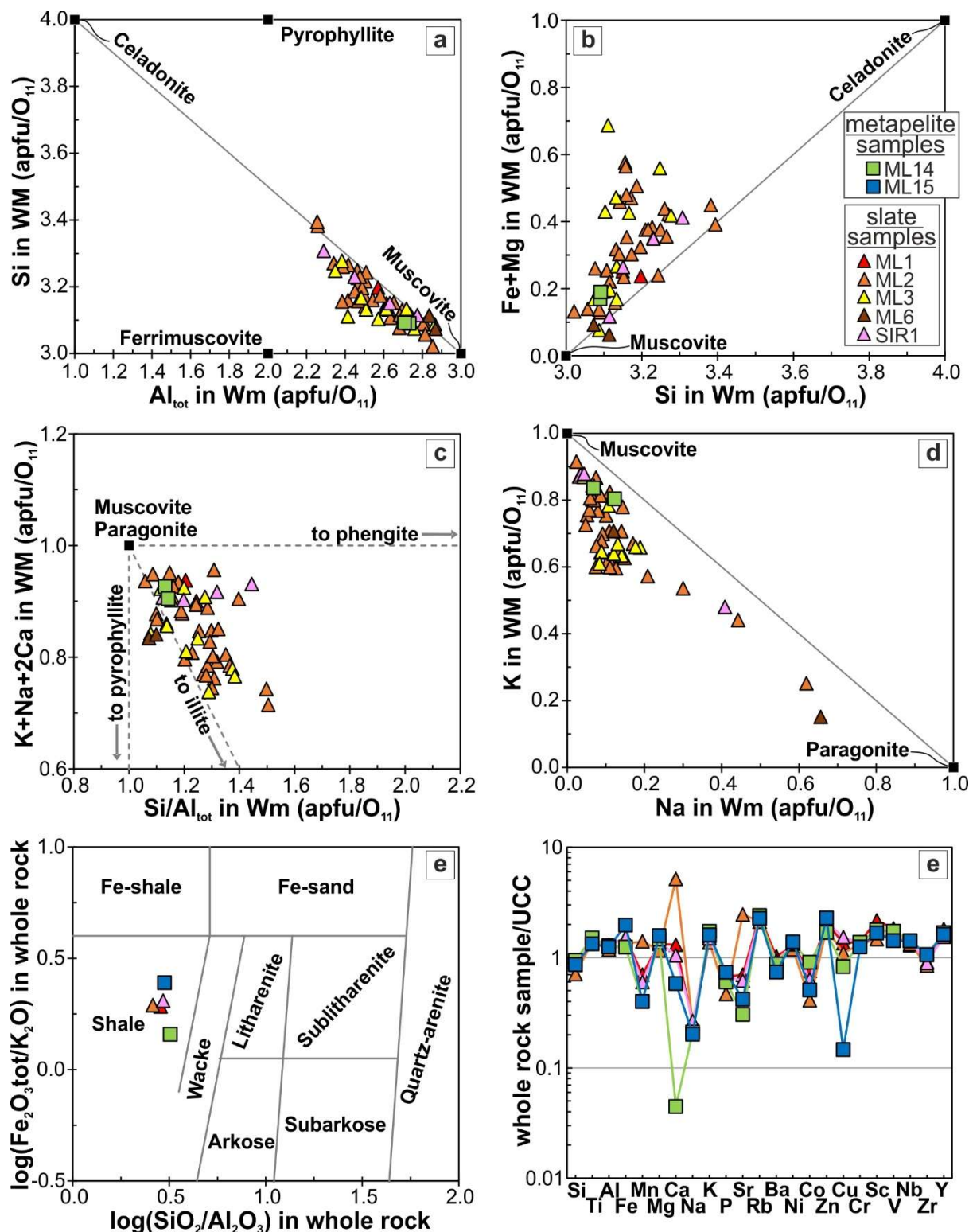


Figure 4-9: (a) Si versus Al_{tot}, (b) Fe + Mg versus Si, (c) K + Na+2Ca versus Si/Al_{tot} and (d) K versus Na in white mica (WM) crystals, (e) log(Fe₂O₃tot/K₂O) versus log (SiO₂/Al₂O₃) (Herron, 1988) and (f) upper continental crust (UCC; (Rudnick & Gao, 2014) normalized major oxides and trace elements in whole-rock for the investigated Tempa Roccarossa mélangé samples.

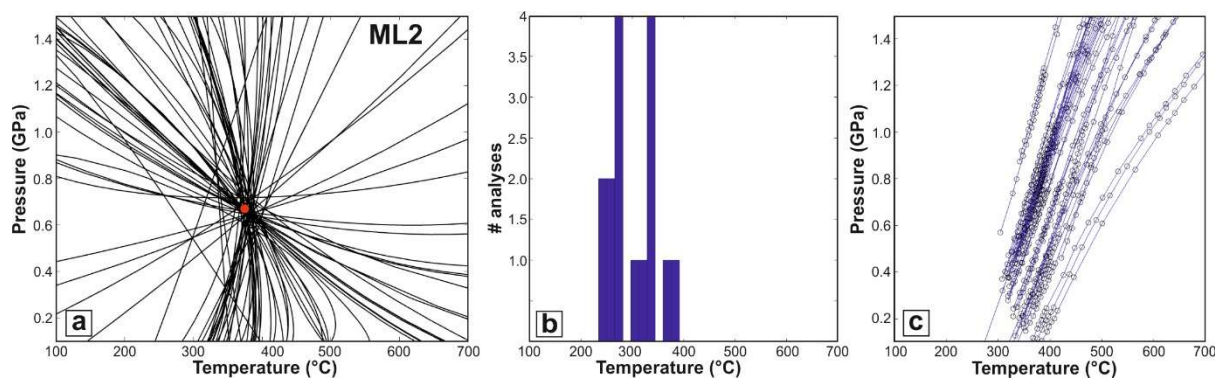


Figure 4-10: Results of (a) Chlorite + Mica + Quartz + H₂O thermobarometry, (b) Chlorite + Quartz + H₂O thermometry and (c) Mica + Quartz + H₂O thermobarometry for sample ML2. In (c) each line represents an equilibrium between Mica + Quartz + H₂O as a P–T line along which the interlayer H₂O content of the mica varies. A full description of each method can be found in (Lanari & Duisterhoeft, 2019).

4.9 Whole-Rock Composition

The analyzed Tempa Roccarossa mélange samples all fall in the shale compositional field of Figure 4-9 e, showing very homogeneous SiO₂/Al₂O₃, while Fe₂O₃_{tot}/K₂O is more variable, especially for the two metapelite samples. Although generally in agreement with the composition of the Worldwide Median Pelite (WMP) estimated by (Forshaw & Pattison, 2023), the investigated samples have lower SiO₂ (47.2–58.3 wt.% vs. 63.7 wt.% of WMP; FeO converted to Fe₂O₃_{tot} and sum of oxides normalized to 100 wt.%) and Na₂O (0.66–0.87 vs. 1.37 wt.%), generally higher Fe₂O₃_{tot} (7.47–11.0 vs. 7.55 wt.%), and higher MgO (2.88–3.93 vs. 2.39 wt.%) and CaO (2.08–18.5 vs. 0.65 wt.%), not taking into account sample ML14. The latter is closer to the WMP, that is, 63.0 wt.% SiO₂, 6.96 wt.% Fe₂O₃_{tot}, 0.16 wt.% CaO. Weight loss on ignition (LOI) is generally higher for metapelite (5.83–6.53 wt.%) rather than slate samples (5.21–5.96 wt.%), except for slate sample ML2, which reaches 8.11 wt.%. Compared with the average estimates for upper crustal compositions (UCC; Figure 4-9 f), only a few significant deviations are recognizable, with all samples showing overall comparable patterns. Very slight enrichments are evident for Ti, Al, Fe, Mg, K, Rb, Zn, Cr, Sc, V, Nb and Y, while Na, P, Ba and Co are depleted, well correlating with the generally (potassic) Wm- and Chl-rich and alkali feldspar-poor mineralogy. Slate sample ML2 is strongly enriched in Ca and, less evidently, Sr, and depleted in Si, well in line with the high calcite and low quartz modal abundances revealed by XRPD analyses and with the higher LOI values. On the other hand, metapelite sample ML15 is enriched in Fe (at almost 2xUCC values) and depleted in Cu (coupled with minor depletions in Ca and Sr), well correlating with the relatively high hematite and low plagioclase and calcite contents. Finally, metapelite sample ML14 is strongly depleted

in Ca and Sr, reflecting the absence of calcite, notwithstanding the highest plagioclase modal abundances.

4.10 Chl-Wm Multi-Equilibrium Geothermobarometry

We applied the three multi-equilibrium methods to samples ML1-2-6-14 and SIR1. Except for sample ML6, for which Chl and Wm appear to have not recorded chemical equilibrium, there is no convergence of the selected equilibria for the chlorite + mica + quartz + H₂O method in the P-T space; the results indicate temperature and pressure ranges between 300 and 380 °C and 0.34 and 0.68 GPa, respectively (Figure 4-10 a). Comparable results were obtained by applying the chlorite + quartz + H₂O method for chlorite (Figure 4-10 b) and the mica + quartz + H₂O method for white mica (Figure 4-10 c).

4.11 Strain and Kinematic Vorticity Analyses

To obtain estimations of 3D finite strain, we analysed only calcareous elliptical markers that were deformed passively, excluding rigid objects. We analysed pictures of polished slabs and thin sections, obtaining the R_f/ϕ plots of Figure 4-11 a, e. The 2D finite strain (along RXZ and RYZ planes) varies between 4.50 and 8.26. The ellipsoids fall in the oblate field when plotted in the Ramsay-Flinn diagram (Figure 4-11 f). The latter plot also shows the field of strain ellipsoids estimated in LMB units (Mazzoli, 1992) and mylonites of other southern Apennines units, formed in shear zones at footwall of a major thrust (Pollino-Ciagola unit; Iannace et al., 2007; Vitale & Mazzoli, 2009) and a regional normal detachment (Lungro-Verbicaro unit; Iannace et al., 2007; S. Vitale et al., 2020b). It is worth noting that, in the analysed samples, the maximum stretching (X-axis) always occurs along the direction parallel to the NW-SE tectonic transport.

The kinematic vorticity analysis, performed on the synthetic S-C' structures (R2), indicates similar results for the four analysed slate samples (**Error! Reference source not found.** a, d). The mean θ angles range between 19 and 22°, with a Gaussian distribution, indicating mean kinematic vorticity numbers W_k of 0.92–0.94. On the other hand, the method of rigid porphyroclast analysis (**Error! Reference source not found.** e, h) yields mean values ranging from 0.88 to 0.90. These values, reported in the diagram of **Error! Reference source not found.** i, provide a percentile of the coaxial strain of 20%–35%, that is, the contribution of the

pure shear with respect to the simple shear (e.g., (Means & Park, 1994; Tikoff & Fossen, 1995; Fossen & Cavalcante, 2017).

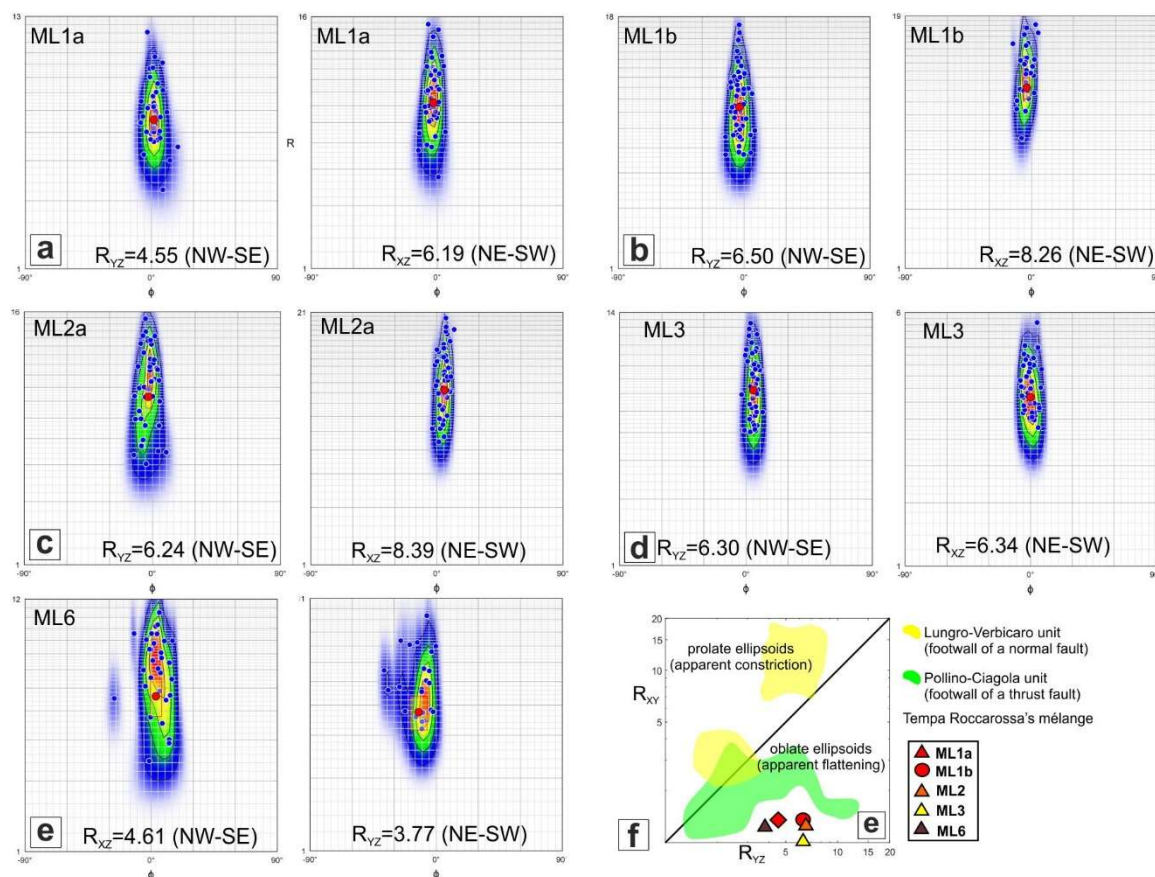


Figure 4-11: (a–e) Rf/φ diagrams and (f) Flinn-Ramsay plot of the investigated Tempa Roccarossa mélange samples. Data from this work and (Mazzoli, 1992; Iannace et al., 2007; Vitale & Mazzoli, 2009).

4.12 Discussion

4.12.1 Petrography, XRPD, Chl-Wm Multi-Equilibrium and Temperature Estimation

By integrating microstructural, mineralogical, mineral-chemical, and whole-rock geochemical analyses, we have investigated the thermal equilibrium at which the studied mélange formed. The mélange is mainly composed of Wm and Chl minerals, which define the relict (S1) and main cleavage (S2). XRPD data show the presence of paragonite and K/Na mica. These minerals occur in the upper part of the late diagenetic zone and are ubiquitous in the anchizone (Kisch, 1987; Leoni et al., 2008; Cavalcante et al., 2012). The temperature information provided by these phases is somewhat controversial. The KI values obtained from the Wm crystals with size $<2 \mu\text{m}$ indicate that the investigated samples seem to have equilibrated at rather low-T conditions, with metapelite pointing to lower grades ($200 \pm 10 \text{ }^\circ\text{C}$, Table 4-1) with

respect to the slate samples (from 220 ± 10 to 250 ± 10 °C). On the other hand, data obtained from empirical geothermometry based on Chl compositions (Figure 4-8 c, d), performed by EDS analysis on larger crystals (5–10 μm thick), not only show a large variability (basically covering the entire spectrum from 125 to >325 °C) but also do not recognize any systematic difference between the two rock types and S1 and S2 cleavages.

A role in these discrepancies could be related to the well-known intrinsic difficulties of low-T Chl geothermometry in general (and the empirical method used in particular), discussed in detail by (Bourdelle & Cathelineau, 2015; Bourdelle, 2021), and (Petroccia et al., 2025). Nevertheless, the large temperature range and the completely overlapping compositions of the crystals aligned along the S1 and S2 foliations could simply indicate the presence of some detrital Chl crystals (e.g., Lanari, 2012; Petroccia et al., 2025). This is particularly evident for the crystals with the highest T values, which fall well within the epizone (or chlorite zone/greenschist facies), and which appear to be at odds with the very fine grain size of the studied Tempa Roccarossa mélange samples. The T estimates obtained from the various multi-equilibrium methods applied are significantly higher than those obtained from the KI. However, since these were obtained using SEM-EDS analyses performed on the larger grains, it seems likely to conclude that the first are detrital in origin, while the second include the neo-formed equilibrium crystals, which thus provide the most reliable geothermobarometric estimates Table 4-1.

Table 4-1: Mineralogical Composition (W %) of Bulk Rock and Clay Fraction.

Mineralogy																
Bulk sample (W%)										Clay fraction (< 2 μm)				KI and Temperatures estimates		
Sample	Ill/Mus	Pg + K/Na mica	Chl + C/S	Qtz	Ant	K-feld	Pl	Cal	Hem	Ill/Mus	Pg	Chl	C/S	FWHM	KI (Basel)	T °C
SIR 1	40.0	3.0	24.0	24.0	1<	1.0	2.0	6.0	0.0	67	9	21	4	0.387	0.310	250 $\pm$ 10
ML 2	36.0	3.0	17.0	14.0	1<	1.0	2.0	27.0	0.0	71	10	14	5	0.428	0.343	240 $\pm$ 10
ML 1	43.0	6.0	23.0	19.0	1<	1.0	2.0	6.0	0.0	80	6	12	2	0.492	0.394	220 $\pm$ 10
ML 14	43.0	tr	23.0	30.0	1<	1.0	3.0	0.0	0.0	81	4	11	4	0.521	0.417	200 $\pm$ 10
ML 15	42.0	tr	23.0	25.0	1<	1.0	3.0	3.0	3.0	79	4	14	3	0.457	0.366	230 $\pm$ 10

However, only detrital Wm crystals have been described in the Monte Facito Formation (G. Ciarapica & L. Passeri, 2000; Passeri, 2010), while Chl crystals have only been found in the Roccarossa mélange (Mazzoli, 1992). It follows that the abundance of Chl and Wm grains in the mélange suggests that they originated from the tectonic erosion of the Paleozoic rocks during the decollement activity. This is a common feature of mélange rocks, including exotic materials (Festa et al., 2019). In this light, the Chl-Wm crystals could be in thermal equilibrium, recording a previous P-T evolution likely associated with the Variscan orogeny, such as observed in metapelite units in Calabria (e.g., Graessner and Schenk, 1999).

The presence of unequilibrated detrital grains is commonly recognized from the observation of zoning patterns, often with the aid of compositional mapping (e.g., Lanari, 2012; Petroccia et al., 2025). However, considering the extremely small grain size of the investigated Chl (as well as Wm) crystals, it is conceivable that the very low-grade conditions reached by the Tempa Roccarossa mélange may have significantly slowed reaction kinetics, possibly allowing P-T re-equilibration only for the finest Wm fraction (size <2 μm).

The trend of increasing KI-estimated T values from the metapelite to the slate samples, which lines up with metapelite samples having lower amounts of paragonite (e.g., Guidotti and Sassi, 1998) and higher LOI values (if the calcite-rich slate sample ML2 is excluded; see Forshaw and Pattison, 2023), could be taken as further evidence for the overall consistency of these T estimates.

In summary, the above observations suggest that T values obtained from Chl geothermometry may be less reliable in the presence of detrital grains than those obtained from KI values. Therefore, the latter will be used in the following discussion to reconstruct the tectono-metamorphic evolution of the Tempa Roccarossa mélange.

4.12.2 Strain, Structures and Kinematics

Strain and kinematic analyses provide significant information on the orogenic evolution of the LMB units. The analyzed meso- and microstructures suggest that the mélange was formed within a shear zone. Hence, these highly deformed rocks can be described as a sheared mélange (*sensu* Needham, 1995). The FSEs, estimated from slate samples (ML1,2,3,6), all located close to the thrust, fall in the apparent oblate field (Figure 4-11 f). However, this flattening may be true or related to a volume loss orthogonal to the shear planes. On the other hand, the results of the kinematic vorticity analysis indicate that a pure shear component is present, ranging

between 20% and 35% (Figure 4-12 i). Therefore, we can hypothesize a further stretching along the Y-axis, that is, orthogonal to the shear direction (X-axis) suggesting that part of the flattening is related to the pure shear component. Although evidence for microstructures associated with a volume decrease, such as the cleavage S1 marked in the calcareous porphyroclasts by stylolites at high angles to the bedding, is poorly represented, we cannot exclude that a volume loss has occurred. This process could also be associated with the water release due to the chemical reactions transforming (dioctahedral) smectite into illite through mixed illite/smectite layers, and subsequently into muscovite white mica (Elliott & Matisoff, 1996; Środoń, 1999; van de Kamp, 2008; Cavalcante et al., 2023), as well as those transforming (trioctahedral) smectite into chlorite through mixed layers chlorite/smectite (e.g., Robinson et al., 2002; Beaufort et al., 2015). This reaction also produces H₂O, silica and other cations, which result in the final growth stages of quartz crystals in the syntaxial veins (Figure 4-5 c and Figure 4-6 j).

The oblate finite strain of the Tempa Roccarossa mélange is like the FSEs calculated for other Apennine units (Figure 4-11 f). Compared to the ellipsoids reconstructed in (Mazzoli, 1992) for the LMB units, including the Roccarossa locality (red area in Figure 4-11 f), our data indicate a more intense deformation toward the oblate field.

Several structures, such as sigma porphyroclasts, synthetic and antithetic S-C' structures (R2 and R1, respectively), S-C structures, late cleavage (S3) and the crenulation lineation L3 mark well the NW-SE shear direction, that is, the X-axis of the FSE. However, the pure shear component was likely characterized by a secondary stretching orthogonal to the shear direction. Accordingly, the ductile shear zone experienced thinning-widening shear, as illustrated in Figure 4-12 j (Tikoff & Fossen, 1999). As a result of the flattening, the mélange rocks are S-tectonites dominated by foliation with no well-marked stretching lineation. The tectonic transport to the E/SE is consistent with the first deformation stage affecting the LMB units as described by (Catalano et al., 2007), and the early fold and thrust verging to the E/SE hosted by Mt. Sirino (Mazzoli, 1992). This orientation is also more generally consistent with the orogenic vergence of the southern Apennines, which ranges from NE to SE (e.g., Ciarcia and Vitale (2025)).

Regarding the kinematic vorticity number estimates, the different W_k values obtained by the two methods, that is, rigid porphyroclasts (W_k = 0.85–90) and the synthetic S-C' structures (0.92–0.94), could be explained considering that the R2 shear planes frequently bound the

porphyroclasts (Figure 4-4 m), reducing the rotation angle, and thus overestimating the pure shear component. However, considering an uncertainty of ± 0.1 (Tikoff & Fossen, 1995; Iacopini et al., 2011), a common interval of 0.85–0.95 is obtained, indicating a percentage of 20%–35% for the pure shear contribution. In any case, the use of different approaches to estimate the kinematic vorticity can reduce the uncertainty of this important tool to evaluate the contribution of coaxial and non-coaxial strain in shear zones (e.g., Xypolias, 2010).

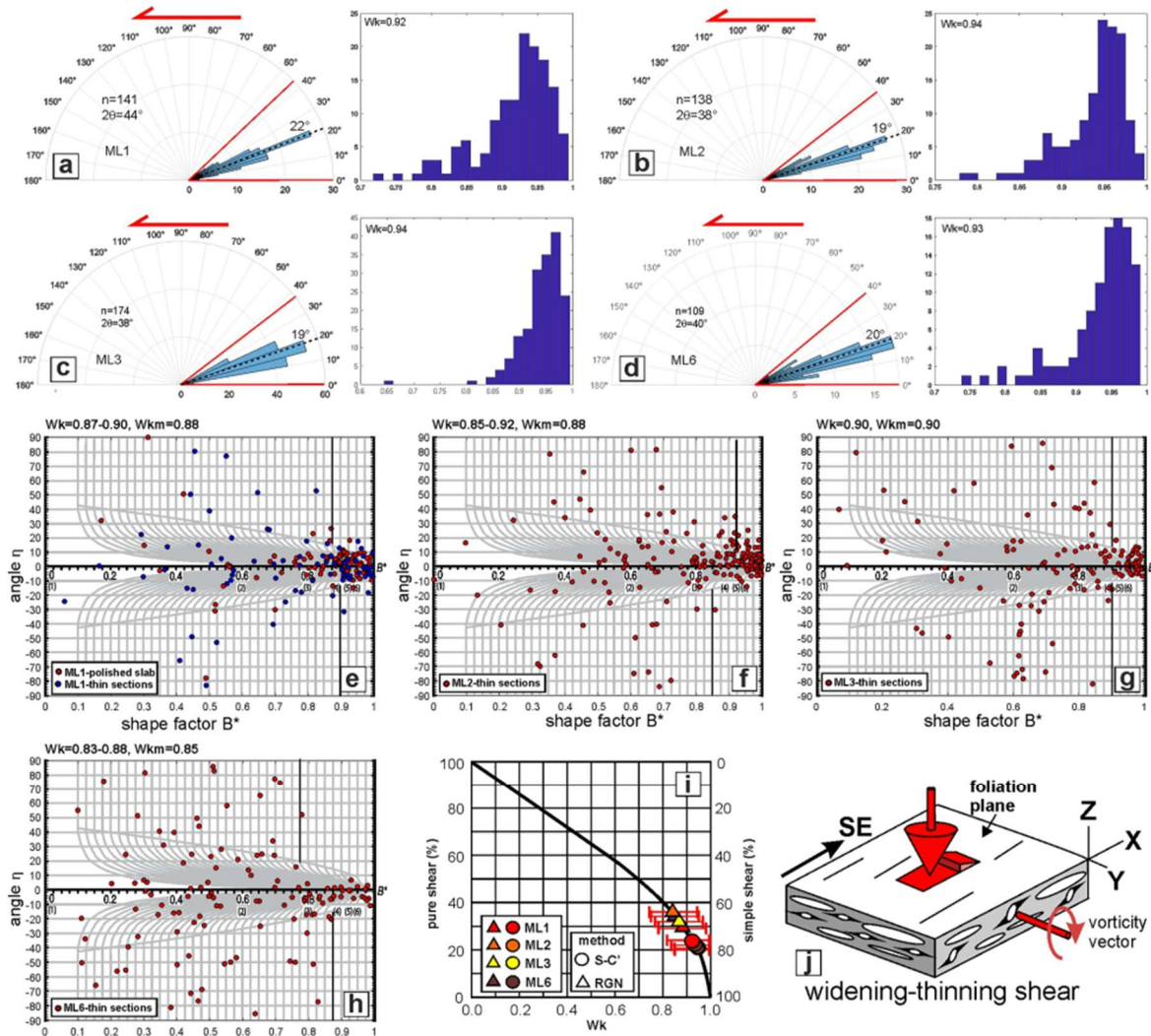


Figure 4-12: Results of the vorticity analyses performed on the investigated Tempa Roccarossa mélange samples. (a–d) S-C' structures method: (a) ML1; (b) ML2; (c) ML3; (d) ML6. (e–h) Rigid porphyroclast method: (e) ML1; (f) ML2; (g) ML3; (h) ML6. (i) Wk versus pure shear/simple shear percentile plot (triangle: S-C' structures method; circle: Rigid porphyroclast method). (j) Model of the widening-thinning shear zone (Tikoff & Fossen, 1999).

Structural analyses of meso- and micro-structures indicate that the Tempa Roccarossa mélange is characterized by three cleavages (S1-S2-S3), with the remnants of the original bedding (S0) occurring only in the large calcareous porphyroclasts (Figure 4-3 b, g). S1 appears as a

disjunctive cleavage preserved in some large porphyroclasts (Figure 4-6 c, e) or in the microlithons of the subsequent crenulation cleavage S2 (Figure 4-12 b and Figure 4-6 b, f). Both S1 and S2 are defined by chlorite and white mica (Figure 4-6 b, f), with chlorite crystals showing no systematic compositional differences associated with each deformation phase. However, the presence of cataclastic relicts around the large porphyroclasts, preserved in the strain shadows (Figure 4-5 h), suggests that the base of the *Monte Facito* Formation was affected by cataclasis prior to ductile strain. In such a tectonic framework, it is hypothesized that part of the chaotic texture results from an early brittle deformation (D0 stage), which was subsequently enhanced by ductile strain (D1-D2-D3 stages). Under this light, the cataclastic rocks associated with the décollement of the LMB units from the autochthonous Paleozoic basement may have acted as a brittle precursor of ductile strain localization, such as observed in other orogenic chains (e.g., Mancktelow & Pennacchioni, 2005; Pennacchioni & Mancktelow, 2007). Further, the cataclastic fabric with displaced clasts of older cataclasite suggests that multiple cataclastic events affected the *mélange* during its early formation.

Further information on the deformation history is provided by the syn-kinematic veins located within the strain shadows of the porphyroclasts, Figure 4-5 b, g). These veins exhibit symmetric antitaxial growth of fibrous calcite (e.g., Bons and Montenari (2005), characterized by sigmoidal or complex shapes, as indicated by the medial line located in the middle of the veins (Figure 4-5 d, f). These geometries can result from the rotation of the strain field or, more likely, from the rotation of the veins (Bons & Montenari, 2005). In this case, vein rotation was a consequence of the porphyroclast rotation during shearing. As typical for brittle-ductile contexts (e.g., Cardello et al., 2024), the veins continued to grow until this syn-kinematic growth ended before the D2 event, as indicated by the deformed veins transposed by the S2 cleavage (Figure 4-5 e). The main S2 foliation is locally deformed by a crenulation with formation of a S3 cleavage and is widely deformed by antithetic and synthetic S-C' structures. S3 cleavage and S-C' structures are likely formed in the same deformation stage (D3).

In summary, the tectonic *mélange* was initially characterized by brittle deformation (D0) and subsequently by a dominant ductile general strain, defined by the interplay between simple shear, pure shear, and volume change, which occurred in three main stages (D1-D2-D3).

4.12.3 Burial and Exhumation Evolution

In this section, we compare our results with those of previous works to shed light on the dynamic evolution of the burial and exhumation of the LMB units during the Cenozoic orogenic construction of the southern Apennines. Unlike (Catalano et al., 2007), who associate the *mélange* formation with the overthrusting of the ApeP onto the LMB, more up-to-date regional geology evidence suggests that the analyzed *mélange* formed in the hanging wall of the major *décollement*. This allowed the detachment of the LMB units from the Palaeozoic succession during the Tortonian-Pliocene orogenic shortening and the following stacking and overthrusting onto the ApuP (Mazzoli, Barkham, et al., 2001; Mazzoli, Zampetti, et al., 2001). This evidence is also supported by the presence of lithoclasts of Fusilinidae-bearing Permian sandstones included in the *mélange* in the Serra Maula (Mt. Facito area; Figure 4-1 a; G. Ciarapica & L. Passeri, 2000) and by Chl and Wm in the analyzed Roccarossa *mélange*. In this sense, the latter, located at the base of the Monte Facito Formation, was progressively formed at the base of the LMB units. It is exposed in several places in the central sector of the chain (e.g., in the Mt. Facito area to the north and in the Mt. Sirino and Mt. Alpi areas to the south), and it has been intercepted in several boreholes (ViDEPI-Project, 2025), the correlation of which indicates that it was formed by a far traveling overthrust along tens of kilometers. The resulting present-day configuration within the LMB units features the *mélange* embedded within the LMB stacked thrust sheets and positioned on top of the ApuP. This scenario is evident at Mt. Alpi, the only place in the southern Apennines where the otherwise buried ApuP succession is exposed, as reported by (Corrado et al., 2002; Mazzoli et al., 2006; La Bruna et al., 2018), and (Cavalcante et al., 2023). The authors highlighted the high strain recorded by the ApuP units, as well as by the upper Messinian clastic deposits located below the tectonic *mélange*. Corrado et al. (2002) analyzed the vitrinite reflectance and clay mineralogy of the upper Messinian clastic units, indicating rapid burial to a depth of approximately 4 km in the Early Pliocene due to the overthrusting of the allochthonous wedge. Mazzoli and Butler (2006) proposed a temperature range of 110–170°C, based on vitrinite reflectance and I/S mixed layers in clayey sediments, and a temperature of 90–170°C, based on fluid inclusions within calcite veins hosted in the Messinian sandstones. I related these temperatures to the maximum burial associated with the overthrusting of the LMB units onto the ApuP. This aligns with the work of Mazzoli et al. (2008), who provided a synthesis of the thermal indicators in the southern Apennines, suggesting that parts of the allochthonous units have been buried to depths >5 km. More recently, Cavalcante et al. (2023) analyzed the clay mineral composition of ApuP, the *mélange*, and the overlying LMB thrust sheet in Mt. Alpi. The results indicated temperatures

of 120–130 °C for the upper Messinian units, corresponding to burial depths of 6–7 km, assuming a geothermal gradient of 20°C/km (Sciamanna et al., 2004; Megna et al., 2014). The Authors also noted that the *mélange* and the overlying LMB unit exhibit slightly higher thermal equilibrium, with temperatures of 130–140 °C. The percentage of illite in the mixed illite/smectite layer determined by different studies (Aldega, 2003; Cavalcante et al., 2023); (Mazzoli et al., 2006; Mazzoli et al., 2008) in the Mt. Alpi area, and temperature estimates determined following (Waliczek et al., 2021).

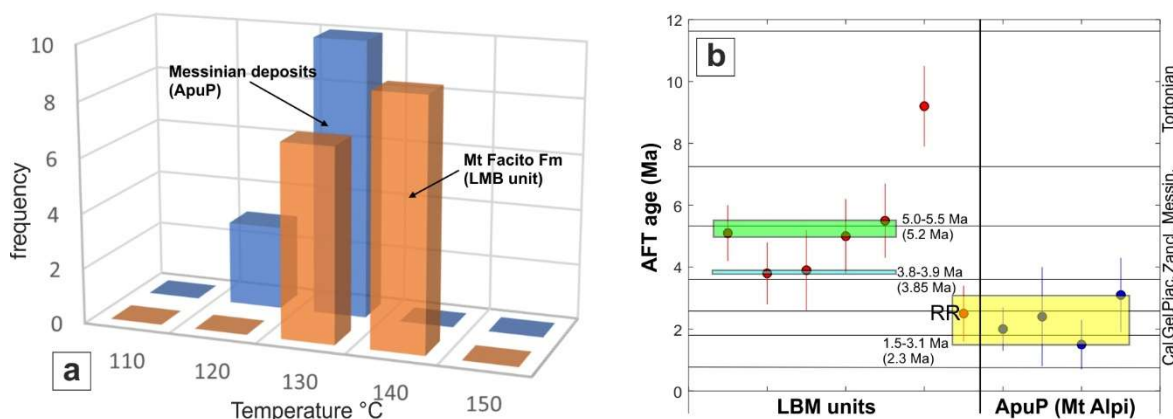


Figure 4-13: (a) Frequency histograms of temperatures calculated by Illite/smectite ratio for the Monte Facito Formation (LMB) and upper Messinian formation of the Apulian Platform (ApuP; Mt. Alpi). Data from (Aldega, 2003; Mazzoli et al., 2008; Cavalcante et al., 2023). (c) Comparison between Apatite Fission Track (AFT) ages of LBM units and ApuP (Mt. Alpi). Data from (Mazzoli et al., 2008).

Despite these regional constraints, few studies have focused on the Tempa Roccarossa succession. Aldega (2003) investigated fluid inclusions in calcite veins hosted in the *Monte Facito* Formation, yielding a temperature range of 130–180 °C. Catalano et al. (2007) determined paleotemperatures using the Kubler Index in samples from the *Monte Facito* Formation, collected at various locations, including the Tempa Roccarossa locality. This analysis suggests a maximum temperature range of 220–250 °C, in good agreement with our estimates, and a maximum burial depth of 8–9 km. Additionally, other rocks of the LMB have been analyzed with the Kubler Index, such as the Jurassic argillites of Mt. Sirino (Mazzoli, Zampetti, et al., 2001), which yield deep diagenetic zone conditions, or the successions in Sasso di Castalda and Pignola-Abriola (Figure 4-1 a; Schiattarella et al. (2006), which also point to the uppermost diagenetic zone.

Our findings further confirm the evidence of low-grade metamorphism recorded by the *mélange* at the base of the LMB unit. Considering that the current geothermal gradient increases from the front of the chain (17–20 °C/km) to the inner sectors (20–26 °C/km) of the

southern Apennines (Sciamanna et al., 2004; Megna et al., 2014) log of Roccarossa01, (ViDEPI-Project, 2025), a gradient of about 20°C/km can be proposed for the early stages of the accretionary wedge growth. Then, the maximum paleotemperatures of 250 ± 10 °C obtained for the slates in Site 1 correspond to a tectonic load between 12 and 13 km (about 0.32–0.35 GPa considering a geobaric gradient of 0.027 GPa/km).

However, in analogy to other shear zones associated with large thrusts (Graham & England, 1976; Mori et al., 2015; Maino et al., 2020; Cardello et al., 2024), the effects of thermal shearing need to be considered when interpreting the thermal evolution and deformation mechanisms of a *mélange*. This is crucial as shear heating can significantly increase local temperatures by at least 50–60 °C (Graham & England, 1976; Mori et al., 2015; Maino et al., 2020; Cardello et al., 2024; Petroccia et al., 2025), thereby affecting recrystallization in shear zones. Therefore, temperatures obtained from mineralogical analyses in high-strain zones may be higher than those predicted solely by the burial depth of the host rock. The effects of shear heating may explain the difference in temperatures estimated for the more deformed slate samples, located close to the thrust fault (Site 1; $T = 250 \pm 10$ °C), and those for the less deformed metapelites, situated further away (Site 2; $T = 200 \pm 10$ °C). In this case, the thermal gradient is 50 °C for a *mélange* thickness of about 25 m, resulting in a value of 2 °C/m.

What is also worth noting is that T estimates obtained from the multi-equilibrium methods do not show any correlation with sample location with respect to the thrust fault. This suggests that the larger (detrital) crystals did not re-equilibrate under the new P - T conditions, notwithstanding the additional heat pulse provided by the shear heating.

Thus, to reconstruct the burial history of the *mélange*, we collected all available geothermal estimates published from the LMB and ApuP successions. The frequency plot of temperatures calculated for the Monte Facito Formation (LMB) and the upper Messinian units (ApuP) (Figure 4-13 a) shows that the Monte Facito Formation reached higher values than the overthrust upper Messinian rocks. This is consistent with the vitrinite reflectance values calculated for the LMB units, which are higher than those values measured in the underlying ApuP succession (Aldega, 2003; Sciamanna et al., 2004; Mazzoli et al., 2008). This implies that the LMB units recorded an earlier phase of burial and heating before piling up and thrusting over the ApuP (Sciamanna et al., 2004; Mazzoli et al., 2006; Mazzoli et al., 2008) highlighted that ApeP units of the Apennine platform, overlying the LMB units, show a lower thermal equilibrium and suggested that the present thickness of the ApeP (generally less than 1 km, as

resulting from geological cross sections; Mazzoli et al. (2008) cannot account for the burial and thermal conditions recorded by the LMB units in the intermediate footwall. Hence, there are paleo-thermal breaks between the LMB units and the overlying ApeP and underlying ApuP rocks. This difference is further amplified by the low-grade metamorphism recorded by the tectonic *mélange*, as evidenced by the clay mineralogy of the Tempa Roccarossa *mélange*.

Information about the exhumation of the LMB units and Mt. Alpi was obtained by apatite fission track analysis (AFT; Figure 4-13 b) by Mazzoli et al., 2008. The Authors suggested that the Tempa Roccarossa rocks recorded a cooling age (above the 110-70°C temperature interval) of about 2.3 Ma, similar to Mt. Alpi. The other AFT ages obtained for the LMB units indicate a cooling age of 9.2 ± 1.0 Ma for the Miocene sedimentary units and two clusters around 5.2 and 3.85 Ma for the Mesozoic rocks. We suggest that the Miocene LMB rocks were off-scraped and detached from the older rocks, and then frontally accreted (e.g., Mattioni et al., 2006), resulting in an older exhumation history. On the contrary, the Mesozoic LMB succession, overthrust by the ApeP units, was buried and exhumed in more recent times.

According to Mazzoli et al., 2006 and Schiattarella et al., 2006; Ciarcia et al., 2011; Giannandrea et al., 2014; Schiattarella et al., 2017, to explain the thermal breaks between the different thrust sheets of the southern Apennines, tectonic delamination by low-angle normal faults (LANFs) should be assumed, which significantly reduced the thickness of allochthonous wedge during the Tortonian- Pliocene time. LANFs are widely reported in the southern Apennines (Ferranti & Oldow, 1999; Casciello et al., 2006; Vitale et al., 2020a; Novellino et al., 2025), including the Cogliandrino fault (Mazzoli et al., 2006) and the area between Mt. Sirino and Mt. Alpi, which puts the LAC units in contact above the LMB thrust sheets. This results in the tectonic elision of the ApeP. According to Mazzoli et al. (2006), the development of LANFs in this region, and in general in the entire southern Apennines (Vitale et al., 2020a; Ciarcia & Vitale, 2025) was associated with the deep-seated thrusting within the ApuP that triggered the tectonic denudation in the Early Pliocene.

4.12.4 Tortonian-Quaternary Evolution of the Tempa Roccarossa *Mélange* and Mt. Alpi

Finally, we summarize all the above-mentioned observations in the cartoon in Figure 4-14, which illustrates the proposed Tortonian-Quaternary evolution of the Tempa Roccarossa *mélange* and Mt. Alpi within the framework of the orogenic development of the southern Apennines.

To estimate depth values from thermal data, we assumed an average geothermal gradient of 20°C/km and an average thickness of the original LMB unit of about 3 km (Figure 4-2 e).

The first tectonic stage occurred in the Tortonian—early Messinian interval, when the ApuP units thrust over the LMB units (T0, Figure 4-14 a), the latter detached from the Paleozoic basement (T1; Figure 4-14 a). The concurrent brittle deformation and increasing burial allowed the LMB units to transition from brittle (D0) to ductile deformation (D1-D2-D3). In this scenario, the orogenic evolution can be explained by the simultaneous activity of thrusting stages T0 and T1, with regional subsidence and thrusting (T0) increasing burial and the décollement (T1), allowing the LMB units to detach from the Paleozoic basement.

After reaching the maximum burial depth, the tectonic exhumation began with the piling up of LMB and the thrusting (T2) of the whole allochthonous wedge onto the ApuP succession. The latter was buried up to 7 km (Figure 4-14 b) according to the geothermometric estimates (120–130 °C; Mazzoli et al., 2006; Cavalcante et al., 2023). In our scheme, the Tempa Roccarossa mélange was buried more than the Mt. Alpi mélange, with a maximum difference of 75 °C (considering shear heating of 50 °C), corresponding to a burial difference of about 3 km.

As discussed earlier, the comparison of burial temperature estimates obtained for the Monte Facito Formation of the LMB with those for the upper Messinian sedimentary units of the ApuP (Figure 4-13 a) suggests that the LMB units were more deeply buried than the ApuP prior to thrusting. The tectonic exhumation of the allochthonous wedge is recorded by the AFT cluster ages of 5–5.5 Ma of the LMB units (Figure 4-13 b), corresponding to the overthrust onto the ApuP.

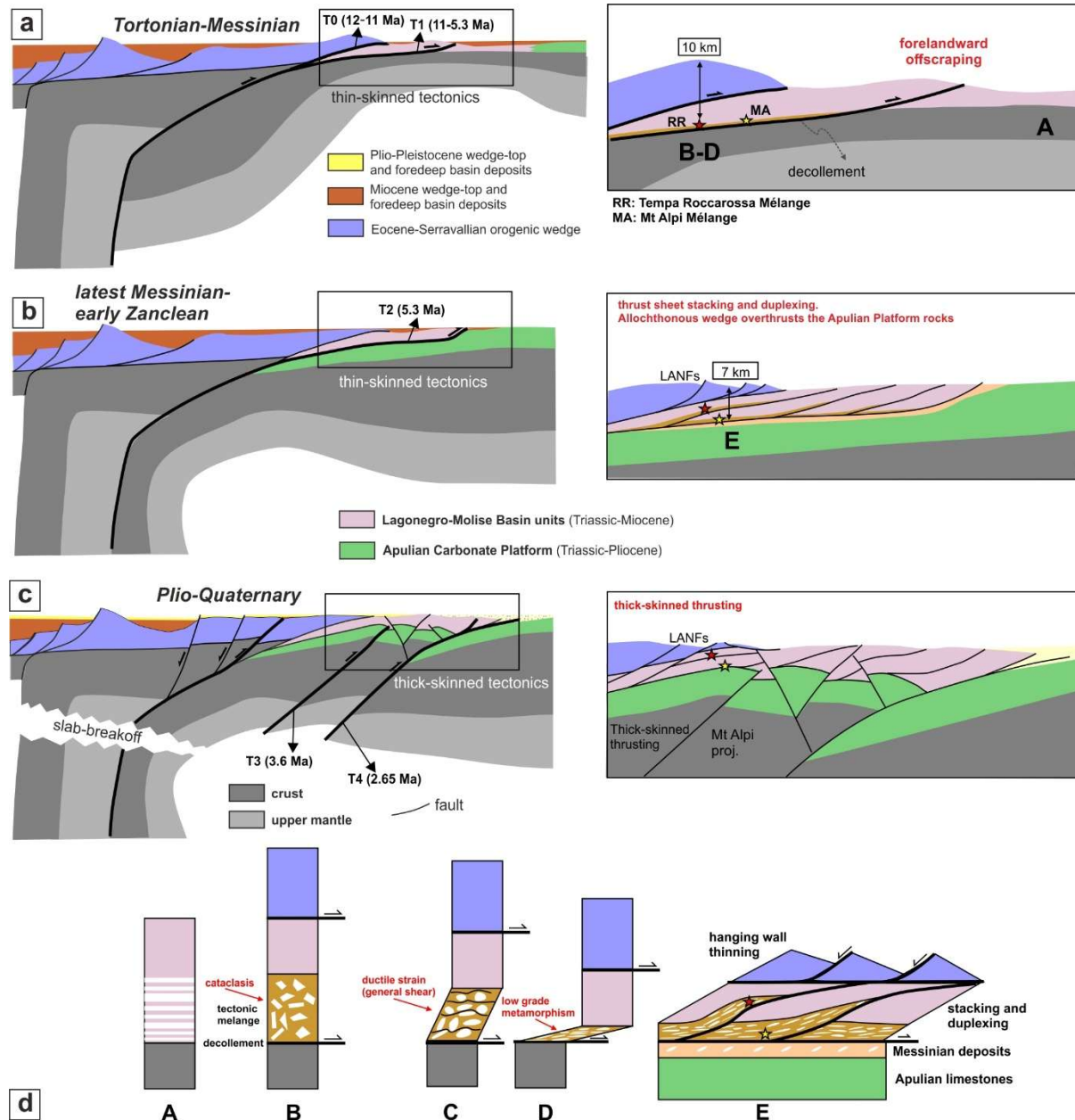


Figure 4-14: (a–c) Schematic crustal section showing the proposed tectonic evolution from Tortonian to Quaternary for the southern Apennines-Calabria-Peloritani System (modified after (Vitale et al., 2019; Vitale et al., 2019b); not to vertical scale). (d) Cartoon showing the tectonic and strain evolution of the tectonic mélangé units (modified after Kimura et al. (2012). LANFs: low-angle normal faults.

We suggest that exhumation and unloading were enhanced by tectonic delamination in the hanging wall by LANFs during the latest Messinian-Early Pliocene stage of the orogenic evolution, which may correspond to the enhancement of erosion in the Mediterranean catchments related to the salinity crisis (e.g., Garcia-Castellanos and Villaseñor (2011)). Finally, in the Plio-Quaternary (Figure 4-14 c), the allochthon wedge was further exhumed by the activity of two main thrusts (T3 and T4), which were likely active in the 3.9–3.8 Ma and 3.1–1.5 Ma time intervals, respectively, as indicated by the AFT clustered ages (Figure 4-13 b). In

this case, the unloading was due to the activity of LANFs (e.g., Mazzoli et al. (2006), likely associated with out-of-sequence thrusts rooted in the buried Apulian succession (e.g., Vitale et al., 2020a; Ciarcia & Vitale, 2025).

Figure 4-14 d outlines the strain and tectonic evolution of the Roccarossa mélange. Stage A represents the original succession of the LMB attached to the Variscan basement. Stage B represents the formation of cataclastic rocks associated with the décollement activity between the LMB and the Variscan basement. The brittle deformation likely formed a clayey matrix-supported cataclasite. Subsequently, this texture was affected by ductile strain and low-grade metamorphism (Stages C-D), reaching temperatures of about 250 ± 10 °C for the highly deformed slates near the principal slip surface of the thrust, which may be partly (50 °C) related to the effects of shear heating. Finally (Stage F), LMB thrust sheets piled up by stacking and duplexing, overthrusting onto the ApuP. Meanwhile, LANFs affect the ApuP units, located at the hanging wall of the LMB units, decreasing the thickness and the lithostatic load.

To better explain the thermobaric evolution of the Tempa Roccarossa mélange and the upper Messinian deposits (ApuP) in Mt. Alpi, we reconstructed their P-T-t paths (Figure 4-15), incorporating all available information, including the results of our study. The base of the Monte Facito Formation in the Tempa Roccarossa area was overthrust (T0) by allochthonous units in the early Tortonian, reaching about 250 ± 10 °C. These thermal values correspond to a maximum depth of 12–13 km; however, considering the possible contribution of the shear heating of about 50°C (Maino et al., 2020; Cardello et al., 2024), a depth of 10 ± 0.5 km can be proposed. Synchronously, the succession was detached from the Paleozoic units and subsequently piled up and exhumed onto other LMB units (T1).

To better constrain the orogenic pulses, we have added some bars to the scheme of Figure 4-15, indicating the ages of the foredeep deposits for the different tectonic units (e.g., Vitale and Ciarcia (2018), including the *Serra Palazzo* Formation (Frigento Unit, LMB), the *San Giorgio* Formation (Sannio Unit, LMB), the *Torrente Flaga* Formation (Fortore Unit, LMB), the inner and outer buried ApuP, and finally, the Bradanic foredeep sediments (ApuP).

In the early Zanclean (earliest Pliocene), the allochthon was thrust over the ApuP (T2), burying the Mt. Alpi succession (MA) to a depth of about 5–7 km (120–130°C; Mazzoli et al., 2006; Cavalcante et al., 2023). It should be emphasized that the vertical distance between the Tempa Roccarossa mélange and the ApuP in the *Tempa Roccarossa I* well (Figure 4-2 e) is currently about 3 km. However, the orogenic position of Tempa Roccarossa is more internal

than that of Mt. Alpi, making it reasonable to assume that the ApuP below the *mélange* was deeper buried than Mt. Alpi.

Finally, both the Tempa Roccarossa and Mt. Alpi successions were exhumed during the T3 and T4 thrusting events, which crossed the 110–70 °C AFT isotherm in the 3.1–1.5 Ma time interval (Figure 4-13 b). Regarding the three ductile deformation phases recorded by the tectonic *mélange* that allowed the formation of the three cleavages (e.g., Figure 4-6 f), we suggest that the D1-D2 stages occurred under low-grade metamorphism at 10 km depth. In contrast, the D3 deformation occurred during the subsequent thrusting and exhumation.

4.13 Conclusions

The present study contributes to the understanding of tectonic *mélanges*, providing an example from the southern Apennines. This work examines the Late Miocene-Early Pleistocene orogenic dynamics of the area in terms of the relationships between tectonics and burial and exhumation history. We examined the tectonic *mélange* that initiated this deformation by detaching the Meso-Cenozoic pelagic succession from the autochthonous Paleozoic basement. Meso- and microstructural analysis, combined with strain and kinematic investigations, and the results of petrographic, mineralogical, mineral-chemical, and whole-rock geochemical analyses, allowed us to reconstruct the P-T-t path for these rock units.

The main result of our study is the recognition of a ductile deformation consistent with the estimated temperature peak of 250 ± 10 °C obtained from the Kubler Index on the fines fraction (size <2 mm) of Wm. However, considering a 50 °C contribution from shear heating, the burial depth is estimated to be 10 km (about 0.3 GPa), assuming an average paleogeothermal gradient of 20°C/km. Furthermore, the thermobaric evolution of the Tempa Roccarossa *mélange* and the upper Messinian ApuP units in Mt. Alpi was reconstructed. The analyzed *mélange* is located at the base of the LMB unit, overriding another LMB thrust sheet. A similar *mélange* is exposed near Mt. Alpi, at the top of the ApuP succession, below the LMB unit, and characterized by lower thermal equilibrium. Both *mélange* units began to form at the same décollement level at the base of the *Monte Facito* Formation. However, the structural position of the Tempa Roccarossa *mélange*, along with its higher thermal equilibrium, suggests that it was in a more internal position and subjected to a higher lithostatic load compared to the Mt. Alpi *mélange*. Further, a thermal break is also identified between the LMB units and the Apulian succession of Mt. Alpi, suggesting that the LMB units were buried to greater depths

prior to orogenic coupling with the Apulian succession. This evidence suggests that the LMB units were exhumed and unloaded before being thrust over the ApuP succession. The unloading of about 3 km probably occurred due to the activity of low-angle normal faults.

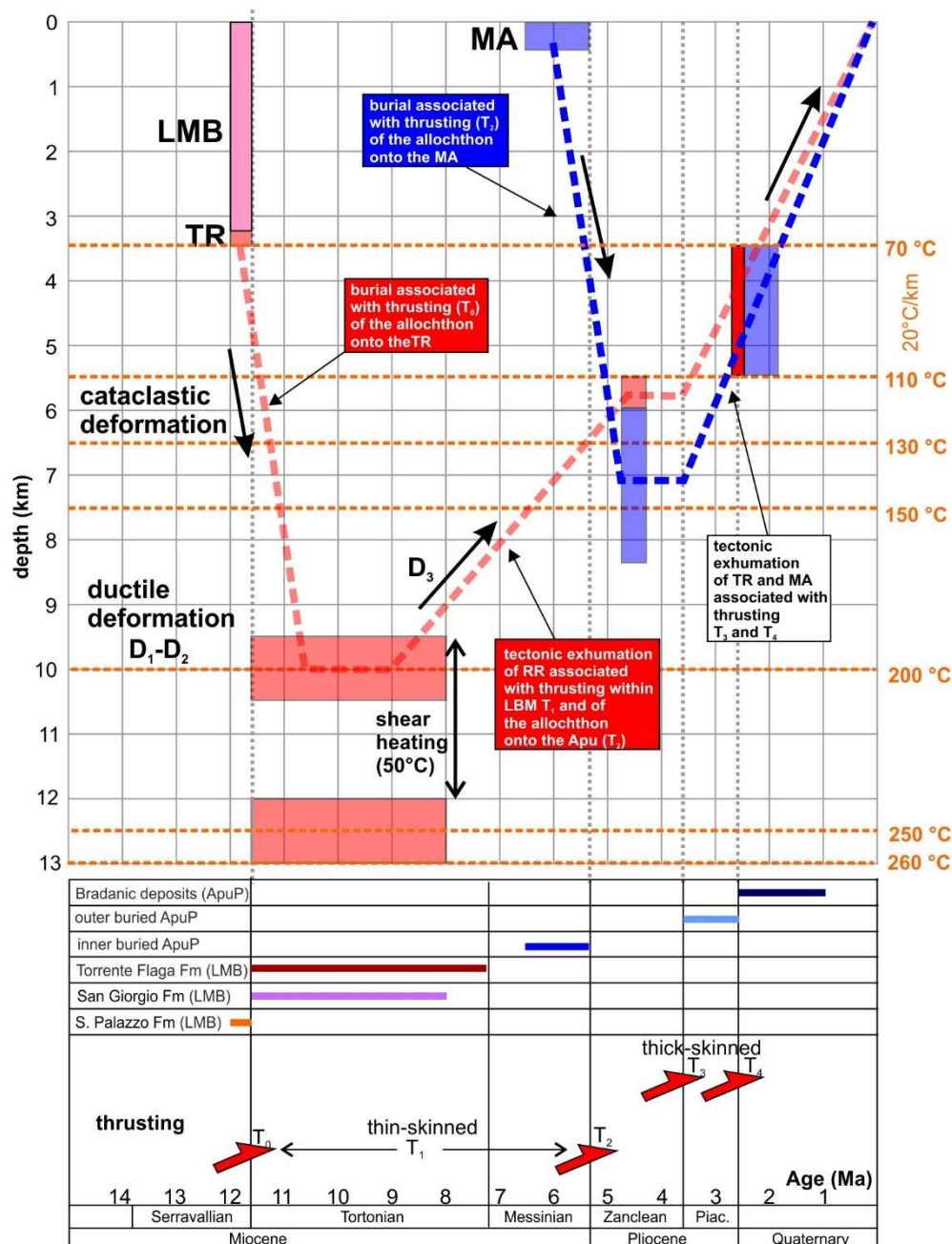


Figure 4-15: Reconstructed P-T-t paths for the Tempa Roccarossa (TR) mélangé and Mt. Alpi upper Messinian units (MA).

Microstructural analysis indicates that the mélangé initially formed under brittle conditions, with the development of a cataclasite (stage D0), and subsequently under ductile conditions (stages D1-D2-D3), leading to the formation of an early disjunctive cleavage (S1), which was later transposed by a main crenulation cleavage (S2). Both foliations consist of thin Wm and

Chl assemblages. A third cleavage (S3) is weakly developed. All meso- and micro- structures indicate a tectonic vergence toward the E/SE. The strain and kinematic vorticity analyses provide valuable insights into the type of 3D deformation and the role of coaxial strain. The calculated finite strain ellipsoids are oblate, with the maximum stretching (X -axis) parallel to the shear direction. The estimated percentage of pure shear is in the range of 20%–35%.

These results provide a comprehensive overview of the tectonic, burial, and exhumation history of the Tempa Roccarossa *mélange* and associated units. They emphasize the significant role of brittle deformation, which acted as a precursor to the subsequent ductile strain and unloading processes, such as near-surface low-angle normal faulting during in- and out-of-sequence thrusting pulses. This accounts for the thermal breaks between the different thrust sheets that form the southern Apennines fold-and-thrust belt. Finally, this work marks the importance of a detailed strain and kinematic analysis to fully characterize the tectonic *mélanges* and their evolution and a petrologic and mineralogic multi-approach to discriminate the role of detrital minerals from ones in equilibrium with the new P-T conditions, especially in low-grade metamorphic rocks, such as one analysed in this wor

Chapter 5 Ligurian Accretionary Complex

5.1 Introduction

The Ligurian Units in the southern Apennines are a key component in understanding the geodynamic evolution of the central Mediterranean region (Figure 5-1). These units, primarily consisting of oceanic and transitional crust, have experienced a complex tectono-metamorphic history driven by the convergence and subsequent collision of the African and European plates. Despite extensive research (Bonardi et al., 1988a; Vitale et al., 2011; Ciarcia et al., 2012; Vitale et al., 2013a; Vitale et al., 2019), several critical questions remain unresolved, particularly in relation to the deformation mechanisms, strain styles, and the thermal and fluid conditions during orogenic burial.

The aim of this study is to dissect the deformation history of the Ligurian Units exposed along the Cilento coastline of southern Italy. Through a multi-disciplinary approach, we will focus on the following specific objectives:

1. To analyse the deformation history of the Ligurian Units in the Cilento region, combining structural survey, strain analysis, and microstructural investigations.
2. To assess the temperature and pressure conditions during key deformation phases, integrating thermobarometric data with structural observations to clarify the tectonic evolution.
3. To unravel the correlation between deformation events by focusing on the structural relationships across different outcrops, shedding light on the tectonic processes that shaped the Ligurian Units.
4. To contribute to the broader understanding of the Apennine orogenic system by integrating local data with regional models of tectonic evolution, providing new insights into the formation of the oldest accretionary prism.

By analysing key outcrops from Pioppi, Punta Telegrafo, and Caprioli, this study aims to provide new, high-resolution data on the deformation, metamorphism, and tectonic history of the Ligurian Units, ultimately refining our understanding of their role within the southern Apennine orogen.

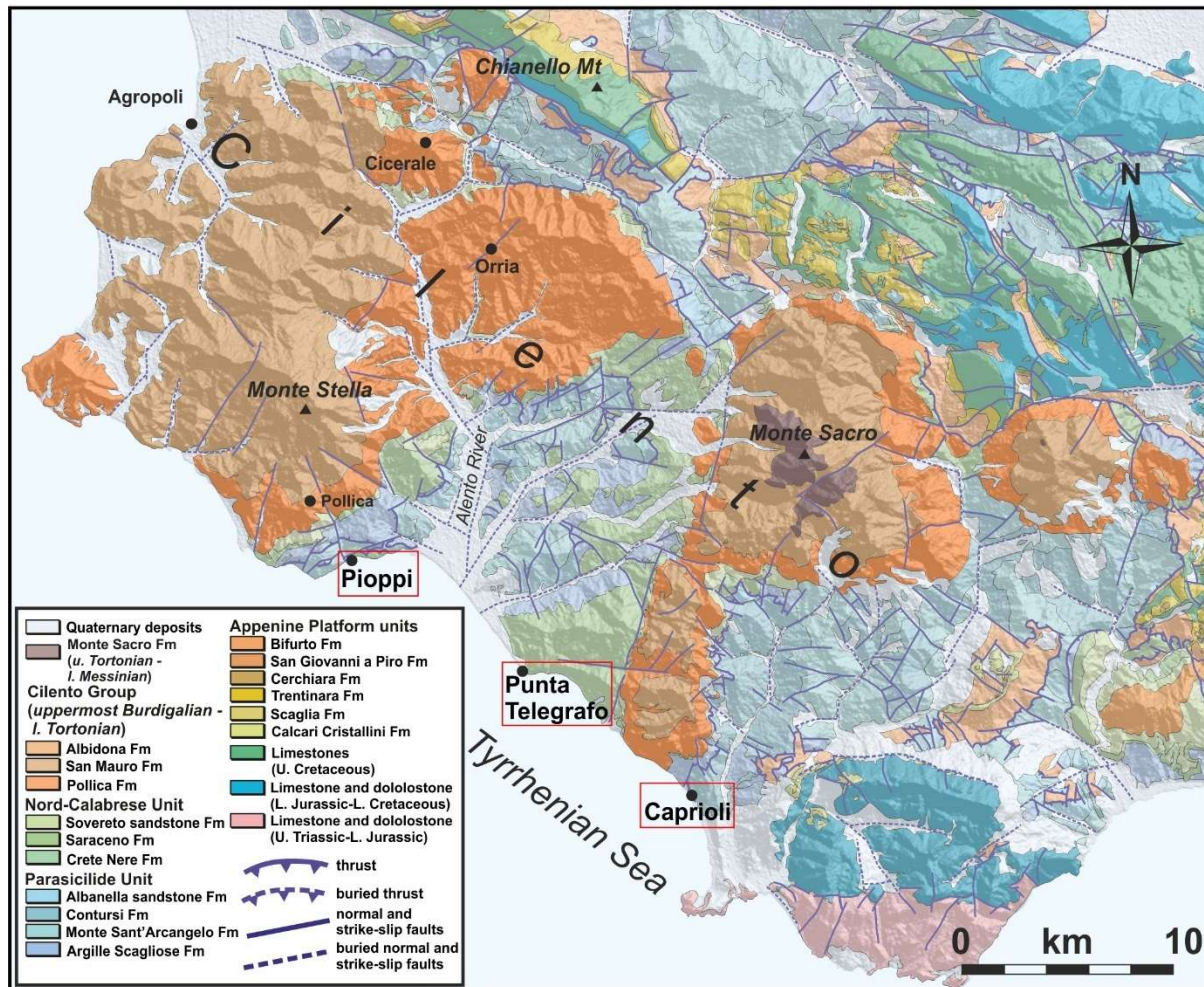


Figure 5-1: Geological map of Cilento with the three study areas indicated.

5.2 Geological settings of the study areas

5.2.1 Pioppi area

At this analysed site, deposits of the Crete Nere Formation of the Nord-Calabrese Unit are present, and, only exceptionally, some sediments typical of the M.S. Arcangelo Formation of the Parasicilide Unit are also present. The Crete Nere Formation is exposed only in its upper part, near the transition to the Saraceno Formation, and consists of alternations of lamellar clays, generally grey and lead-grey, sometimes greenish and hazelnut-coloured, with intervals of thin and medium blackish layers. In the upper part, thin and medium intercalations of siliciclastic and carbonate turbidites are increasingly frequent. The Parasicilide Unit (Bonardi et al., 1988a; Ciarcia et al., 2009a; Vitale et al., 2011; Vitale et al., 2019a), tectonically underlaying the Nord-Calabrese Unit, is present in this area, only with rare marls and

limestones of the Monte Sant'Arcangelo Fm, Upper Eocene -Aquitania p.p. in age (Ciarcia et al., 2009a).

The succession is clearly visible along the western side of the Pioppi coastline and shows substantial continuity of exposure (Figure 5-2 a, b). In the easternmost portion, deposits related to the Crete Nere Formation are present, characterised by their dark, clayey facies. Further along, intercalations of thin layers of predominantly carbonate turbidites become more frequent. The thicknesses of the succession are difficult to measure due to the intense deformation; however, several tens of meters are certainly exposed.

5.2.2 Punta Telegrafo

At this site, deposits exclusively referable to the Saraceno Formation are observed, which consists of its lower member (Punta Telegrafo member), predominantly calcareous turbidite, subordinately marly, and with blackish pelitic intercalations. The layers are generally tabular and medium to thick, but locally thin or very thick. The carbonate turbidites are whitish and greyish, medium to fine, sometimes coarse, also with bands and nodules of dark chert in the lower part of the formation; the pelites are lamellar clays, generally blackish, sometimes bluish-grey and greenish grey; the arenite/pelite ratio is generally greater than 1. The composition of carbonate sandstones varies from calcarenites to hybrid sandstones, the composition of sandstones varies from quartzolitic to lithoarenites and feldspathic lithoarenites (Critelli, 1991;1993). Towards the top, the carbonate fraction decreases and the pelitic fraction increases. Frequent veins of calcite and quartz are observed.

The most interesting outcrops are found along the coast, south of the town of Ascea, at Punta del Telegrafo, where they form a small promontory (Figure 5-2 c, d). The exposed succession, consisting almost exclusively of carbonate turbidites interbedded with blackish shales, is intensely deformed and folded, so much so that a reference thickness cannot be estimated, although it is certainly several tens of meters.

Caprioli, similar to the Pioppi site, features deposits from the Crete Nere Formation of the Northern Calabrian Unit, with its upper portion near the transition to the Saraceno Formation. The sedimentary succession here consists of alternating layers of lamellar clays, generally gray and lead-gray, occasionally greenish and hazelnut-colored, interspersed with fine- and medium-grained blackish layers of siliciclastic and calciclastic turbidites.

The outcrops are well exposed along the coastal stretch from the end of Saline beach to almost Pisciotta Marina (Figure 5-2 e, f). Toward the top of the succession, thin and medium-grained interbeds of carbonate and siliciclastic turbidites become more abundant. The latter, in the northwestern sector, occur in medium and coarse-grained layers and reveal flute casts and ichnofacies at the base of the layers, sometimes overturned. The thicknesses of the succession are difficult to measure due to the intense deformation, but even here, several tens of meters are certainly exposed.



Figure 5-2: Pioppi area: (a) outcrop of black shales (Crete Nere Fm); (b) Convoluted layers within a calcareous turbidite (passage between Crete Nere Fm and Saraceno Fm). Punta Telegrafo area: (c) panoramic view of the site showing calcareous turbidites (lower part of Saraceno Fm); (d) pelitic layers within the calcareous succession. Caprioli area: (e) black shales and arenaceous layers (Crete Nere Fm); (f) flute casts in overturned arenaceous beds.

5.3 Pioppi area

5.3.1 Mesostructures

5.3.1.1 Folds

Different deformation structures have been recognised in the study area, including two sets of folds (F1 and F2), frequently associated with thrust faults. Some shear zones strongly deform the whole succession, including a major structure (Pioppi Shear Zone). Figure 5-3 a shows various examples of F1 folds, generally characterised by isoclinal to tight geometries (Figure 5-3 a, d) and a fold hinge thickening of the more pelitic layers (Figure 5-3 c, d). An axial plane cleavage (S1) is developed (Figure 5-3 a). An intense boudinage is observed in the high deformation zones, such as the overturned limbs of major F1 folds (Figure 5-3 e, g). In some cases, these shear zones show unrooted isoclinal F1 folds and boudins (Figure 5-3 e). In some cases, the boudins are shortened by the successive deformation phase (Figure 5-3 g). F1 folds are, in turn, folded by the F2 folds forming an interference pattern (Figure 5-3 a, c) of the 3-type or between 2 and 3-type of the Ramsay's classification.

F2 folds (Figure 5-4 a-f) show isoclinal to open geometries compared to the F1 ones. F2 folds show rounded geometries (Figure 5-4 a) or angular with rounded hinge (Figure 5-4 b). A S2 crenulation cleavage deforming a previous disjunctive S1 cleavage is well-developed (Figure 5-4 c). Some F2 show a box shape (Figure 5-4 d) or are associated with thrust faults (Figure 5-4 e). Finally, frequently F2 folds deform F1 structures (Figure 5-4 e).

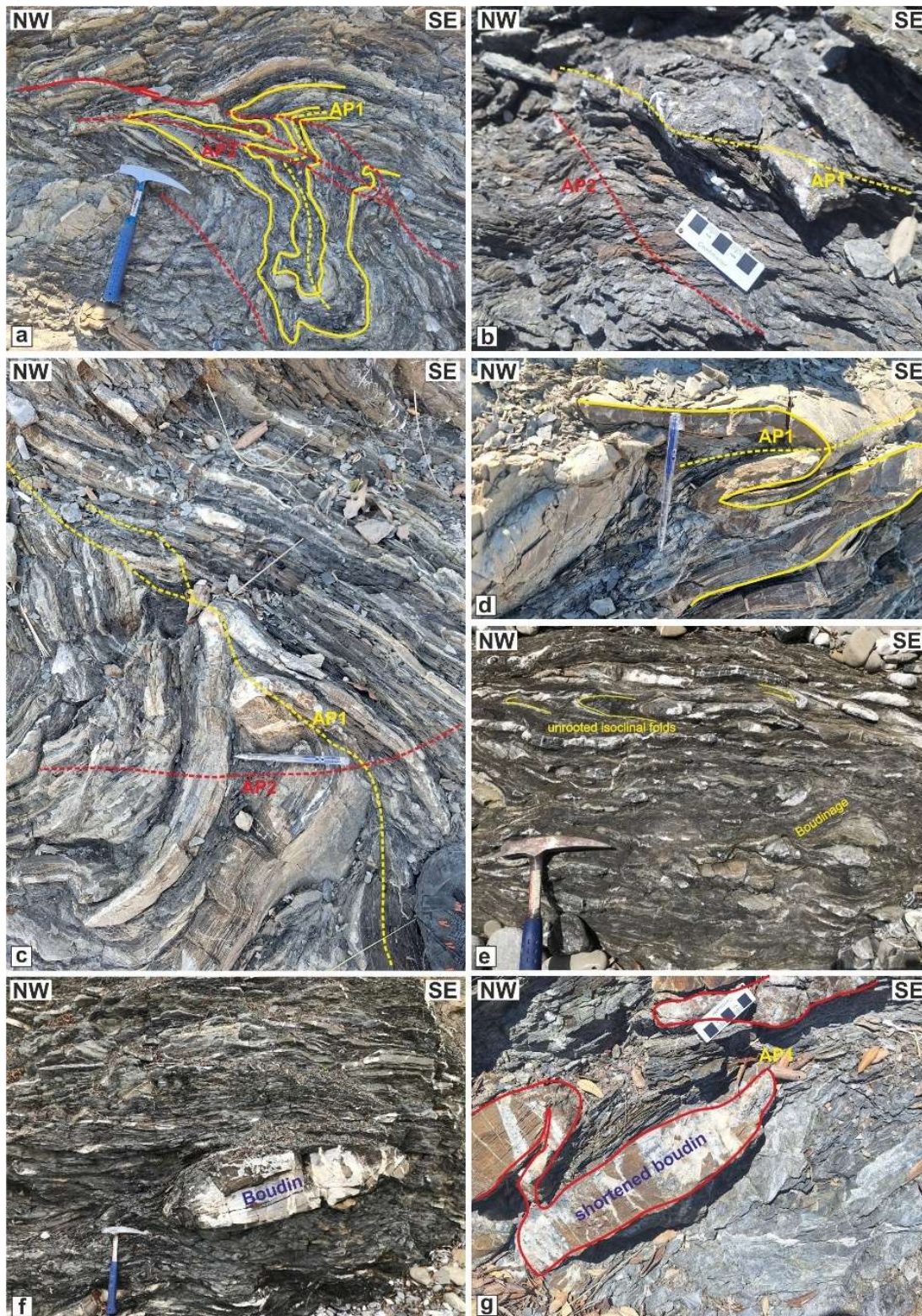


Figure 5-3: (a-b) Isoclinal F1 fold deformed by a F2 fold forming an interference pattern of type 3. (c) Angular and tight F1 fold deformed by an F2 fold. (d) fl isoclinal folds. (e-f) Highly deformed rocks showing unrooted F1 folds and boudins. (g) A shortened boudin.

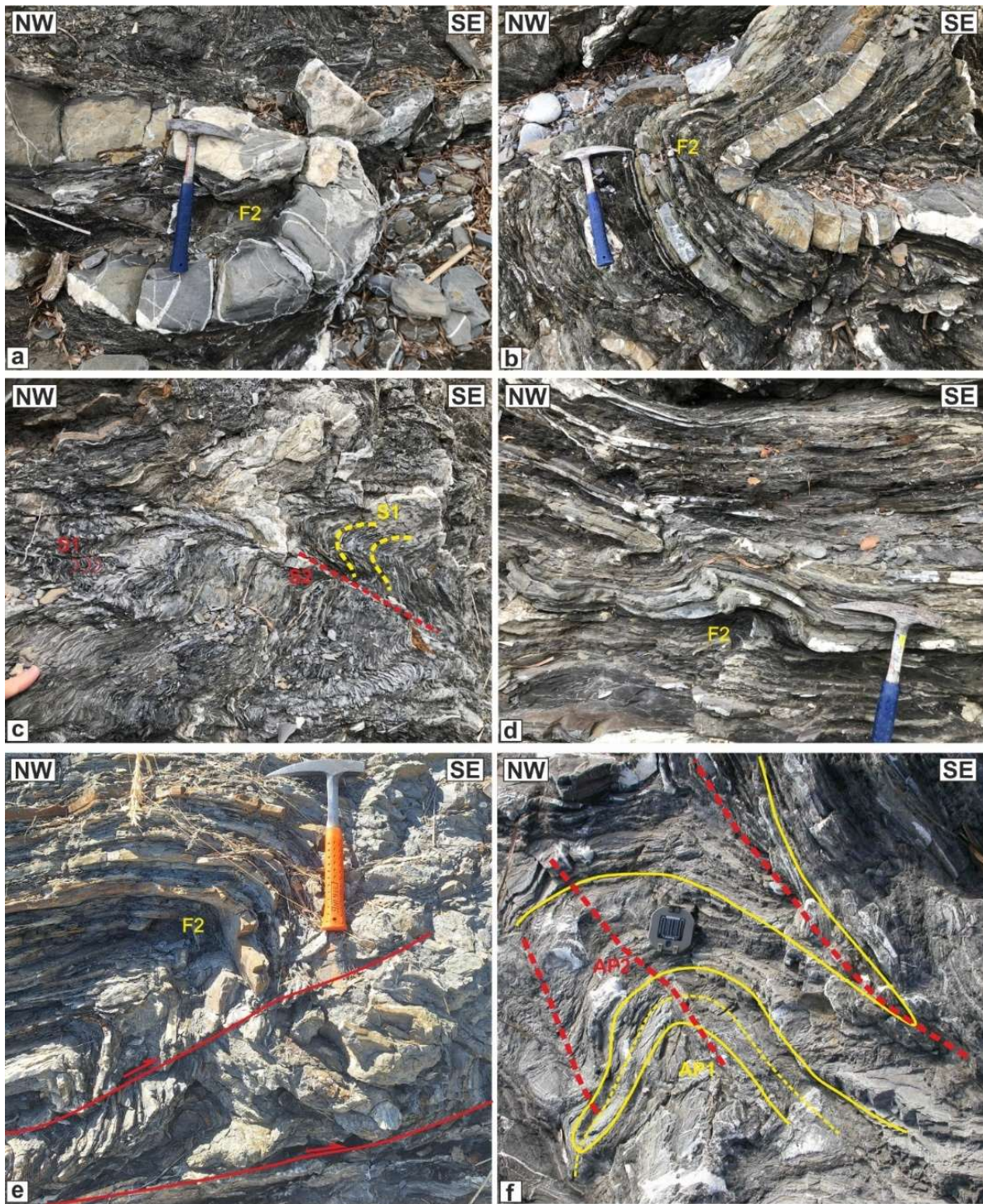


Figure 5-4: (a) Calcareous layer isoclinally folded (F2 fold). (b) Angular F2 fold with a rounded hinge and a subhorizontal axial plane. (c) S2 crenulation cleavage deforming a previous disjunctive S1 cleavage. (d) F2 box fold. (e) F2 fold associated with thrust faults. (f) Isoclinal F1 folds deformed by a F2 fold.

5.3.1.2 Faults

Several faults characterise the succession exposed at Pioppi, most of which show reverse kinematics. Figure 5-5 a illustrates a well-developed thrust fault cutting through a layered succession made of black shales and arenaceous layers. It is worth noting the drag folds formed in the hanging walls (antiformal) and footwalls (synformal). These minor thrust faults show a ramp-flat geometry, with the flat planes parallel to the bedding. Generally, these reverse faults show a displacement of a few cm (Figure 5-5 b, c). Locally, some thrust faults exhibit a displacement of a few meters (Figure 5-5 c), somewhere crosscutting previous reverse faults (Figure 5-5 c), frequently accompanied by hanging wall antiforms and footwall synforms (Figure 5-5 a, c, d, f). Some structures are pre-buckle thrusts, especially those located at the hinge of F1 and F2 folds (Figure 5-5 e). Finally, S-C structures are present (Figure 5-5 f) associated with major thrust faults.

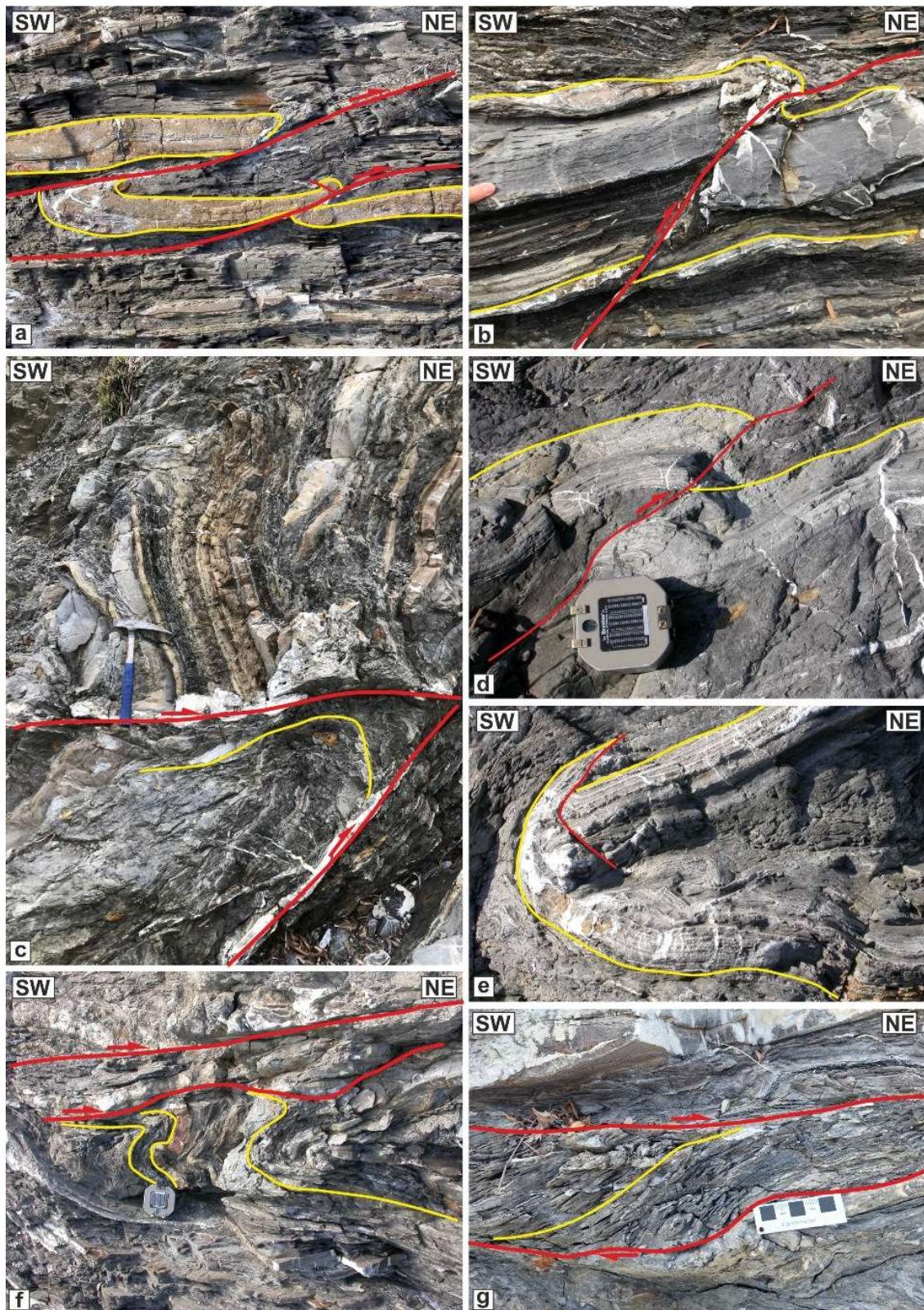


Figure 5-5: (a) minor thrust faults dislocating an arenaceous layer, forming hanging wall antiforms and footwall synforms; (b) minor thrust fault deforming a calcareous layer; (c) a major sub-horizontal thrust fault crosscutting a previous high-angle reverse fault with associated hanging wall antiform; (d) minor thrust fault deforming a calcareous layer; (e) pre-buckle thrust in the fold hinge; (f) sub-horizontal thrust faults with associated footwall folds; (g) S-C structures.

5.3.2 Pioppi Shear Zone

The succession exposed at Pioppi hosts some ductile shear zones. One of these shows a few meters of thickness (Figure 5-6 a) cutting F2 folds. Due to its structural importance, this structure was designated as the Pioppi Shear Zone (PSZ). Only the basal shear plane is exposed (Figure 5-6 a). Generally, the PSZ hosts a chaotic assemblage of porphyroclasts embedded in a slate matrix (Figure 5-6 b-f), giving the outcrop a *mélange* appearance. The highly deformed rocks are affected by Riedle (R1) and anti-Riedle (R2) shears (Figure 5-6 b-f), indicating a top-to-the NE shear direction. Porphyroclasts are composed of limestones and sandstones, typically exhibiting a phi-shape.

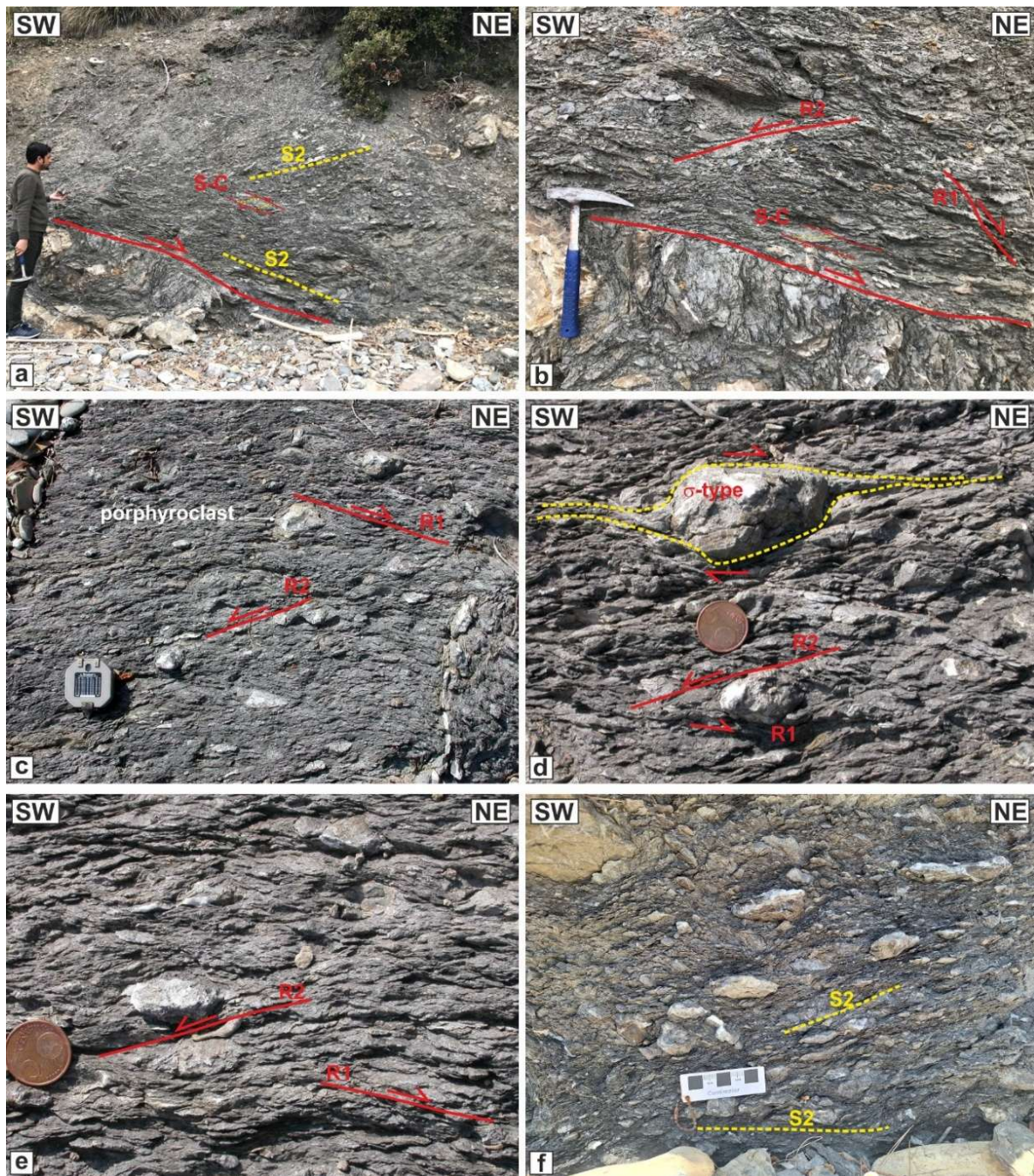


Figure 5-6: Pioppi Shear Zone. (a) Panoramic view of the PSZ. (b-f) Particulars of the PSZ showing a well-developed foliation (S2), R1-R2 structures and several calcareous and arenaceous porphyroclasts.

5.3.3 Stereographic Projections

Generally, the S1 cleavage is parallel to the bedding S0. The stereographic projection (lower hemisphere, equi-area net) of S0/S1 poles to planes (Figure 5-7 a) indicates a girdle distribution with a theoretical fold axis (π_1) of 311/14. A similar girdle distribution, but more dispersed, characterizes the F1 fold axis (Figure 5-7 b) with a theoretical fold axis (π_2) of 148/25. Poles to F1 folds (Figure 5-7 c) indicate again a girdle distribution with a theoretical fold axis (π_3) of 300/35. F2 fold axes (Figure 5-7 d) show a cluster distribution with a mean value of 308/24. Poles to F2 fold axial planes (Figure 5-7 e) indicate a girdle distribution with two clusters indicating mean planes of 245/32 and 002/39. Finally, the R2 (Figure 5-7 f) and R1 (Figure 5-7 g) shear planes, projected as cyclographs with indicated slip vectors, suggest a NE-SW shear direction, with a mean shear top-to-the NE.

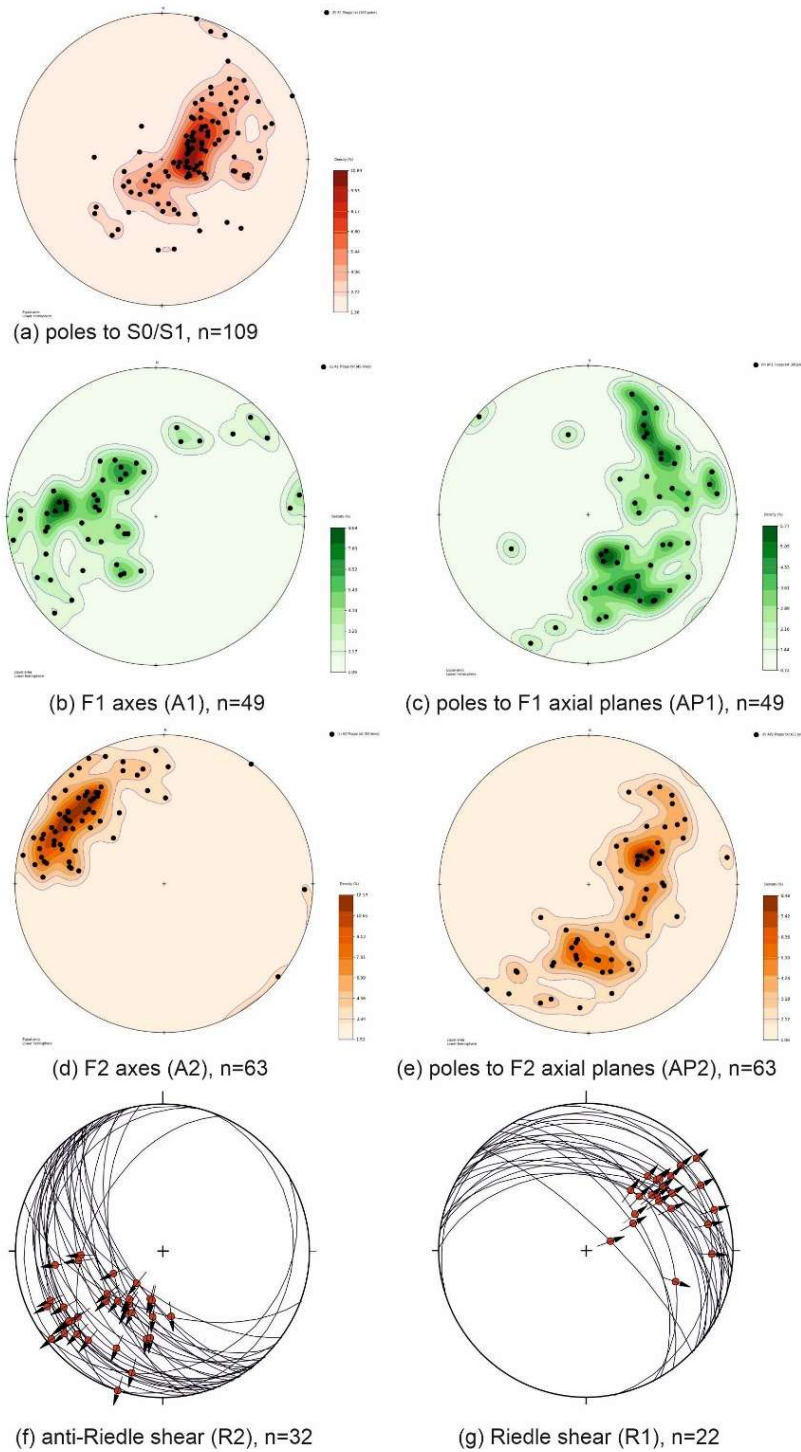


Figure 5-7: Stereographic projections (lower hemisphere, equiareal net): (a) poles to planes of bedding (S0) and D1-cleavage (S1). (b) F1 fold axes. (c) Poles to axial planes of F1 folds. (d) F2 fold axes. (e) Poles to F2 axial planes. Shear planes with slip vectors: (f) R2 (anti-Riedel shear); (g) R1 (Riedel shear).

5.3.4 Fold analysis

We analysed the orientations of the F1 and F2 fold axes and the axial planes. The Fleuty diagram presents a comparative classification of F1 and F2 folds from the Pioppi area (Figure 5-8 a).

The x-axis represents the dip of the axial plane (from 0° horizontal to 90° vertical). The y-axis shows the plunge of the fold axis (from 0° to 90°). The diagram (Figure 5-8 b) is divided into geometrical fold fields defined by Fleuty (1964), such as: gentle folds (low dip, low plunge), tight to isoclinal folds (high dip, low plunge), plunging folds (low dip, high plunge), upright and reclined folds (along the diagonal).

In the plot of Figure 5-8 a, green dots represent F1 folds and orange dots represent F2 folds. The distribution of F1 folds is relatively concentrated within the fields of inclined to recumbent folds. This consistent geometry suggests that F1 folds developed during an early compressional phase (D1), most likely associated with stacking or layer-parallel shortening. These folds typically show moderately to steeply dipping axial planes and gently to moderately plunging fold axes, indicating a relatively uniform stress regime during their formation.

In contrast, the F2 folds show a broader and more dispersed distribution across the diagram, with many plotting above the diagonal line. This indicates that F2 folds are dominantly upright, with horizontal to gently dipping plunging fold axes and variably dipping axial planes.

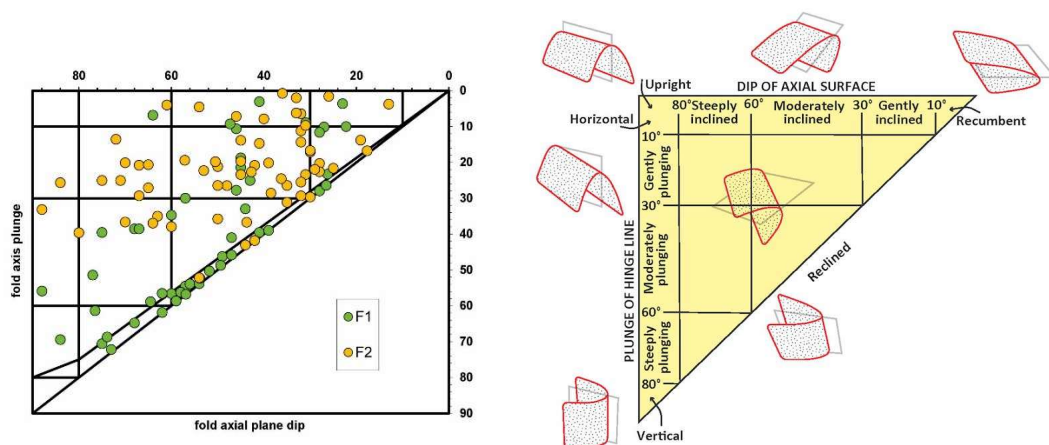


Figure 5-8: Fleuty diagram of F1 and F2 folds at Pioppi.

5.3.5 Microstructures

5.3.5.1 Optical microscope

In the Crete Nere Formation, bedding (S0) is still recognisable in the field and in thin section, although it is locally transposed by the first foliation (S1). S1 is a spaced crenulation/pressure-solution cleavage associated with the early F1 folding event. In most outcrops, S1 represents the dominant foliation at the mesoscale, whereas in thin section it appears as a subtle alignment of phyllosilicates and pressure-solution seams (Figure 5-9 a).

A second deformation phase produced F2 folds that overprint both S0 and S1. Within the thin sections, these F2 folds carry a weak but consistent axial-plane cleavage (S2). The small thrusts observed in thin section show millimetric offsets and cut across S1; they are interpreted as localised shear structures related to the late stages of the F2 deformation. Their presence suggests that strain became progressively localised along the hinges of F2 folds and in associated shear zones (Figure 5-9 b, c). S2 is best developed in the inner arcs of the folds, whereas in the outer arcs, strain is accommodated mainly by antitaxial calcite veins, indicating deformation partitioning (Figure 5-9 d). A picture of an entire thin section (Figure 5-9 e) shows a F2 fold with calcite veins in the outer arc and cleavage (S2) in the inner arc. A small thrust in the hinge is present.

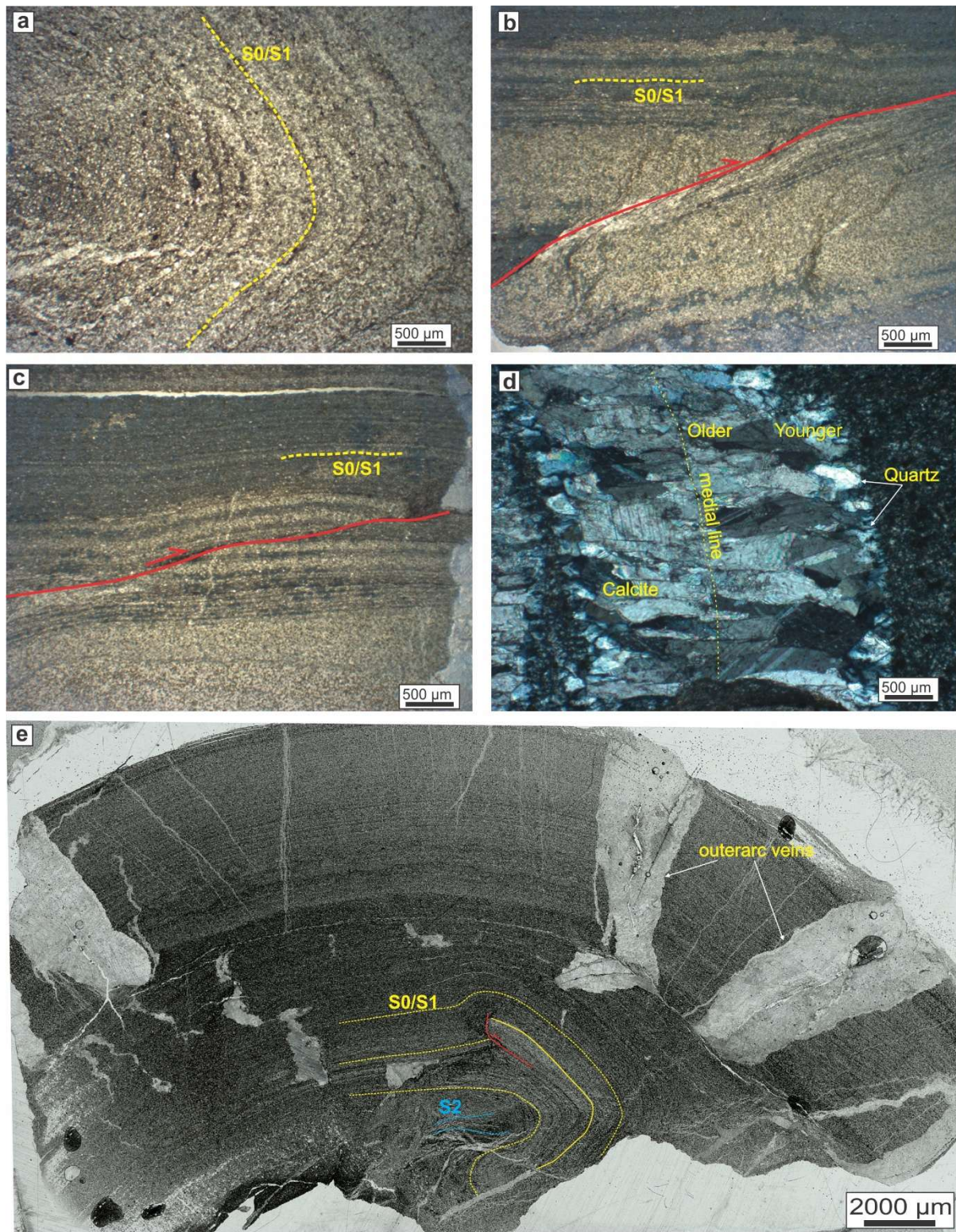


Figure 5-9: Microphotographs of thin sections by optical microscope. (a) F2 fold. (b, c) minor thrust faults with millimetric displacement. (d) Antitaxial calcite vein with younger cements made of quartz. (e) F2 fold with calcite veins in the outer arc and cleavage (s2) in the inner arc. A small thrust in the hinge is present.

The SEM micrographs reveal that deformation in the shear zone is dominated by a pervasive foliation (S2), which is the main foliation observable at the microscale (Figure 5-10). S2 defines the overall planar fabric and wraps around porphyroclasts, consistent with its interpretation as the axial-plane foliation of the F2 folding phase. This foliation is locally cut by R1 shear planes, indicating a second-order fabric associated with progressive simple shear and strain localisation (Figure 5-10 a, b). Porphyroclasts of calcite/dolomite and quartz display rounded to polygonal shapes, and many preserve internal relics of earlier fabrics (Figure 5-10 c-h). In several grains, a pre-existing S0/S1 foliation is visible as an inclusion trail or internal compositional layering, typically oriented at a high angle to the external S2 foliation (Figure 5-10 e, g). Kinematic indicators within the porphyroclast population include ϕ -type, σ -type, and δ -type geometries reflecting varying degrees of asymmetric strain accommodation (Figure 5-10). In other cases, they show a sigma (Figure 5-10) or a delta symmetry (Figure 5-10). The SEM data confirm that S2 is the dominant foliation within the shear zone, overprinting S1 and S0, and that deformation progressively localised into discrete shear bands during the D2 phase.

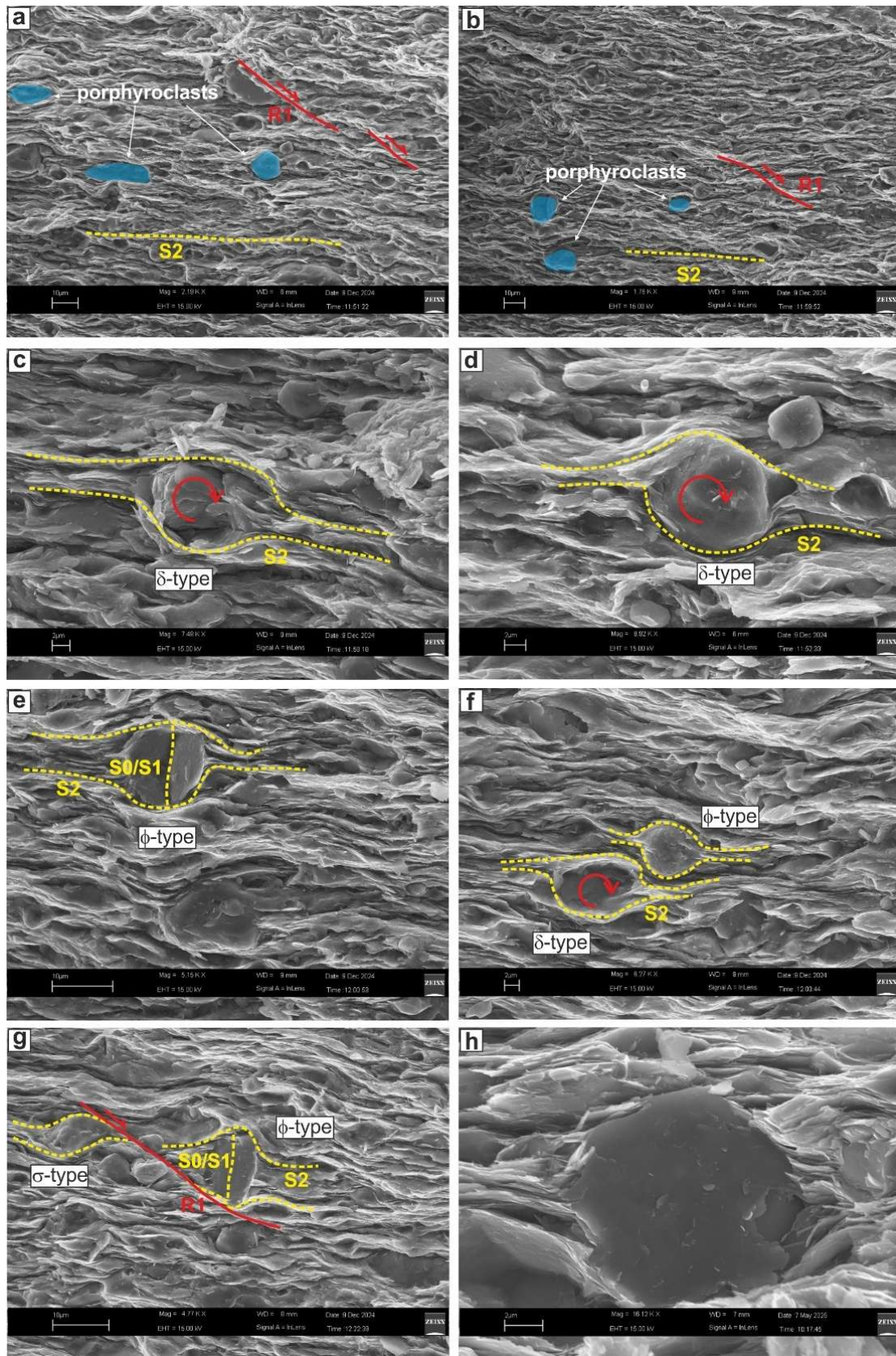


Figure 5-10: (a-b) mylonitic fabric of the deformed rocks within the PSZ, showing a main S2 foliation, porphyroclasts and Ridle shears (R2). (c-g) porphyroclasts showing a symmetry of phi- sigma- and delta-type and relic of the S0/S1 foliation. (h) Zoomed area showing the S2 foliation wraps the porphyroclast.

5.3.7 Pioppi Shear Zone: 3D strain analysis

To reconstruct the Finite Strain Ellipsoid (FSE) of the Pioppi Shear Zone, SEM pictures of 6 samples cut parallel to XZ and YZ planes of the FSE were analyzed. Samples include: PP1a (5 YZ and 4 XZ pictures; Figure 5-11), PP1b (3 YZ and 3 XZ pictures; Figure 5-12), PP1c (3 YZ and 3 XZ pictures; Figure 5-13), PP1d (3 YZ and 3 XZ pictures; Figure 5-14), PP1e (3 YZ and 3 XZ pictures; Figure 5-15), and PP2b (5 YZ and 5 XZ pictures; Figure 5-16).

The resulting FRY plots are shown in Figure 5-11 to 5-16. Due to the different strain values in XZ and YZ sections for each sample, all possible combinations have been plotted in the Ramsay Diagram (Logarithmic Flinn Diagram) to reconstruct the FSEs. For each couple of (R_{xz} , R_{yz}), the R_{xy} has been calculated as R_{xz}/R_{yz} . To better visualise the ellipsoid distribution, a density contouring was added (Figure 5-17). The Ramsay Diagram indicates that most of the FSEs fall within the oblate field.

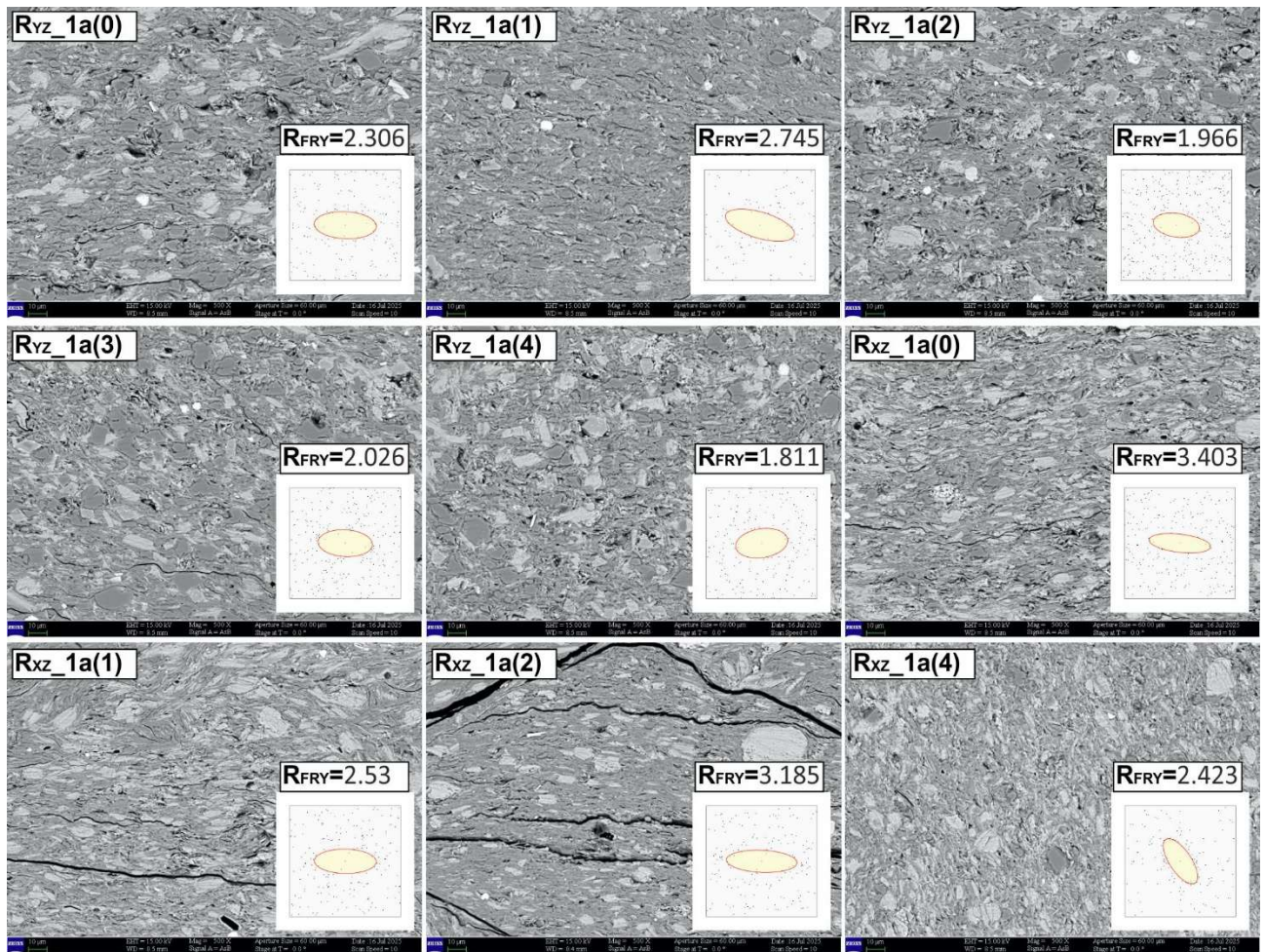


Figure 5-11: SEM images of sample PP1a (YZ and XZ sections) and corresponding FRY plots.

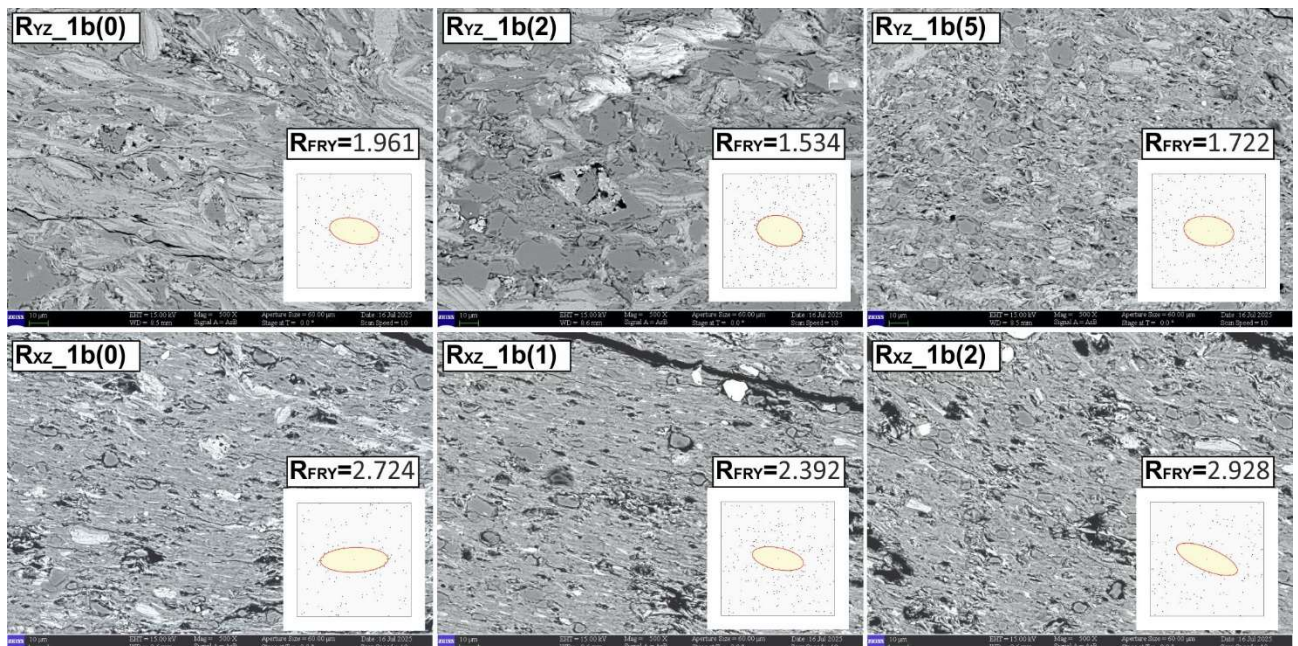


Figure 5-12: SEM images of sample PP1b (YZ and XZ sections) and corresponding FRY plots.

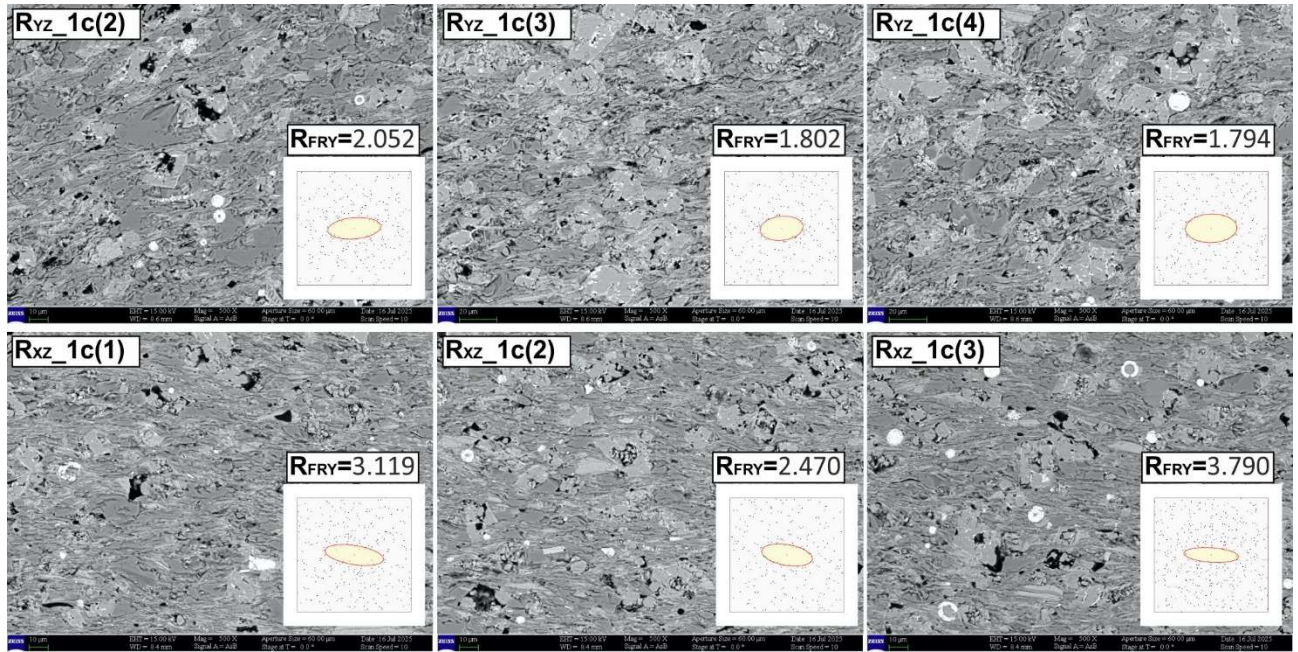


Figure 5-13: SEM images of sample PP1c (YZ and XZ sections) and corresponding FRY plots.

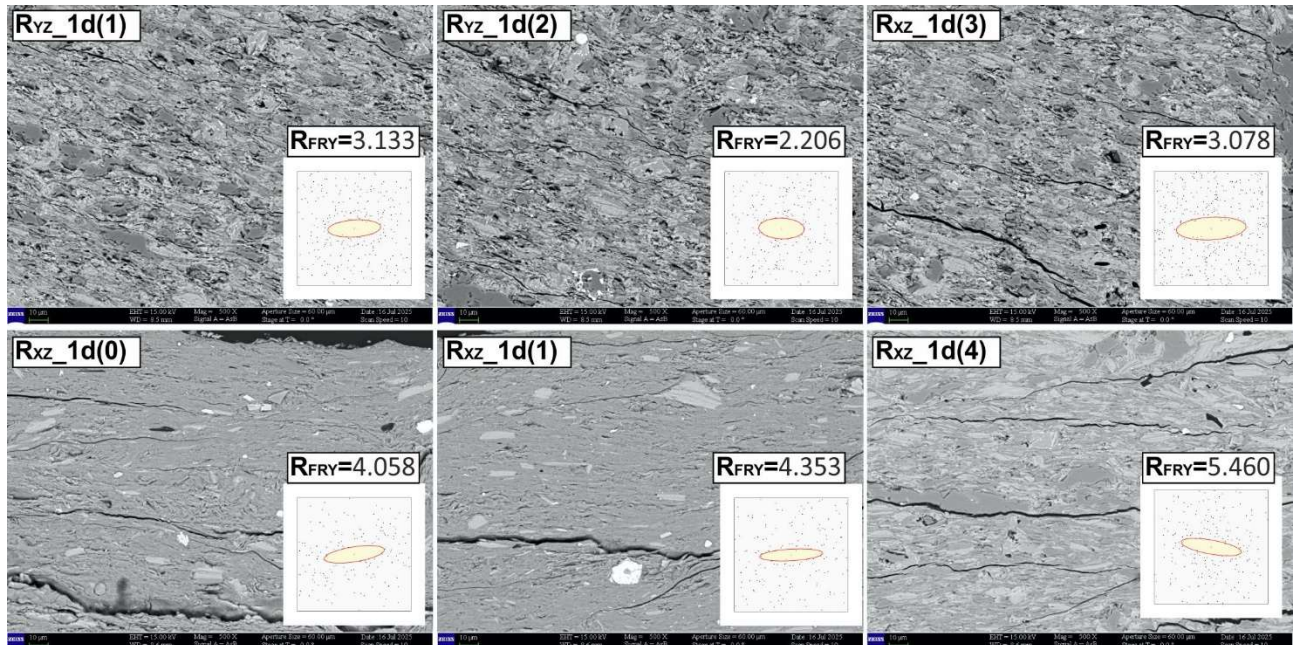


Figure 5-14: SEM images of sample PP1d (YZ and XZ sections) and corresponding FRY plots.

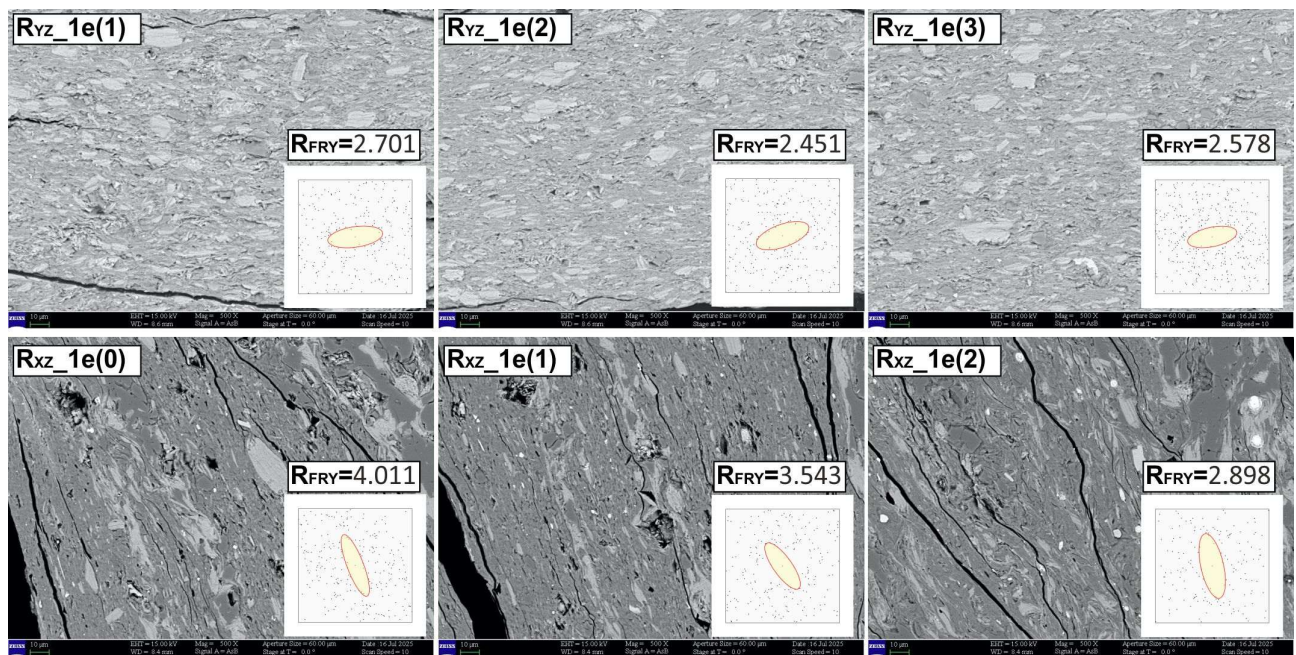


Figure 5-15: SEM images of sample PP1c (YZ and XZ sections) and corresponding FRY plots.

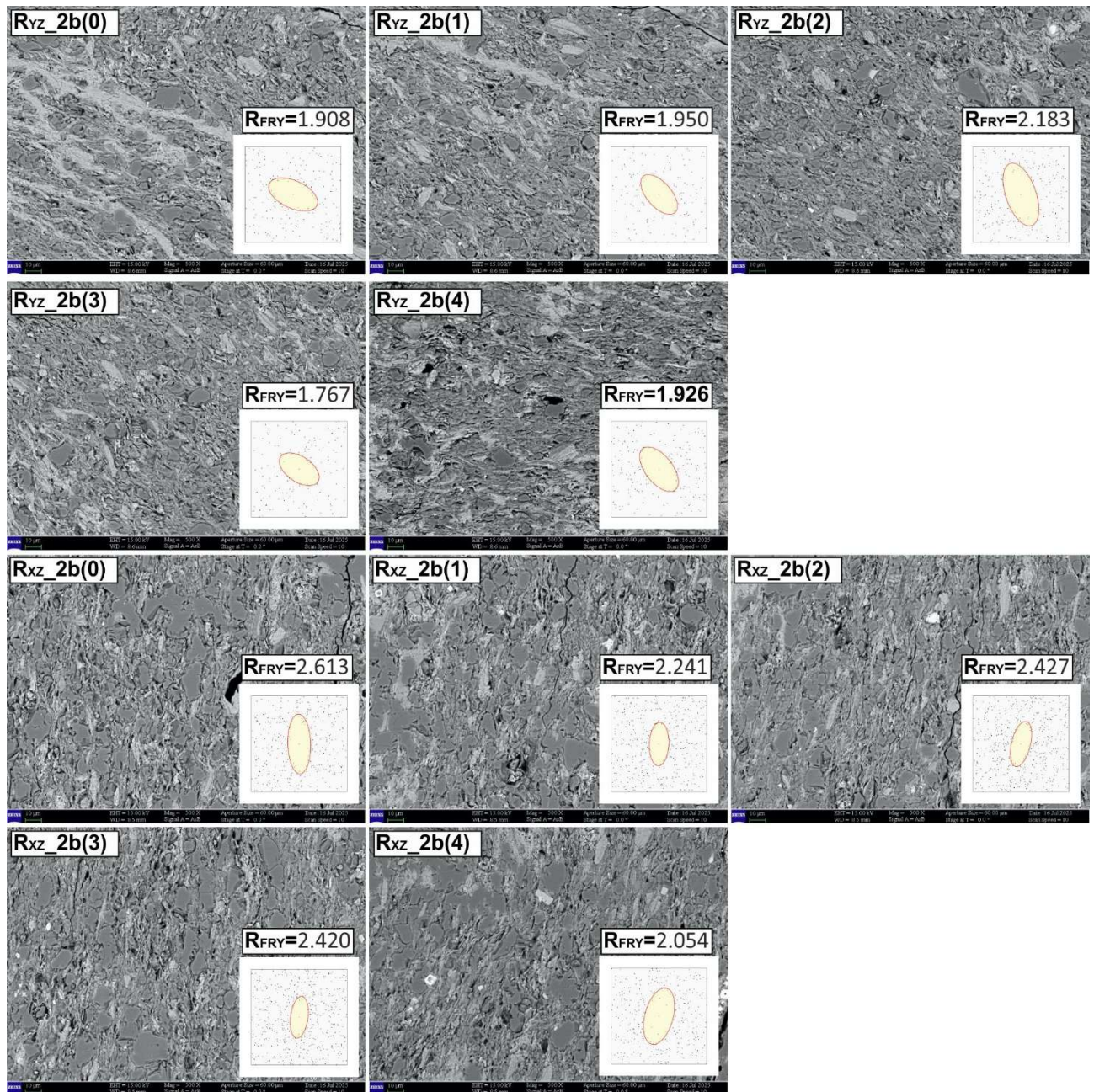


Figure 5-16: SEM images of sample PP2b (YZ and XZ sections) and corresponding FRY plots.

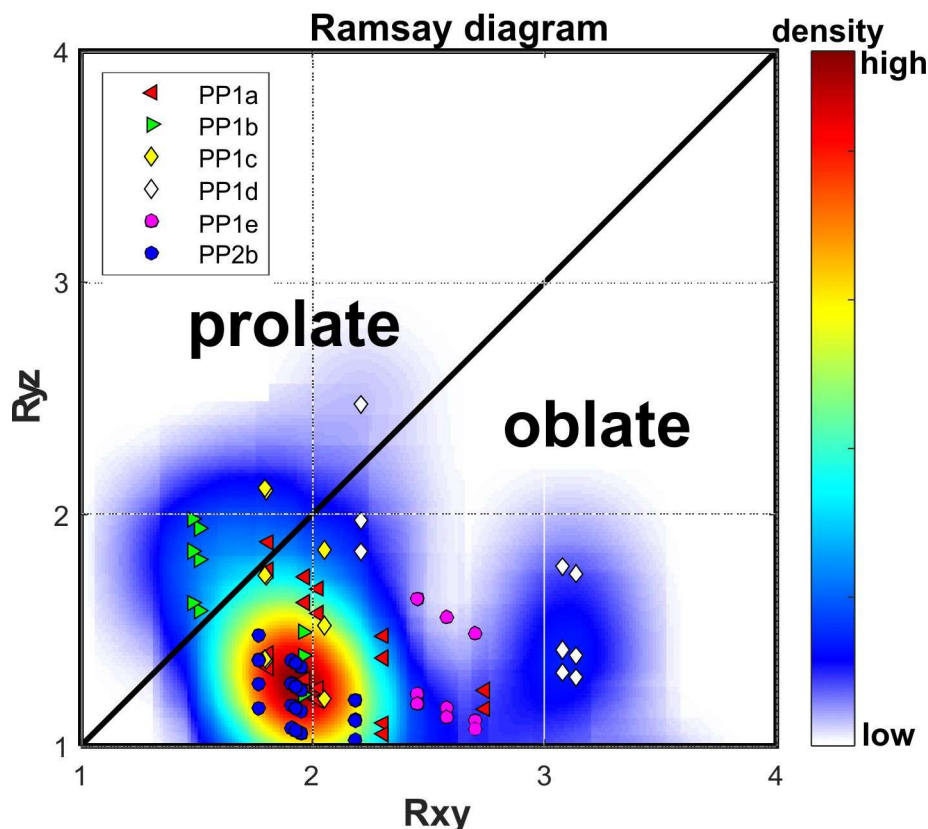


Figure 5-17: Ramsay diagram of Finite Strain Ellipsoids.

5.3.8 Pioppi Shear Zone: Kinematic vorticity analysis

We analysed several images from the field and SEM (Scanning Electron Microscope) of samples of the Pioppi Shear Zone (Figure 5-18), to estimate the kinematic vorticity and evaluate the type of shear (simple, pure or sub-simple shear).

Both R1-R2 and rotated porphyroclast (RGN) methods were used (see methodology section). Results are presented in Figure 5-19 a, g. The results are summarized (Figure 5-19 f) in terms of kinematic vorticity number (W_k), which is a non-dimensional parameter representing the ratio of simple shear to pure shear (with $W_k = 1$ being simple shear, $W_k = 0$ being pure shear, and $0 < W_k < 1$ being sub-simple shear).

Each diagram of Figure 5-19 a–e, corresponds to data derived from specific SEM images labelled P41, P91, P21, P11, and P01. The plots show the rose diagram (angle frequency plot) of the measured angle between the main foliation and the R1 shear plane. W_k was calculated by the mean value of theta angle. The arrows in red show the shear sense. Figure 5-19 f, uses the Ramsay-Grasemann-Nordin (RGN) technique to derive W_k . The distribution is scattered, and a no-clear threshold can be identified, hence a conservative value of $W_k \approx 0.9$ is considered.

Figure 5-19 g summarizes all kinematic vorticity estimations, where W_k value are indicated by the x-axis and the corresponding percentage of pure shear vs. simple shear by the y-axis. W_k data range between 0.87–0.98, indicating that deformation in these rocks was dominated by simple shear (~67–97%), but with a component of pure shear of ~33–3%, confirming deformation under general shear conditions.

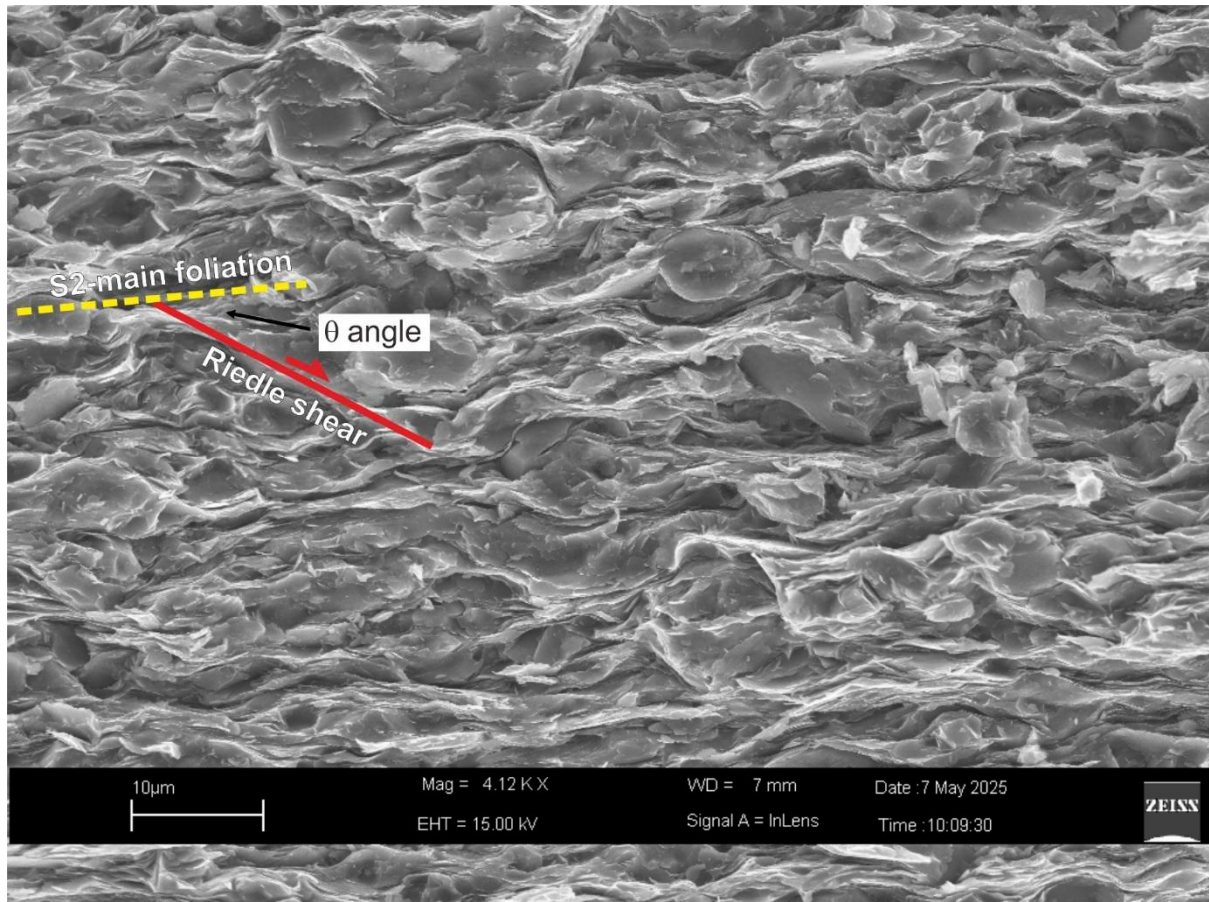


Figure 5-18: Example of SEM image analyzed to estimate the theta angle between main foliation and Riedle shear.

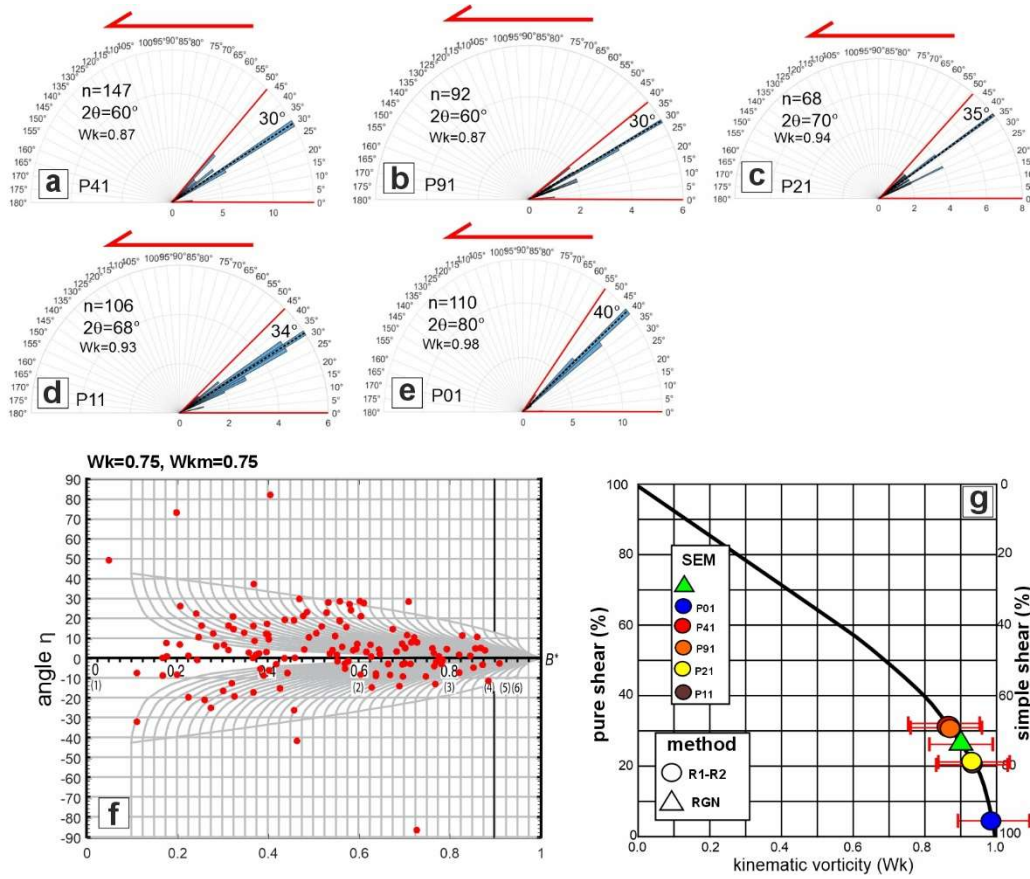


Figure 5-19: Results of the vorticity analyses performed on SEM pictures from Pioppi shear zone. (a–e) S-C' structures method: (a) P41; (b) P91; (c) P21; (d) P11; (e) P01. (f) Rigid porphyroclast method; (g) Wk versus pure shear/simple shear percentile plot.

5.4 Punta Telegrafo

5.4.1 Mesostructures

3.4.1.1 Folds and Cleavages

Two superposed sets of folds intensely deform the succession exposed in Punta Telegrafo. The first fold set (F1) is characterised by isoclinal to tight geometries (Figure 5-20 a, c). Generally, they have rounded hinges in the calcareous folded layers and a hinge thickening in the pelitic layers. F1 folds show a large orientation variability, from upright (Figure 5-20 a) to vertical (Figure 5-20 c). The F1 folds are invariably deformed by the subsequent fold set (F2), forming interference patterns that range between type 3 (Figure 5-20 a) or between types 2 and 3 (Figure 5-20 d), indicating parallel F1-F2 axes or an angle between the two-fold axes. F2 folds show horizontal axes and more open geometries (Figure 5-20 e). Some ptygmatic F2 folds are present (Figure 5-20 b), formed by calcite veins embedded in the pelitic matrix. They generally show a tectonic vergence top-to-the SE.

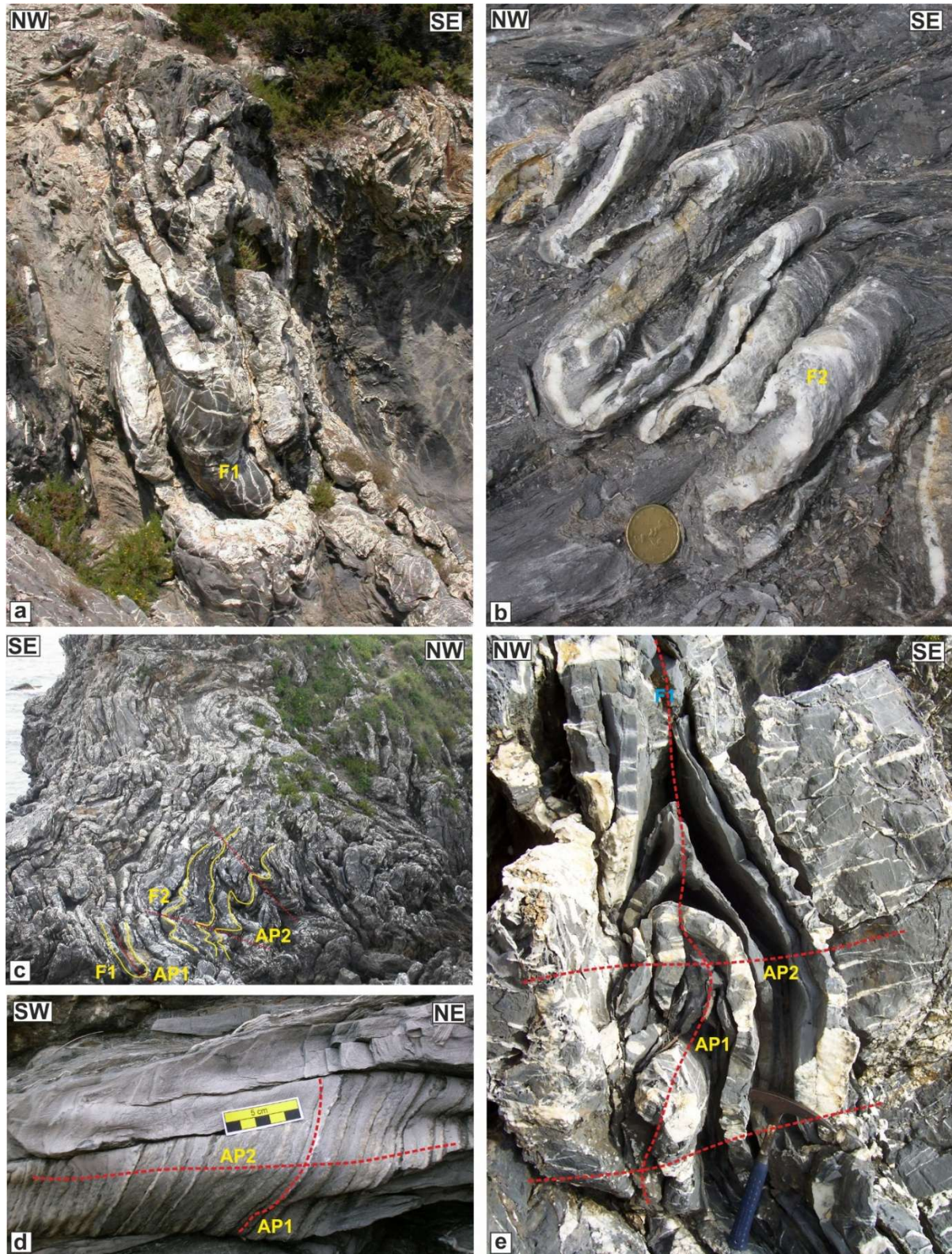


Figure 5-20: (a) Vertical F1 fold. (b) F2 ptygmatic folds of a calcite vein within a pelitic matrix. (c) F2 folds verging to SE. (d) vertical F1 ptygmatic folds refolded by F2 horizontal folds. (e) interference pattern of type 3 between F1 and F2 folds, both with a sub-horizontal fold axis.

Interference pattern of type 3 between an isoclinal F1 fold and F2 folds, both with subhorizontal fold axes. F1 and F2 folds are associated with the development of two pervasive cleavages S1

and S2 (Figure 5-21 a-e). The S1 cleavage is parallel to the bedding S0, and parallel to the F1 axial planes. It is a disjunctive cleavage. S2 foliation is a crenulation cleavage (Figure 5-21 a-e). The S2 is well-developed in the tight hinges of F2 synforms (Figure 5-21 e).

Veins are widely developed. They are associated with a severe boudinage affecting the stiffer calcareous layers (Figure 5-21 f, g). Generally, veins form different sets with different orientations; however, two orthogonal sets are widespread, forming a chocolate tablet boudinage (Figure 5-21 g, Figure 5-22 a, b).

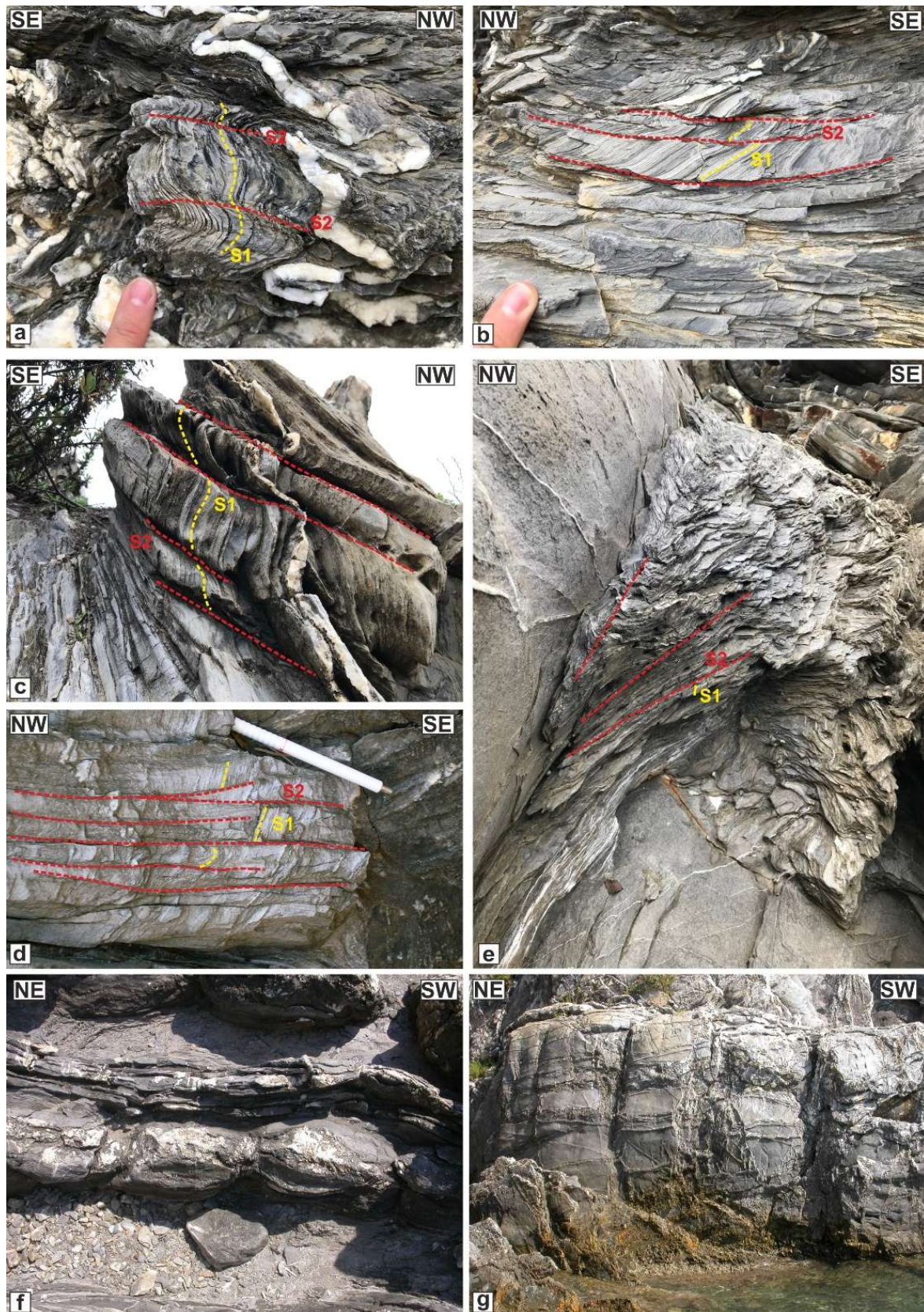


Figure 5-21: (a-e) S2 crenulation cleavage superposed to the S1 disjunctive cleavage. (f) boudinage of a calcareous layer embedded by pelitic layers. (g) Chocolate tablet boudinage affecting a calcareous layer.

3.4.1.2 Veins

Some veins show an *en echelon* geometry (Figure 5-22c), indicating both dextral and sinistral sense of shear associated with conjugate veins. As described before, other veins form ptigmatic folds (Figure 5-22 d, f) associated with the D2 folding stage.

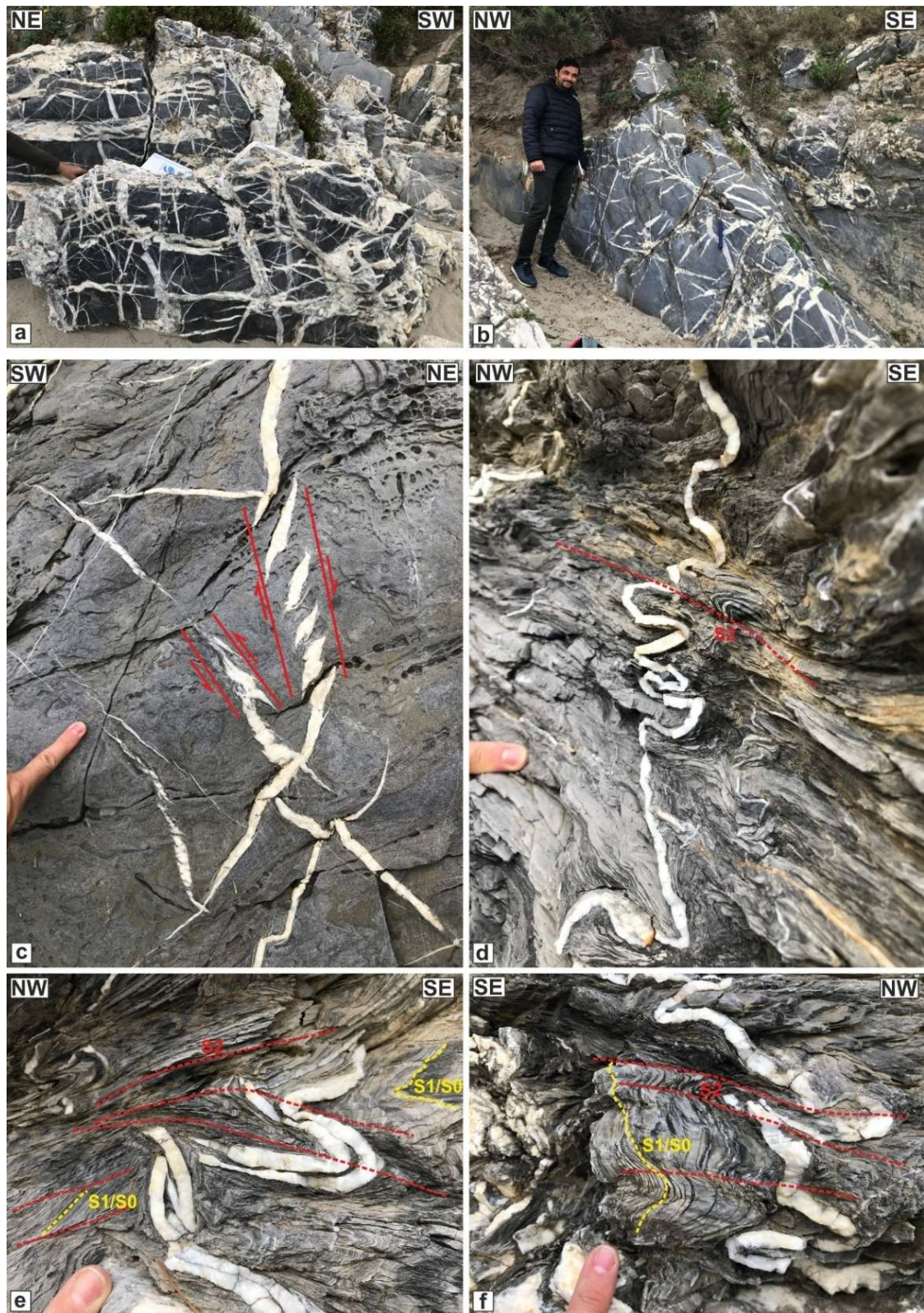


Figure 5-22: (a) Chocolate tablet boudinage affecting a calcareous layer. (b) Chocolate tablet boudinage affecting a F1 fold. (c) Conjugate veins with en echelon geometry. (d-f) Ptygmatic veins with associated S2 crenulation cleavage deforming a S1/S0 foliation.

5.4.2 Stereographic projections

Poles to bedding (Figure 5-23 a) indicate a broad girdle distribution with a theoretical fold axis (p) of 051/03. Poles to veins (Figure 5-23 b) show three main clusters corresponding to three average vein sets: V1=178/78, V2= 065/81 and V3=139/70. When restored, rotating the host layer to the horizontal, the projection (Figure 5-23 c) indicates that the V2 and V3 set do not change, on the contrary the V1 set change furnishing an average plane of 316/45. This indicates that the V2 and V3 sets are closely related to the F1-F2 folding, whereas the V1 was formed before the folding stages. The stereographic projection of F1 axes (Figure 5-23 d) forms a girdle distribution with NE-SW direction, whereas poles to axial planes (Figure 5-23 e) form two clusters with average planes of 149/69 and 314/82. Finally, projection of F2 axes (Figure 5-23 f) indicates a main cluster with an average axis of 328/00, whereas the poles to F2 axial planes (Figure 5-23 g) suggest a main cluster with an average plane of 219/06.

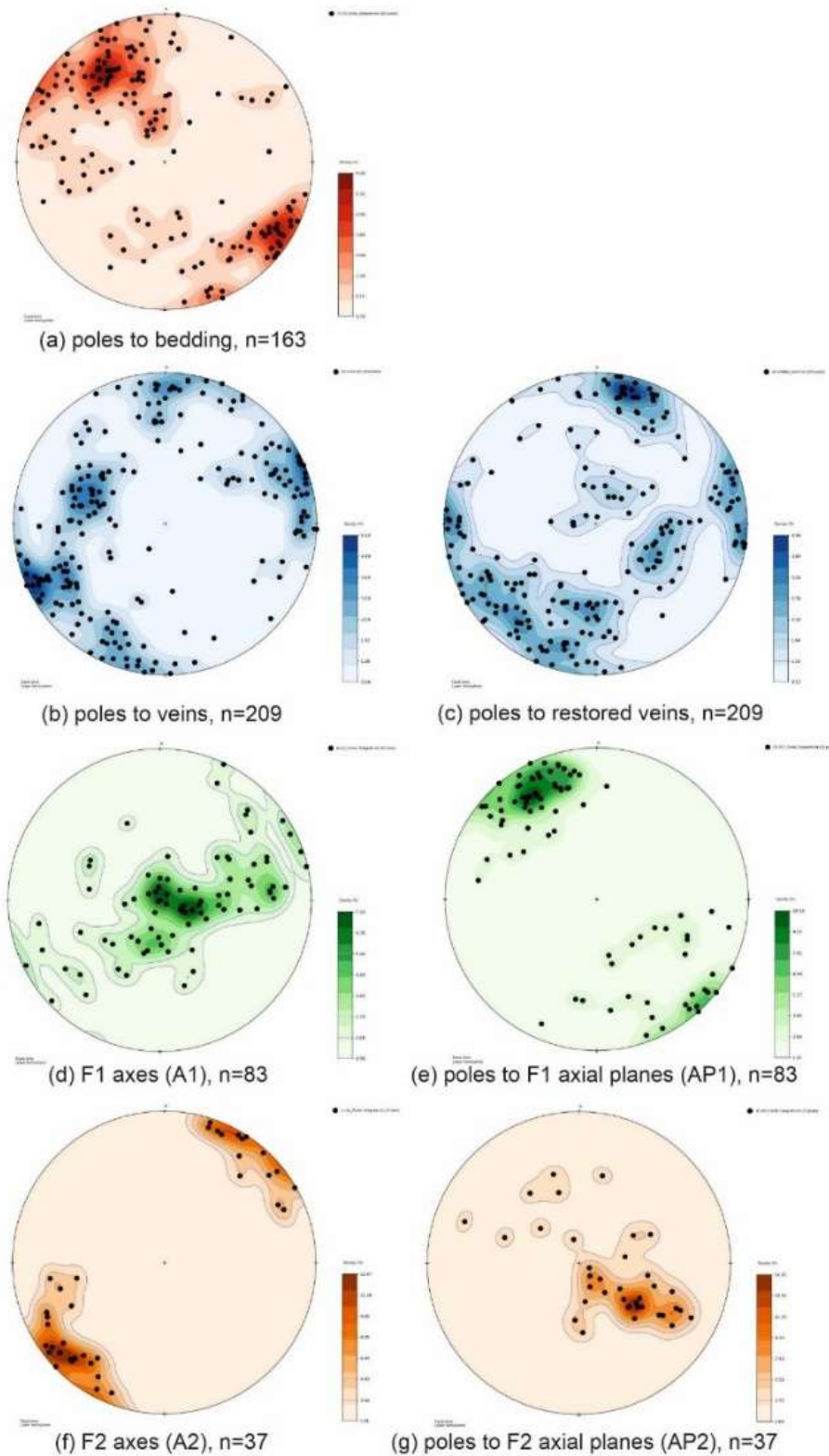


Figure 5-23: Stereographic projections (lower hemisphere, equilateral net) of structures at Punta Telegrafo: (a) Poles to bedding; (b) Poles to veins distribution; (c) Restored projection after rotating the host layer to horizontal; (d) F1 fold axes with a NE-SW direction; (e) Poles to axial planes of F1, showing two clusters; (f) F2 fold axes; (g) Poles to F2 axial planes.

5.4.3 Fold analysis

We analysed F1 and F2 data through the Fleuty diagram (Figure 5-24 a) to understand the geometric classification. The graph plots fold axial plane dip (horizontal axis) against fold axis plunge (vertical axis) to provide insight into the geometry of the F1 and F2 folds observed in the field (Figure 5-24 b).

The F1 folds (green) are mostly concentrated in the lower left and along the diagonal, where: Axial plane dips are steep, generally between 50° and 90° . Fold axis plunges are low to moderate, from 0° up to $\sim 40^\circ$ (Figure 5-24 a). This cluster suggests that F1 folds are predominantly upright to inclined, tight to isoclinal folds.

The F2 folds (orange) appear more scattered but tend to cluster toward the upper right corner of the plot (Figure 5-24 a): Fold axes have low plunges, mostly between 0° and 30° . Axial plane dips are generally gentle to moderate, from 0° to 50° . This indicates that F2 folds are largely open to gentle folds with sub-horizontal axial planes and commonly plunging gently or nearly horizontal.

The Fleuty diagram shows a clear kinematic and geometric contrast between F1 and F2 folds. F1 folds reflect an early compressional phase, forming tight and upright folds typical of deep-seated high deformation and initial nappe emplacement. F2 folds suggest a later deformation event under a lower strain regime.

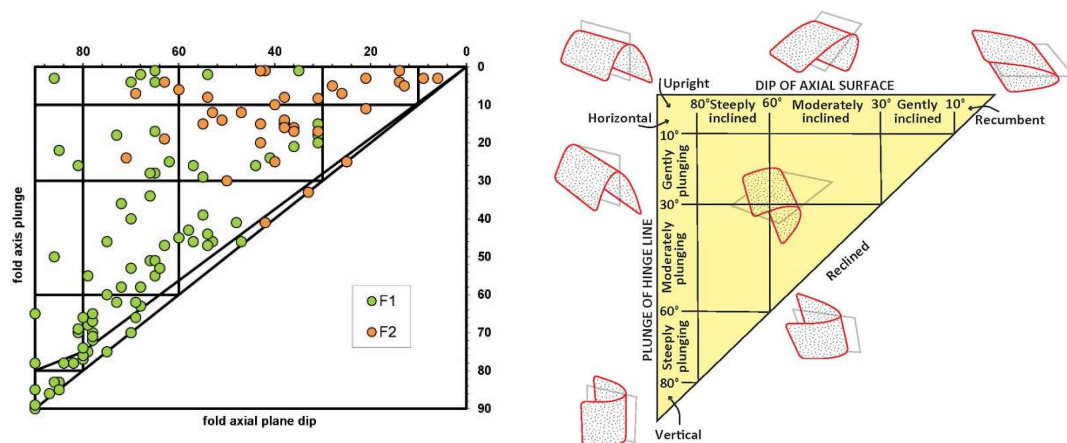


Figure 5-24: Fleuty diagram of the F1 and F2 folds at Punta Telegrafo.

5.4.4 Microstructures

5.4.4.1 Optical microscope

Some samples of veins were collected, prepared as thin sections, and observed using an optical microscope. Most of the veins (V_B) are orthogonal to the bedding, generally forming conjugate sets (Figure 5-25 a). The number of veins and their thickness depend on the lithology and grain size. In Figure 5-25 a, two layers with different grain size but the same lithology host the same vein system; however, the veins in the coarser layer are fewer and thicker than veins in the finer layer, where there are a lot of thin veins. Veins about orthogonal to the bedding are formed by different superposing sets, here named $V_{B1,2,3}$ (Figure 5-25 b). Generally, they are formed by elongate or fibrous calcite in the centre (antitaxial veins) and quartz at the boundaries. Generally, associated with the veins V_B , stylolites occur orthogonal to the veins (Figure 5-25 c). The latter structures have been used as preferred surfaces for the new veining stage (V_T). In this case, vein V_B are deformed by orthogonal veins V_T (Figure 5-25 d, e). Some veins show a blocky growth of calcite, however, always with quartz at the boundaries (Figure 5-25 f). These blocky crystals are equidimensional sparry calcite affected by mechanical twinning (Figure 5-25 g). Finally, some F2 folds with associated crenulation cleavage S_2 are present in the slaty samples (Figure 5-25 h).

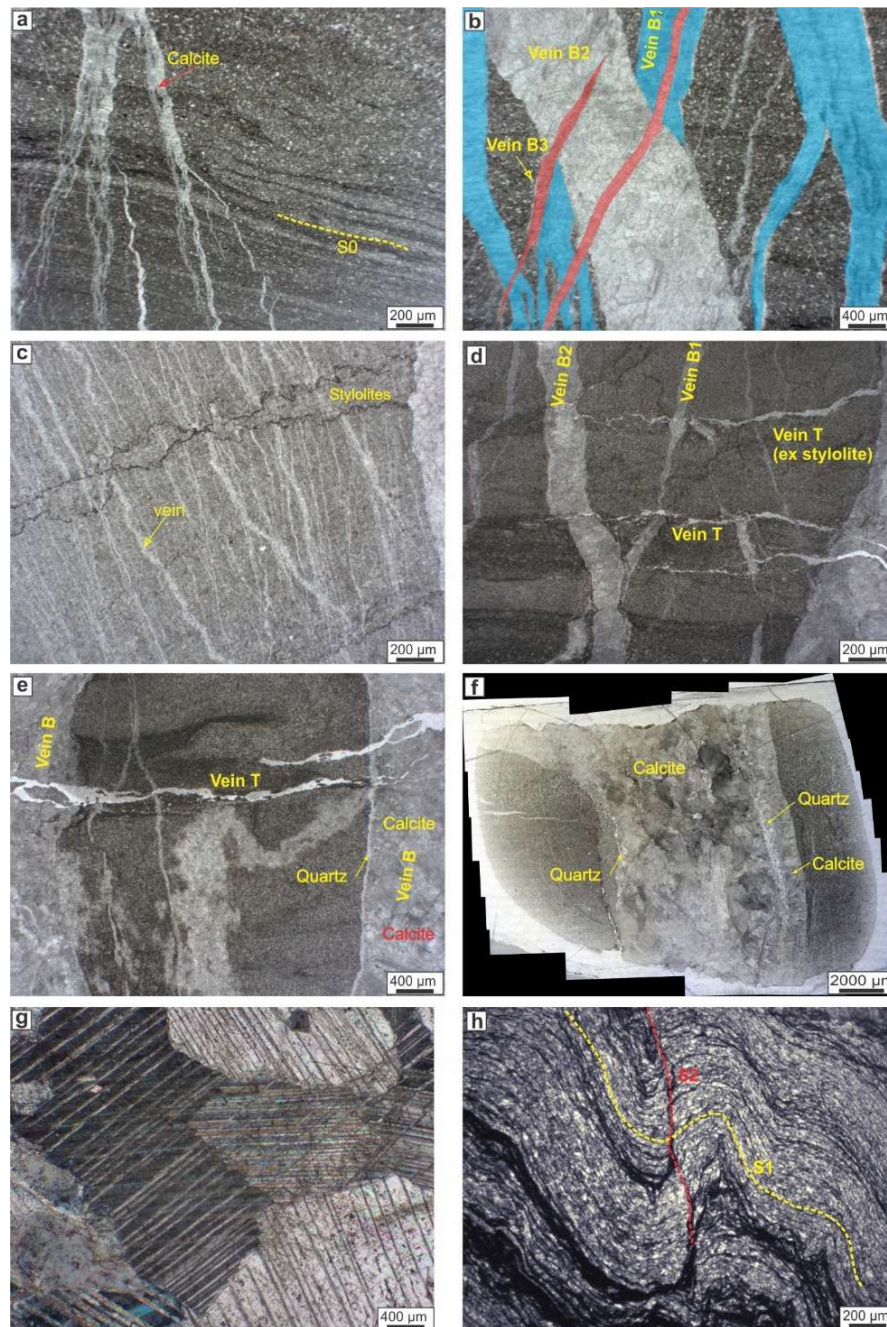


Figure 5-25: Microphotographs of thin sections by optical microscope. (a) Orthogonal veins to the bedding, forming conjugate sets, (b) Superposing vein sets (VB1, VB2, VB3), (c) Stylolite's occurring orthogonal to the veins, (d, e) Deformation of veins by orthogonal veins, (f) Blocky calcite growth within veins, with quartz at the boundaries, (g) mechanical twinning, (h) F2 folds with associated crenulation cleavage (S2).

5.4.4.2 SEM

SEM images have been analyzed to investigate the microstructures (Figure 5-26 and Figure 5-27). Punta Telegrafo pelites preserve two superposed foliations, S1 and S2 (Figure 5-26 a, g). S1 is a disjunctive to spaced cleavage, developed during the early deformation phase (D1), and is best recognised in the microlithons where it forms intrafolial microfolds (Figure 5-26 a,

d). S2 corresponds to a later crenulation cleavage that overprints and folds the earlier S1 fabric (Figure 5-26 a, g). In less deformed samples, SEM images clearly show S1 being folded into microfolds that define the axial-plane S2 crenulation cleavage. This demonstrates that S2 developed during the D2 deformation phase through progressive crenulation and reorientation of the pre-existing S1 foliation (Figure 5-27). Here, the S1 cleavage is folded by microfolds, developing the S2 crenulation cleavage. The S1 is composed of alternating cleavage domains rich in white mica and microlithons composed mainly of quartz and carbonate minerals, indicating mechanical anisotropy during D1 deformation (Figure 5-26 g, h). The Riedel shears appear locally, indicating zones of strain localisation within the D2 crenulation domains. These shear bands record discrete increments of simple shear associated with the later stage of the D2 deformation (Figure 5-26 g).

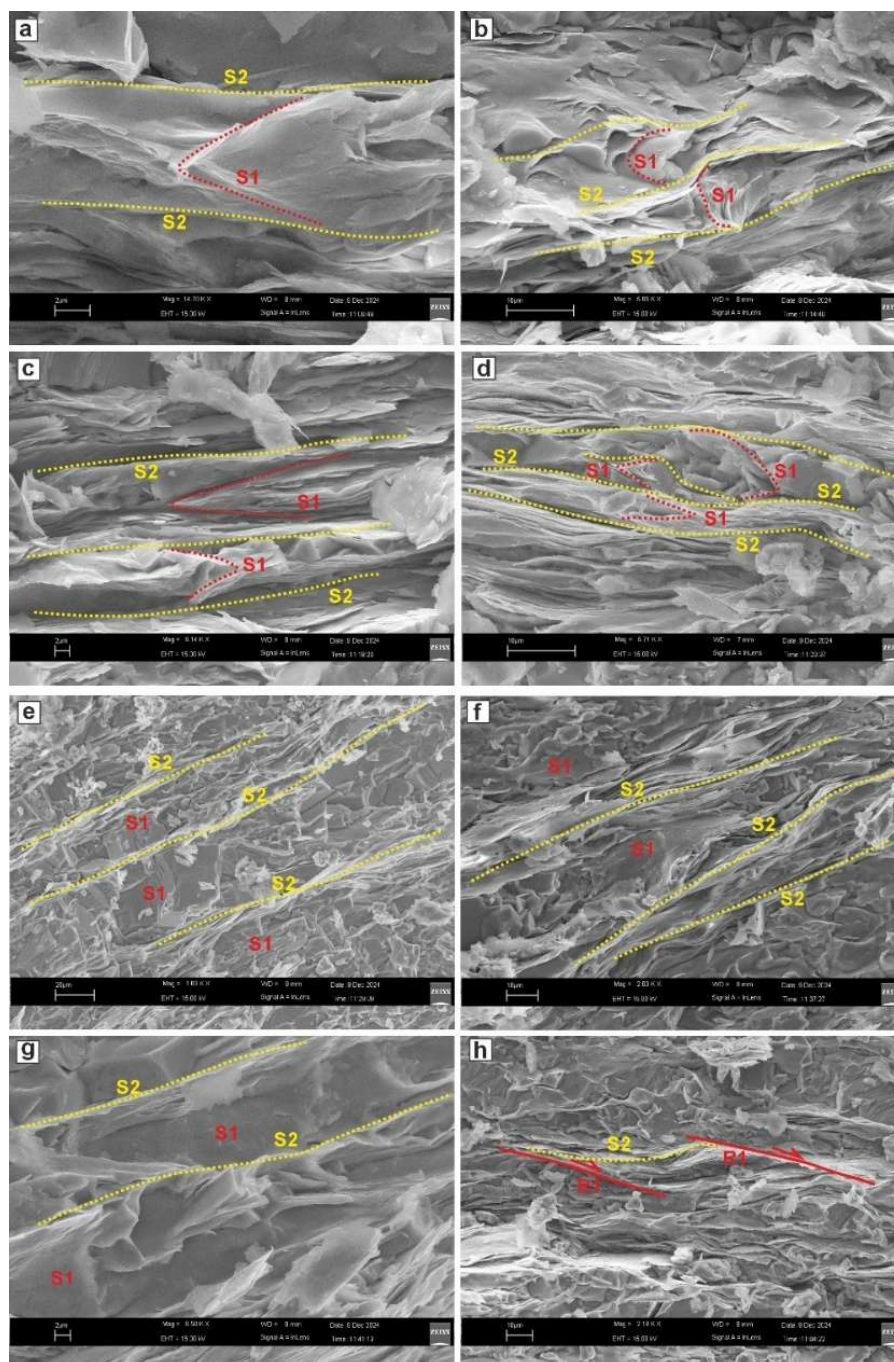


Figure 5-26: (a-g) Examples of S2 crenulation cleavage with F1 intrafolial folds within the microlithons. (h) Riedle shears deforming S2 main foliation.

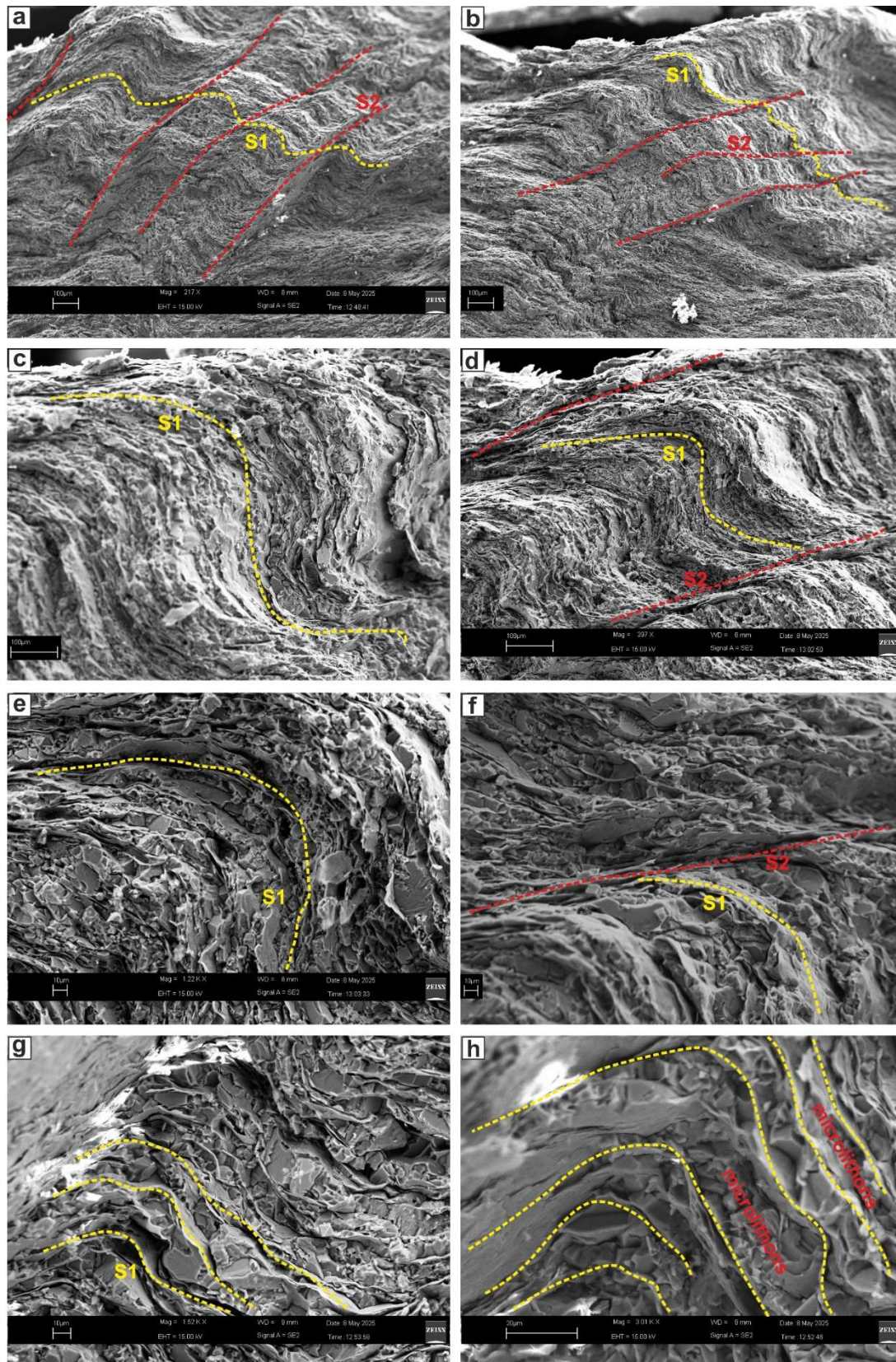


Figure 5-27: (a-f) Examples of crenulation and crenulation cleavage (S2). (g-h) S1 disjunctive cleavage folded by F2 crenulation, showing the cleavage domain, dominated by white mica and microlithons dominated by quartz and carbonate minerals.

5.5 Caprioli

5.5.1 Mesostructures

5.5.1.1 Folds

The Caprioli outcrops expose a highly deformed succession like the previous sections, but here the structural complexity is greater due to the presence of three superposed fold sets. The earliest deformation phase (D1) produced isoclinal to tight F1 folds whose axial planes are subparallel to the original bedding (S0) (Figure 5-28 a). These folds are associated with a pervasive axial-plane cleavage (S1), which locally transposes S0 where deformation is most intense. In zones of high strain, competent arenaceous layers underwent severe boudinage, resulting in stretched and rotated boudins (Figure 5-28c, d), indicating strong competence contrast during D1 deformation.

A subsequent deformation phase (D2) generated a superposed set of folds (F2), producing a Type-3 interference pattern with the earlier F1 structures (Figure 5-28 b). F2 folds range from tight to open and are commonly associated with minor thrust faults, suggesting that shortening during D2 involved both ductile folding and brittle–ductile shear (Figure 5-28 g). It is worth noting that the fold length varies with the thickness of the deformed layer, frequently forming polyharmonic folds (Figure 5-28 b). Figure 5-28 (g), also shows that a previously boudinaged layer is shortened by minor thrust faults. Frequently, ductile shear zones formed between two stiffer layers, within the pelitic bed, where previous veins are deformed: some of them are stretched, others are shortened (Figure 5-28 h).

5.5.1.2 Crenulation cleavages

The third fold set (F3) occurs only as a crenulation. In fact, the most spectacular features of this area are the occurrence of F2-F3 crenulation and crenulation lineations (L2 and L3) (Figure 5-28). S2 foliation is a crenulation cleavage, deforming the previous S1 foliation (Figure 5-29 a, b, c). Frequently, F2 crenulation on the S1 planes consists of non-cylindrical microfolds, generating a complex pattern of lens-shaped hinges (Figure 5-29 d, e). Often, F2 and F3 crenulations superpose with a high angle (Figure 5-29 f, g, h).

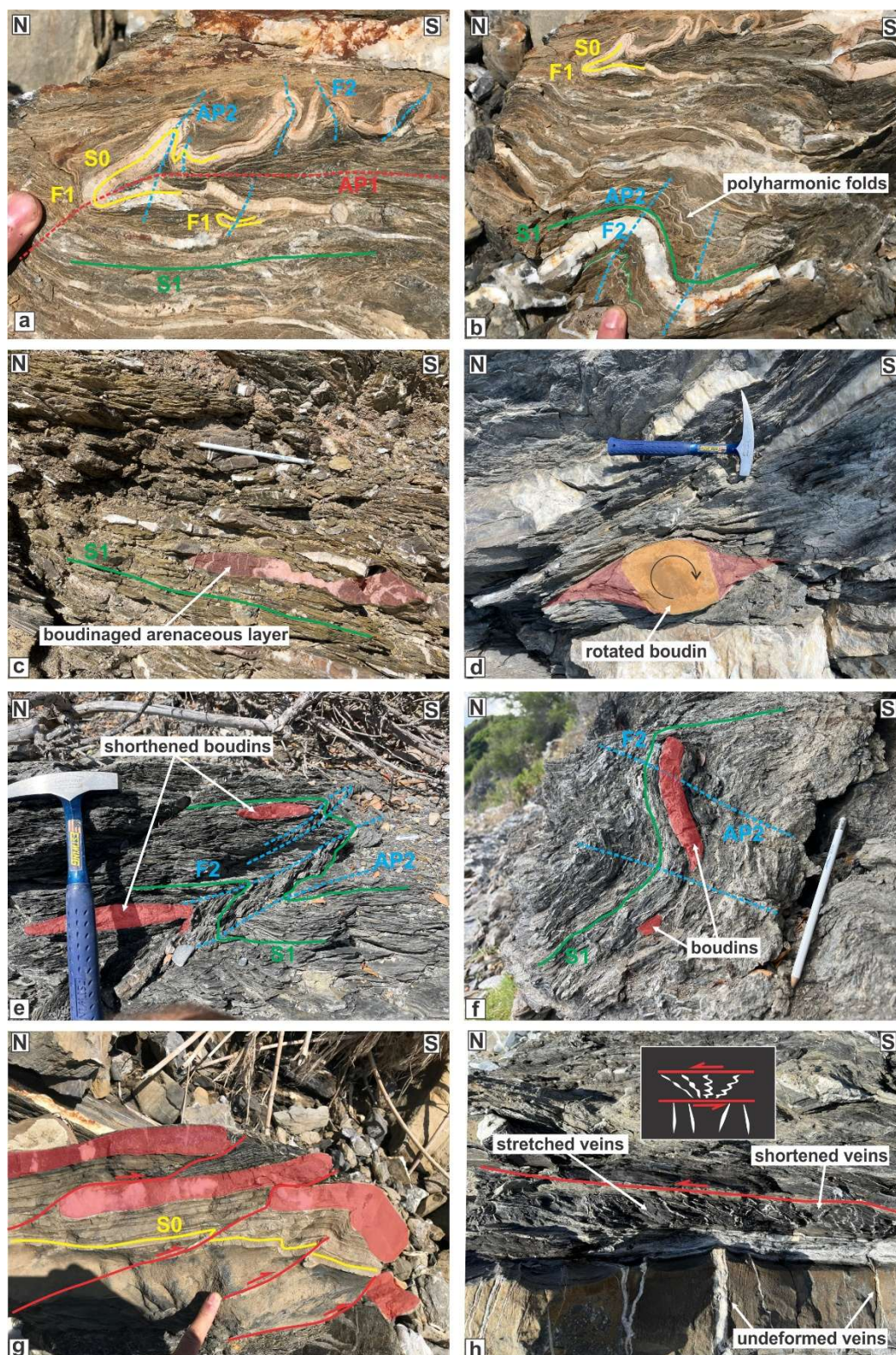


Figure 5-28: (a) F1 fold, refolded by F2. (b) F2 fold with minor polyharmonic folds. (c) Highly deformed zone with boudinaged layers. (d) Rotated boudin. (e) F2 fold verging to S. (f) F2 fold verging to N. (g) Minor top-to-the S thrust faults, shortening a previously boudinaged arenaceous layer. (h) Ductile shear zone within a pelitic layer, showing stretched and shortened veins (Flexural flow) compared to the underlying stiffer arenaceous layer bearing undeformed veins. The inset shows the scheme of the deformed veins.

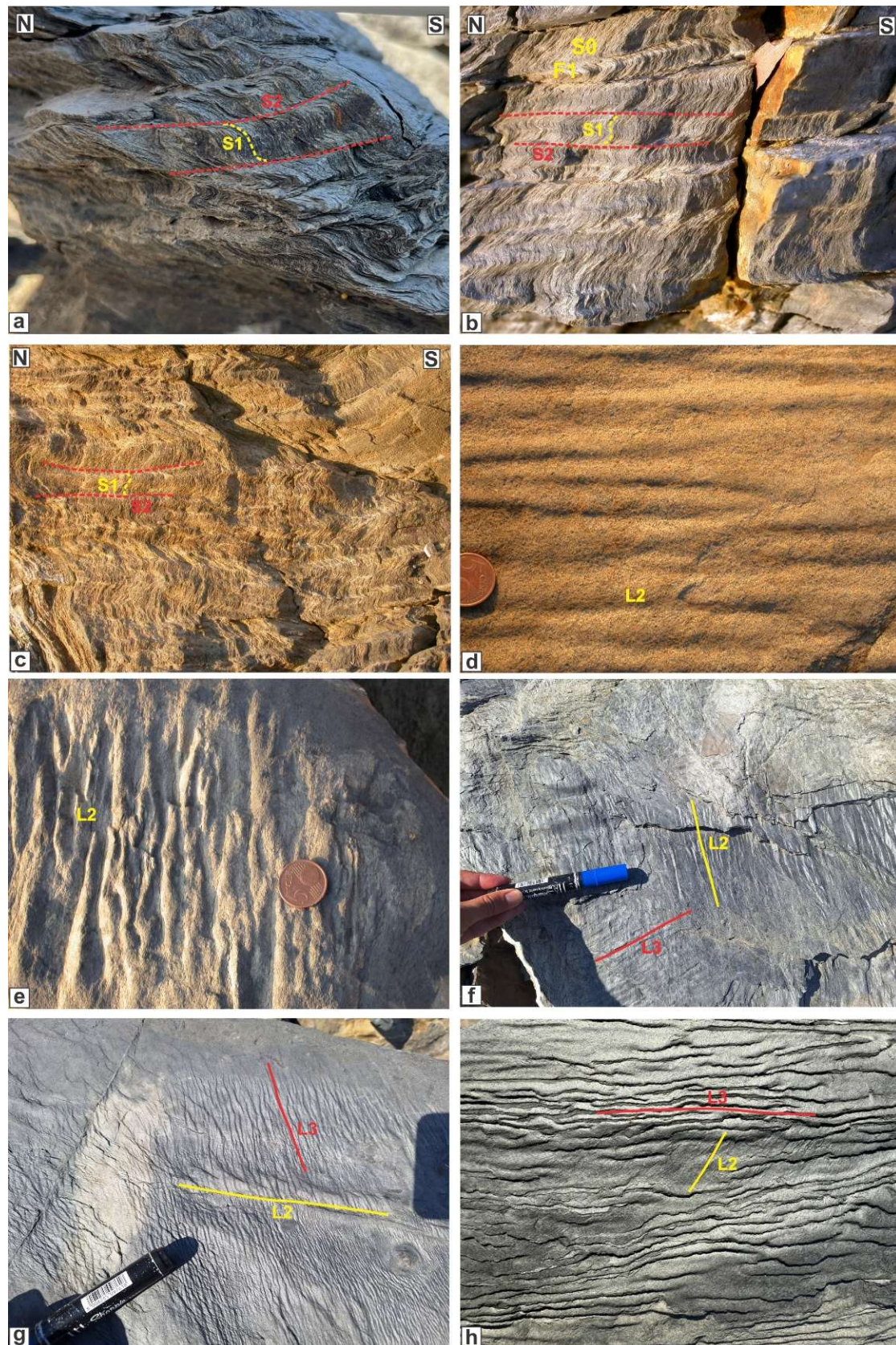


Figure 5-29: (a, b, c) F2 crenulation with S2 crenulation cleavage. (d, e) F2 crenulation hinges, showing non-cylindrical axes and a lens-shaped pattern. (f, g, h) Superposition between L2 and L3 crenulation lineations.

5.5.2 Stereographic projections

The projections of F1 fold axes (Figure 5-30 a) form a girdle distribution with a theoretical fold axis (π_1) of 083/11. Poles to F1 axial planes (Figure 5-30 b) indicate average planes dipping to N/NE. F2 fold axes (Figure 5-30 c) indicate a mean subhorizontal axis with a E-W direction. Poles to F2 axial planes (Figure 5-30 d) suggest an average plane dipping to N. Crenulation lineation associated with the F2 folds (Figure 5-30 e) form a cluster indicating a mean subhorizontal lineation with a ENE-WSW direction. Poles to S0/S1 planes (Figure 5-30 f) form a semi-girdle distribution with a theoretical fold axis (π_2) of 300/04. Finally, Riedle (Figure 5-30 g) and anti-Riedle shears (Figure 5-30 h) suggest a tectonic transport top-to-the SE.

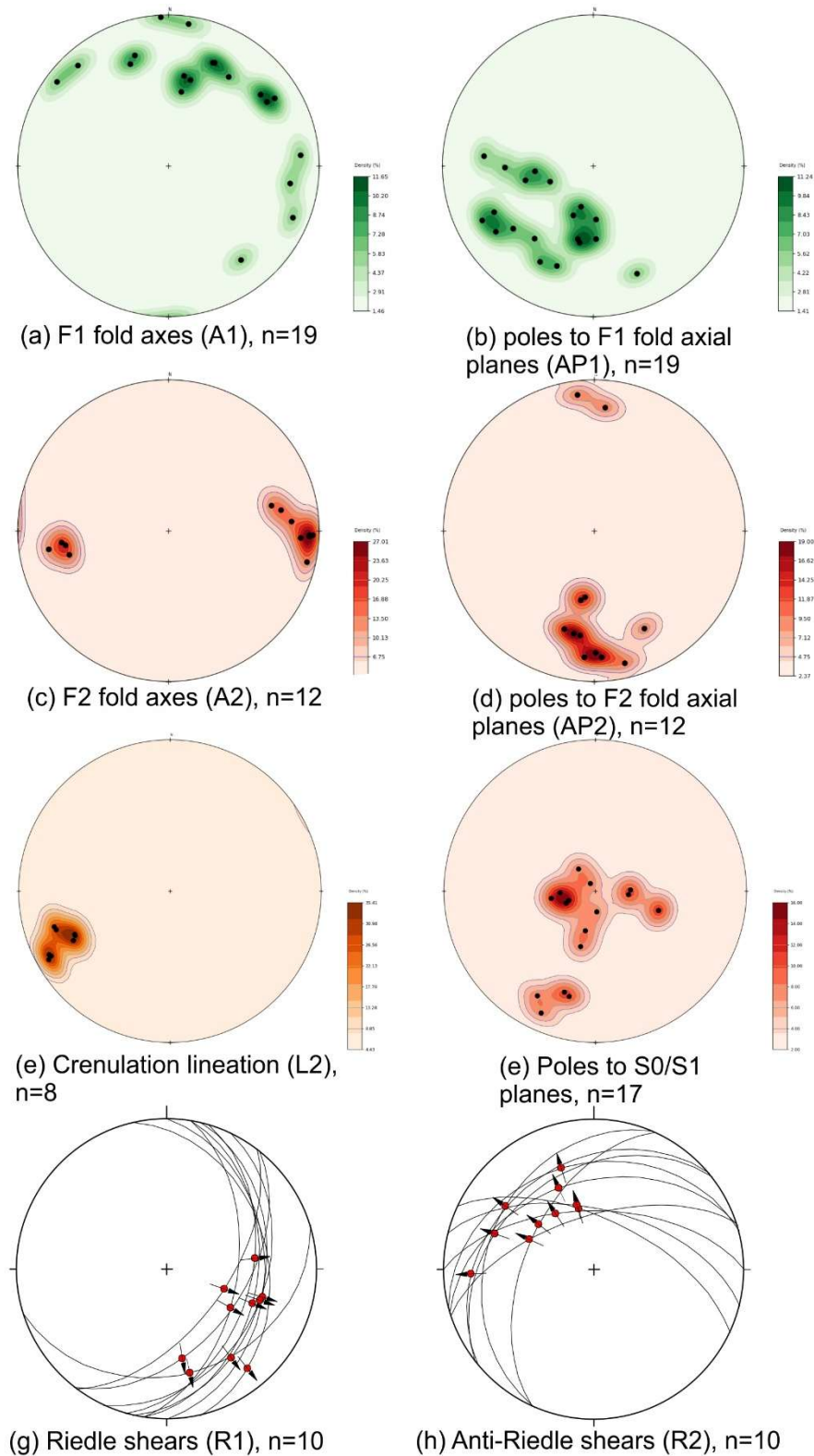


Figure 5-30: Stereographic projections (lower hemisphere, Equal area net) at Caprioli. (a) F1 fold axes; (b) Poles to F1 fold axial planes. (c) F2 fold axes; (d) Poles to F2 fold axial planes. (e) L2 crenulation lineation. (f) Poles to S0/S1 planes. (g) Riedle shears. (h) Anri-Riedle shears.

5.5.3 Microstructures

5.5.3.1 Optical microscope

Thin-section analyses of the slate samples reveal a clear overprinting relationship between the early S1 cleavage and the later F2 folding and associated S2 crenulation cleavage. The S1 foliation, developed during the D1 deformation phase, is defined by an alternation of white-mica-rich domains and quartz-rich layer, reflecting compositional layering enhanced by cleavage-parallel dissolution and mica alignment (Figure 5-31 a). This earlier S1 fabric is variably folded by F2 microfolds, which generate a superposed crenulation cleavage S2. The intensity and geometry of S2 vary depending on strain levels. In weakly deformed domains, the S2 cleavage is only incipient and marked by thin, discontinuous films of white mica that outline the hinges of minor F2 folds (Figure 5-31 b, c, d). In highly deformed slates, the crenulation cleavage S2 is well developed, formed by cleavage domains and microlithons where relics of the previous S1 are present (Figure 5-31 e, f). Frequently, the S2 cleavage is anastomosing, forming a complex array of surfaces (Figure 5-31 f). In finer slates, the F2 crenulation shows a kink/chevron geometry, marked also by folded thin veins parallel to the S1 cleavage.

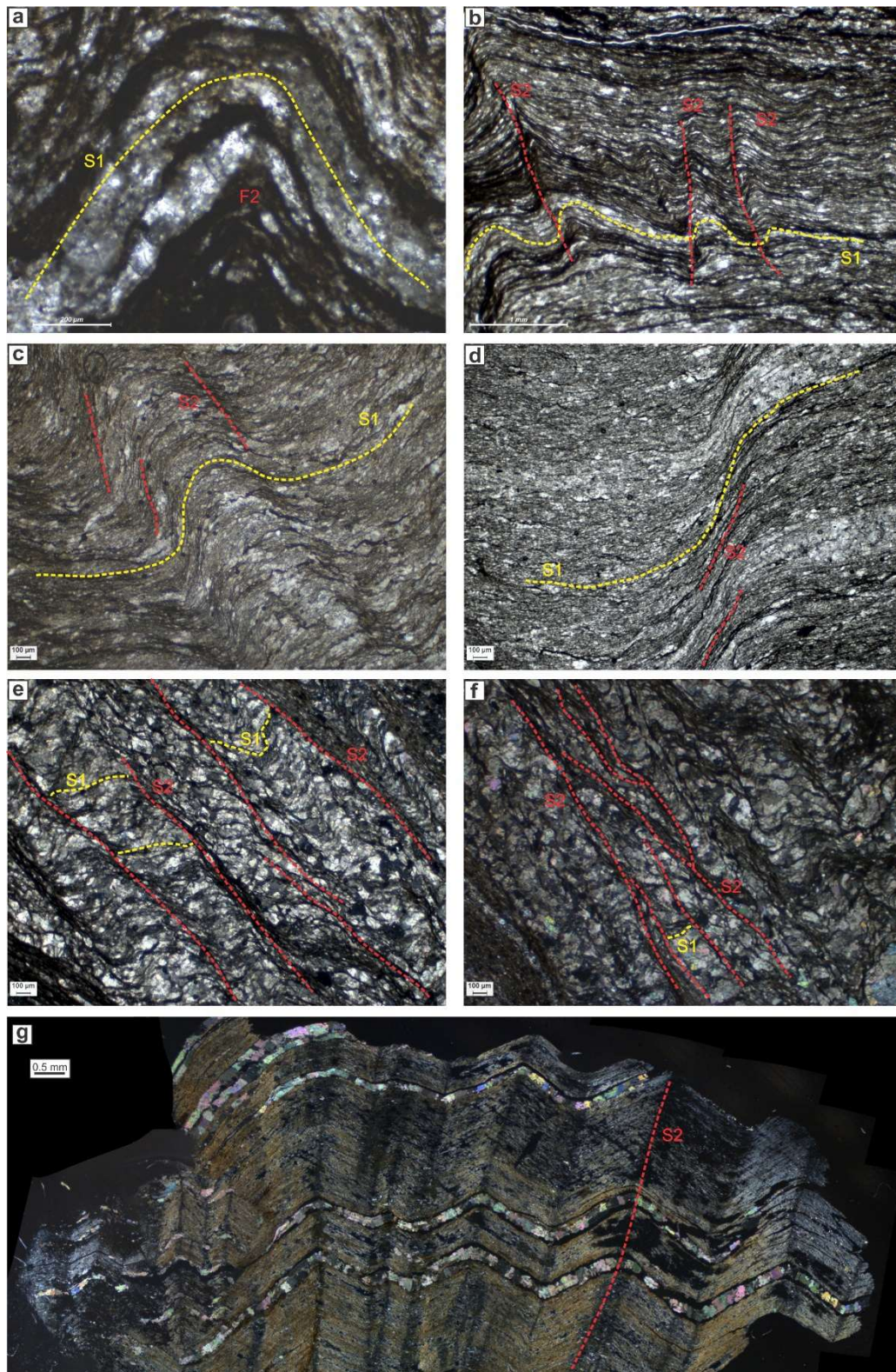


Figure 5-31: Microphotographs of thin sections by optical microscope. (a) F2 fold deforming a previous S1 disjunctive cleavage. (b, c, d) Examples of F2 crenulation with a weak development of S2 cleavage. (e, f) Highly deformed slate with a penetrative crenulation cleavage S2. (g) Crenulation with a chevron geometry.

5.5.4 SEM

SEM observations further clarify the overprinting relationships among the deformation fabrics in the slates (Figure 5-32). The S2 foliation appears as a well-defined crenulation cleavage that folds and reorients the earlier S1 disjunctive cleavage. Within individual microlithons, S1 is preserved as a spaced cleavage and is locally wrapped into intrafolial F2 microfolds (Figure 5-32 a, g), demonstrating the progressive development of S2 as the axial-plane cleavage to the F2 crenulations. These relationships confirm that S2 formed during the D2 deformation phase and is a superposed fabric overprinting S1. Along sections parallel to the F2 crenulation axis, F3 micro-folds can be recognized (Figure 5-32 h).

The matrix contains abundant porphyroclasts of quartz, carbonate minerals, and Fe-oxides, which behave as competent rigid grains around which the S2 cleavage is deflected. Their presence contributes to strain partitioning and results in the heterogeneous distribution of cleavage domains and microlithons.

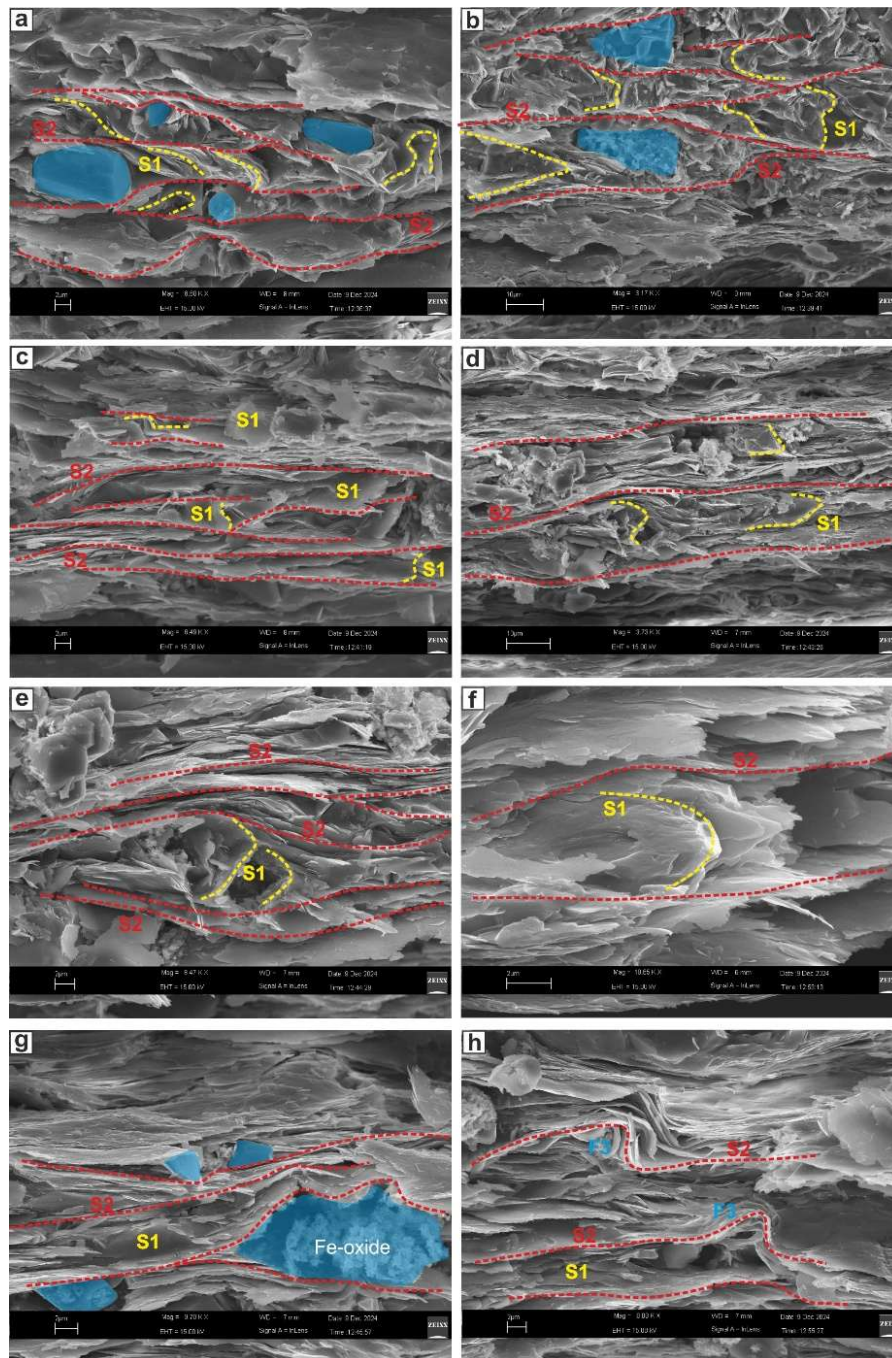


Figure 5-32: (a-g) S2 crenulation cleavage deforming a S1 disjunctive cleavage forming intrafolial folds within microlithons. Porphyroclasts of quartz, carbonates and Fe-oxides are present. (h) S2 foliation folded by F3 micro-folds.

5.6 Scan Electron Microscopy (EDS)

5.6.1 White-Mica composition

Figure 5-33 a, shows the ternary classification diagram plots of white mica compositions in terms of end-member proportions. The analysed samples cluster mainly along the muscovite–paragonite join, with some compositions extending into the phengitic field. Most analyses

avoid the compositional gap, though a few points define a spread between muscovite and paragonite.

The binary plot (Figure 5-33 b) of K apfu versus Na apfu highlights the substitution between muscovite and paragonite. The data define two main clusters: a K-rich muscovite–phengite group and a Na-rich paragonitic group, separated by a compositional gap.

Figure 5-33 c, illustrates the relationship between K content and the degree of Tschermak/phengitic substitution (represented by Si/Al_{tot}). The majority of samples plot within the muscovite–phengite field, trending toward higher Si and lower K contents. A subset of analyses trends toward paragonite compositions. This distribution confirms the coexistence of muscovite, paragonite, and phengitic white micas within the samples.

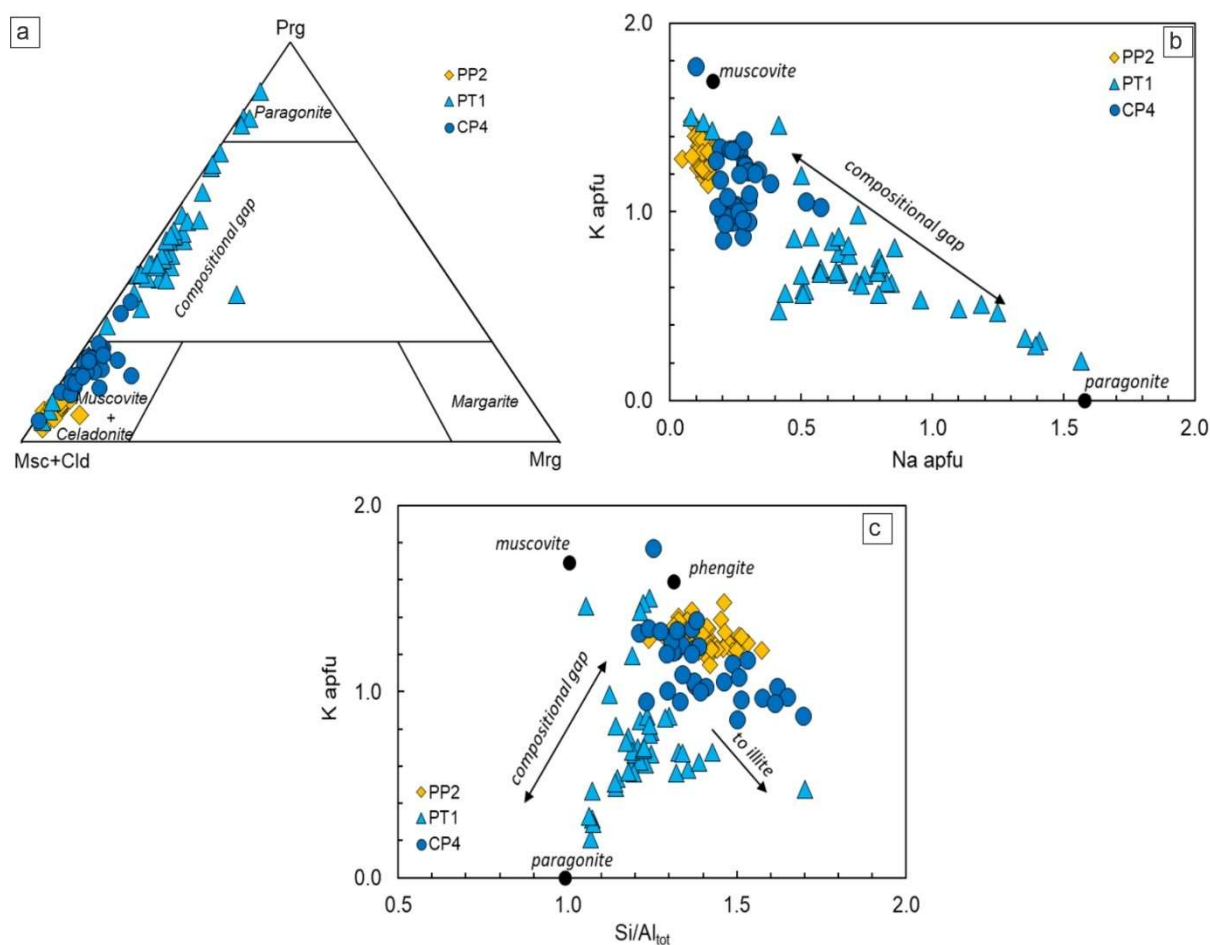


Figure 5-33: SEM analysis: (a) Ternary classification diagram of white mica compositions (muscovite–paragonite–margarite). (b) K apfu vs. Na apfu diagram for white mica. (c) K apfu vs. Si/Al_{tot} diagram.

5.6.2 Chlorite composition

Figure 5-34 a, shows the plots of the chlorite compositions using Fe against Si (apfu). The analyzed samples fall predominantly within the fields of clinochlore, ripidolite, and pycnochlorite, with minor scatter toward chamosite. This distribution indicates Mg–Fe rich trioctahedral chlorites.

Figure 5-34 b, indicates the relationship between Si and divalent cations ($\text{Mg} + \text{Fe}^{2+}$), with reference lines indicating constant Al^{3+} contents. The samples cluster in the clinochlore–ripidolite–sheridanite fields, suggesting limited Tschermak substitution.

The ternary classification (Figure 5-34 c) highlights the relative proportions of Al, Mg, and Fe. The data points plot entirely within the Type I field, confirming the presence of trioctahedral chlorites. Their position along the Mg–Fe join reflects variable Mg/Fe ratios, with moderate Al contents.

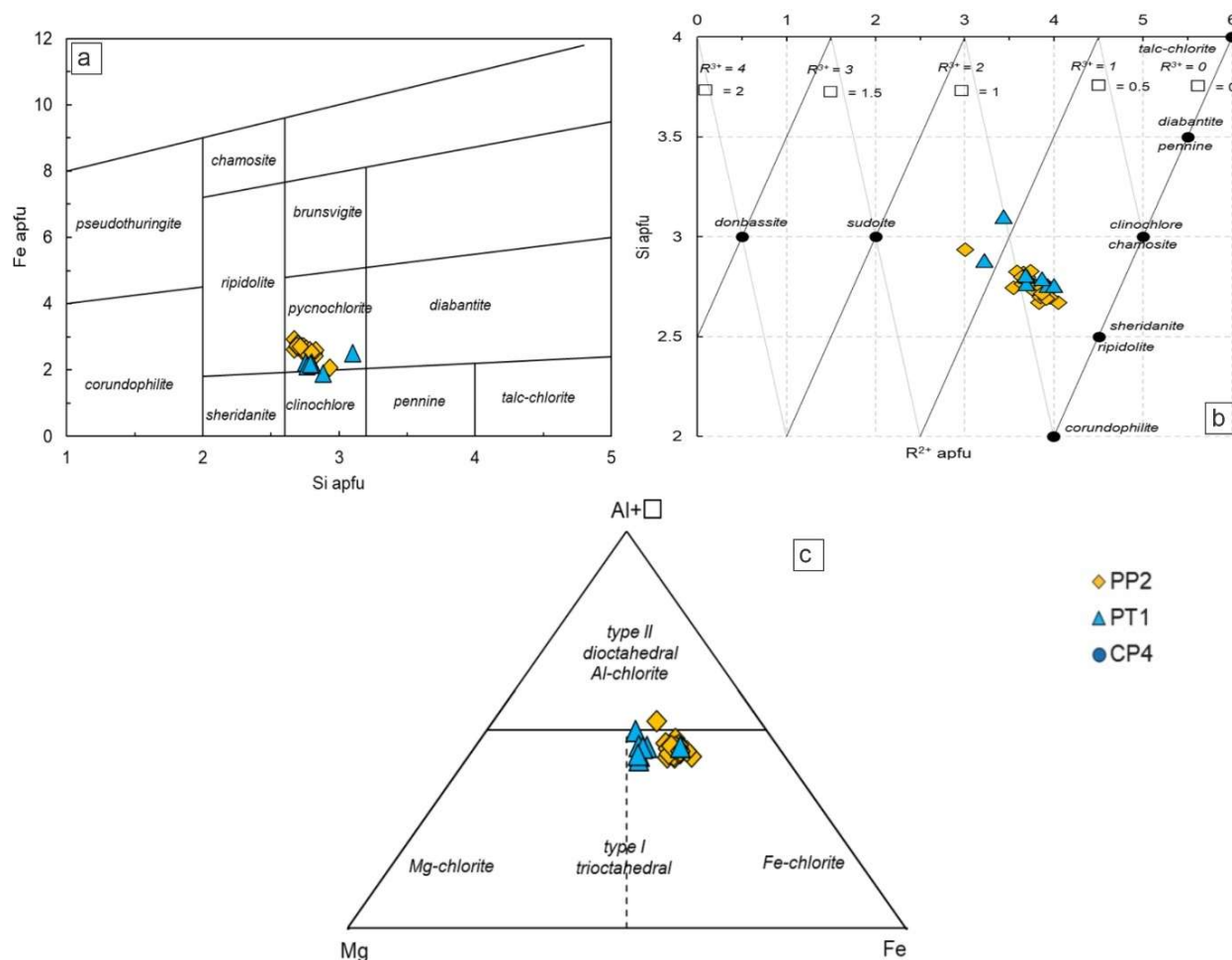


Figure 5-34: SEM analysis (a) Classification diagram of chlorite based on Fe vs. Si (apfu). (b) Si apfu vs. R^{2+} apfu (Mg + Fe²⁺) diagram illustrating the degree of Tschermak substitution in chlorite. (c) Ternary Al–Mg–Fe classification diagram for chlorite.

5.7 Electron microprobe analysis (EPMA-WDS)

5.7.1 White-Mica composition

The ternary classification plot (Figure 5-35 a) shows the proportions of white mica endmembers (Table 5.1). The analysed samples mostly fall along the muscovite–paragonite join, with some extending into the celadonic substitution field. A well-defined compositional gap separates muscovite-rich from paragonite-rich analyses.

Figure 5-35 b, highlights the Tschermak and celadonic substitutions in white mica. Most analyses cluster around muscovite compositions, with Si values between 6.0–7.0 apfu and corresponding Al contents. Several analyses trend toward celadonic substitution (higher Si, lower Al), while only minor scatter is observed in the siderophyllite and eastonite fields.

The binary plot (Figure 5-35 c) illustrates the substitution between K-rich (muscovite–phengite) and Na (paragonite). The data define two clear fields: K-rich muscovite–phengite and Na-rich paragonite, separated by a compositional gap.

Figure 5-35 d, shows the evolution of white mica compositions in relation to phengitic substitution. Most samples fall within the muscovite–phengite field, trending toward higher Si and lower K. A subset of analyses trends toward illitic compositions, while others point toward paragonite. This distribution documents the coexistence of muscovite, paragonite, and phengite.

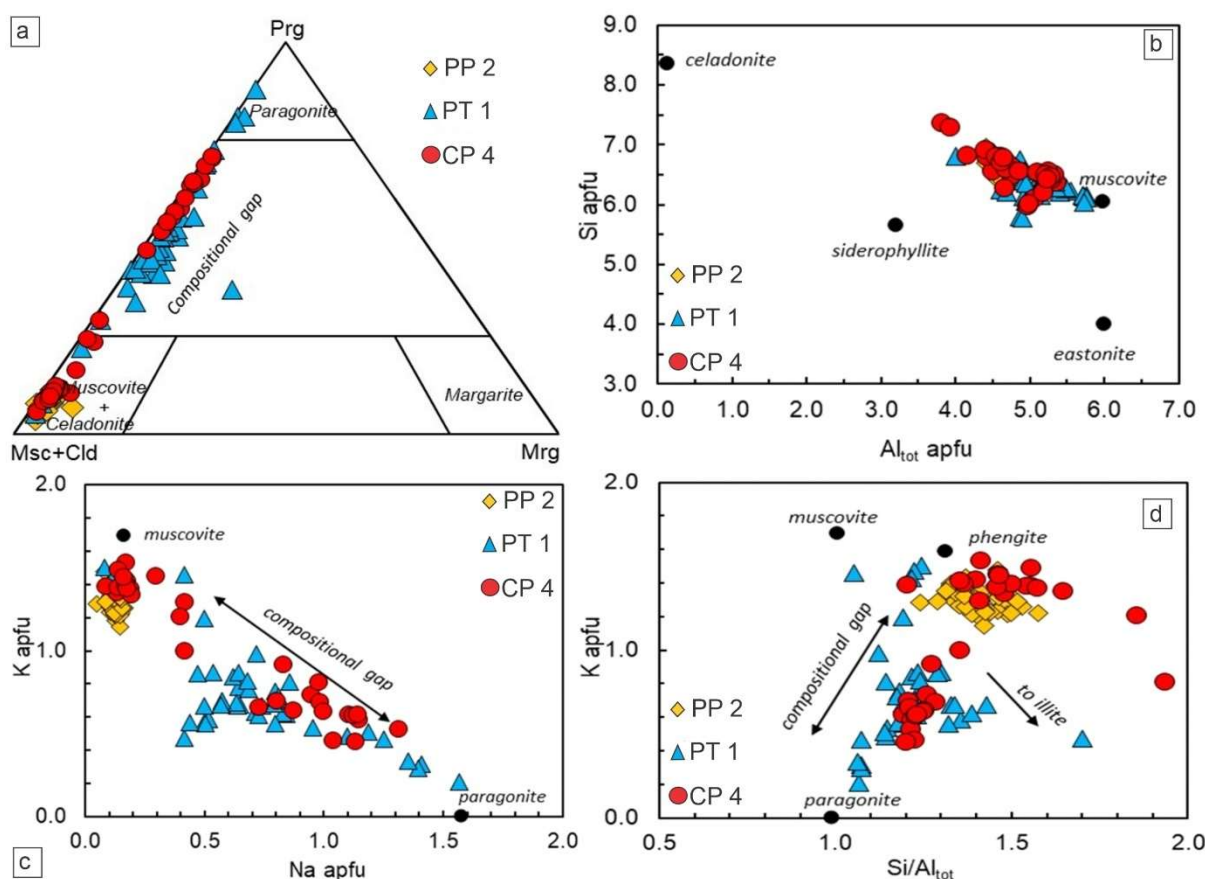


Figure 5-35: EPMA analysis (A) Ternary classification diagram of white mica compositions (muscovite–paragonite–margarite). (B) Si apfu vs. Al_{tot} diagram for white mica. (C) K apfu vs. Na apfu diagram illustrating the muscovite–paragonite substitution. (D) K apfu vs. Si/Al_{tot} diagram for white mica.

Table 5-1: Mineralogical composition of white mica based on EPMA.

Spot No.	MnO	FeO	CaO	TiO ₂	Na ₂ O	MgO	K ₂ O	SO ₃	SiO ₂	Al ₂ O ₃	Total
1	0.0	3.0	1.0	4.7	0.6	6.5	7.7	0.4	44.7	27.0	95.6

2	0.0	3.4	0.3	0.4	0.7	6.2	8.6	0.3	45.4	30.4	95.8
3	0.0	2.7	0.6	1.1	0.8	5.0	9.3	0.1	46.5	30.9	96.9
4	0.0	0.7	0.6	0.1	0.8	2.5	11.7	0.2	46.8	33.1	96.4
5	0.0	1.0	0.5	0.4	0.8	3.4	10.1	0.1	50.2	30.5	97.0
6	0.0	1.1	0.9	1.4	1.0	3.9	8.9	0.1	48.9	30.5	96.8
7	0.0	3.4	0.6	1.1	0.6	6.7	6.6	0.1	46.6	29.0	94.9
8	0.1	1.2	0.5	0.3	0.4	3.5	12.6	0.0	46.2	30.9	95.7
9	0.0	1.6	0.5	0.0	0.1	3.7	11.1	0.2	45.9	32.4	95.5
10	0.0	2.6	0.6	1.0	0.9	6.8	7.6	0.0	46.2	28.8	94.3
11	0.1	1.6	3.5	1.3	0.6	4.9	8.9	0.2	45.5	28.5	95.0
12	0.1	3.4	0.8	2.1	0.7	6.8	7.3	0.1	46.6	29.2	96.8
13	0.0	1.6	0.8	1.0	0.2	1.2	9.1	0.1	49.9	32.7	96.7
14	0.1	2.5	0.6	2.3	0.5	6.0	8.8	0.4	45.5	28.9	95.5
15	0.0	1.8	0.9	1.5	0.6	5.0	9.0	0.0	46.3	30.5	95.6
16	0.0	1.2	2.2	0.5	0.4	2.9	5.3	0.3	52.7	29.8	95.4
17	0.0	1.9	1.2	1.2	0.7	4.9	8.7	0.0	49.2	27.7	95.6
18	0.0	1.3	0.6	1.0	0.4	4.8	10.1	0.1	47.5	29.0	94.8
19	0.0	3.2	0.2	0.8	0.7	6.6	7.8	0.1	49.0	26.6	95.2
20	0.0	2.4	3.4	1.8	0.6	5.2	7.1	0.1	49.2	24.7	94.6
21	0.0	1.7	1.0	1.1	0.8	5.2	9.6	0.1	47.1	29.8	96.4
22	0.1	2.9	1.1	1.3	0.5	5.9	8.7	0.1	45.1	28.7	94.3
23	0.1	4.1	1.3	1.7	0.7	7.6	6.8	0.2	47.3	25.9	95.6
24	0.1	3.8	0.4	1.6	0.4	7.1	9.9	0.2	47.0	26.6	97.0
25	0.1	3.3	0.4	2.0	0.7	6.9	7.0	0.2	48.5	25.0	94.1
26	0.0	3.3	0.6	1.2	0.5	4.9	8.3	0.1	47.5	28.9	95.5
27	0.0	2.3	0.5	0.9	0.8	4.9	9.1	0.1	45.7	29.1	93.5
28	0.1	2.8	0.4	0.4	1.0	6.5	7.7	0.2	49.4	26.8	95.3
29	0.1	4.4	1.0	2.0	0.5	8.1	9.0	0.2	44.8	26.9	96.9
30	0.0	2.2	1.8	1.5	0.7	6.0	8.8	0.1	47.8	27.9	96.6
31	0.0	4.6	0.7	0.8	0.6	8.3	7.4	0.0	44.8	28.8	96.1
32	0.1	3.4	0.6	1.0	0.6	6.5	7.6	0.0	45.6	28.9	94.3
33	0.0	1.0	0.3	1.9	0.5	3.6	9.8	0.1	48.7	28.3	94.3
34	0.0	1.4	1.9	0.6	1.0	4.3	8.5	0.0	48.6	27.8	94.1
35	0.0	1.4	0.4	0.8	0.7	3.7	9.8	0.1	48.2	30.8	96.0
36	0.1	4.3	5.0	1.5	0.5	7.0	6.3	0.0	50.3	20.5	95.3
37	0.0	4.3	0.3	2.3	0.6	7.0	5.9	0.0	54.6	19.9	94.9
38	0.0	3.0	0.8	0.7	0.5	5.5	7.2	0.2	52.6	24.6	95.1
39	0.0	3.2	0.2	2.0	0.6	6.1	7.2	0.1	50.7	24.2	94.3
40	0.0	3.9	0.8	0.3	1.3	7.0	4.6	0.1	52.3	25.4	95.8
41	0.1	1.7	1.2	1.0	0.5	4.5	7.0	0.1	56.0	22.0	93.9
42	0.0	2.0	1.2	1.2	0.6	5.1	6.2	0.2	55.9	23.7	96.1
43	0.1	2.2	3.5	1.0	0.7	5.6	7.4	0.1	51.1	23.9	95.5
44	0.0	2.2	0.8	0.4	0.8	4.4	6.5	0.1	57.1	22.4	94.6
45	0.1	1.2	3.5	1.0	0.6	4.2	8.7	0.2	47.2	28.1	95.0
46	0.0	3.9	2.0	1.0	0.7	7.0	6.2	0.1	50.9	24.6	96.3
47	0.1	3.9	1.3	1.2	0.8	7.0	6.9	0.1	49.0	26.7	96.9

48	0.0	0.9	0.7	0.5	0.7	3.1	10.2	0.1	52.3	28.4	96.9
49	0.0	1.3	0.6	1.6	0.8	3.6	7.3	0.1	55.6	26.0	97.0
50	0.1	3.3	1.2	0.8	0.5	7.5	6.0	0.1	52.9	23.3	95.7
51	0.0	2.7	1.7	2.4	0.4	5.7	6.5	0.1	55.9	22.0	97.4
52	0.0	2.2	2.1	0.5	0.6	4.6	6.3	0.1	58.0	21.1	95.4
53	0.0	1.3	0.3	0.5	0.2	4.7	10.2	0.0	52.9	23.2	93.4
54	0.0	1.0	0.2	2.2	0.8	3.5	8.8	0.0	52.1	26.3	94.9
55	0.1	1.3	0.7	3.0	0.5	3.2	8.9	0.1	53.1	25.7	96.5
56	0.0	2.0	0.3	1.5	1.9	4.4	7.3	0.1	50.1	27.0	94.5
57	0.0	2.9	1.4	0.5	0.7	6.6	6.5	0.1	53.3	22.5	94.6
58	0.0	2.7	0.5	0.5	0.7	4.9	6.5	0.1	53.8	23.7	93.3
59	0.0	1.3	0.6	1.8	0.7	3.8	8.2	0.2	52.3	26.0	95.0
60	0.1	3.8	0.8	1.0	0.6	8.1	7.2	0.1	46.2	27.4	95.3
61	0.1	2.2	0.5	0.9	0.5	5.3	9.7	0.1	49.4	26.8	95.3
62	0.0	1.3	1.6	1.0	0.5	3.3	5.4	0.0	55.6	26.0	94.8
63	0.0	2.6	1.1	2.2	0.6	4.9	7.6	0.1	50.0	25.0	94.2
64	0.0	2.3	0.3	3.7	0.4	4.4	5.3	0.1	53.6	25.5	95.7
65	0.0	4.2	0.3	0.3	0.7	7.8	7.1	0.1	47.6	27.1	95.2
66	0.0	0.9	1.2	1.5	0.6	3.2	6.4	0.2	54.8	26.0	94.7
67	0.1	2.4	3.7	0.5	1.6	6.1	4.1	0.0	51.5	24.8	94.8
68	0.1	1.8	1.2	0.8	0.8	3.4	7.2	1.4	50.3	27.9	94.9
69	0.0	2.9	0.3	0.5	0.7	5.8	6.9	0.0	53.5	25.6	96.3
70	0.0	2.1	0.6	0.9	0.8	4.9	8.0	0.1	52.5	26.5	96.4

5.7.2 Chlorite composition

This diagram of Figure 5-36 a, classifies chlorite compositions based on their Fe and Si contents. The majority of analyses plot within the clinochlore, ripidolite, and pycnochlorite fields, with minor scatter toward chamosite (Table 5.2).

The relationship between Si and divalent cations ($\text{Mg} + \text{Fe}^{2+}$) shows (Figure 5-36 b) that the samples mainly occupy the clinochlore–ripidolite–sheridanite fields. The limited spread along the Al^{3+} substitution lines suggest moderate Tschermak substitution.

Figure 5-36 c, shows the ternary classification, which highlights the relative proportions of Al, Mg, and Fe. cll data points fall within the Type I field, confirming the trioctahedral nature of the chlorites.

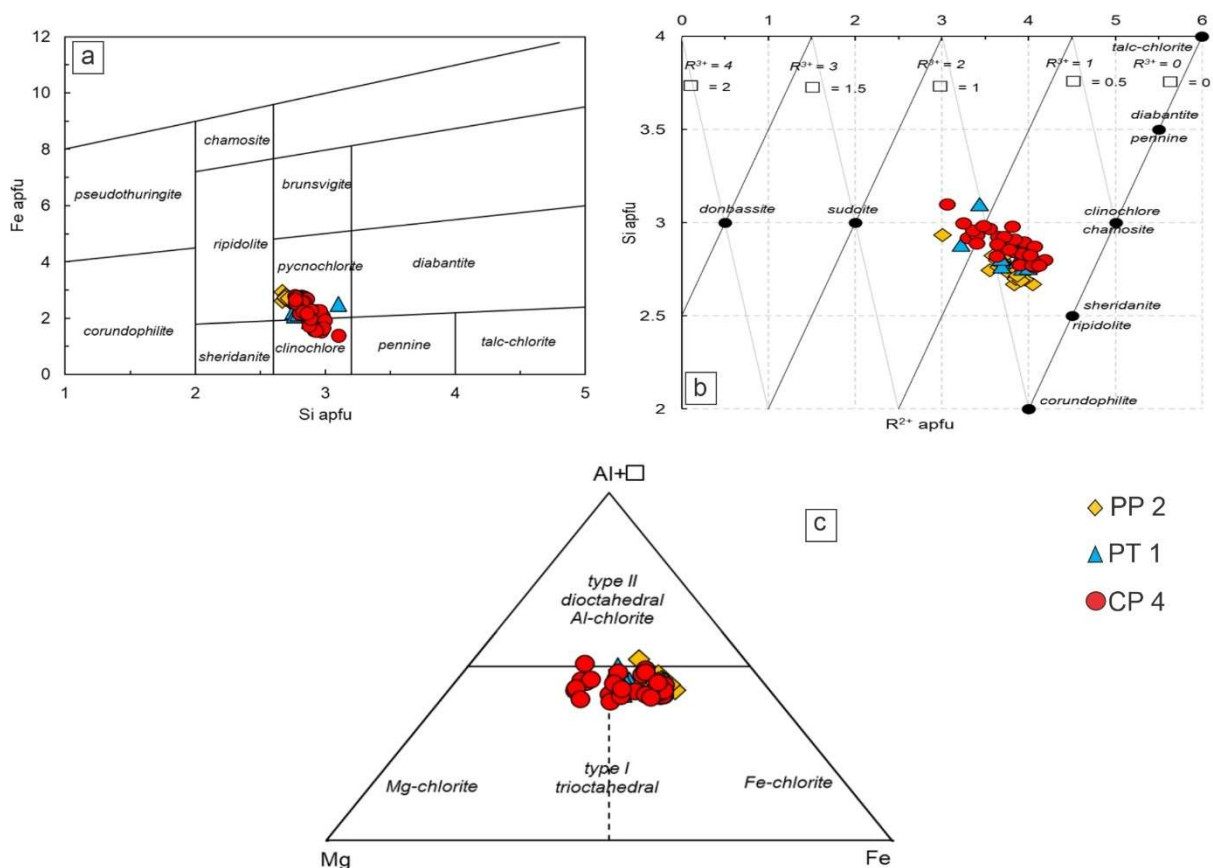


Figure 5-36: EPMA analysis: (a) Fe apfu vs. Si apfu classification diagram for chlorite. (b) Si apfu vs. R^{2+} apfu (Mg + Fe^{2+}) diagram. (c) Al–Mg–Fe ternary classification diagram for chlorite.

Table 5-2: Chlorite composition based on EPMA.

MnO	FeO	CaO	TiO2	Na2O	MgO	K2O	SO3	SiO2	Al2O3	Total
0.2	18.3	0.8	0.0	0.1	14.0	0.5	0.0	32.2	22.5	88.6
0.0	19.5	0.2	0.1	0.2	14.1	0.0	0.0	31.4	21.6	87.2
0.0	18.9	0.3	0.2	0.2	14.7	0.5	0.0	29.7	23.0	87.5
0.1	20.5	0.2	0.1	0.2	15.2	0.4	0.0	29.7	20.5	87.1
0.1	21.6	0.2	0.1	0.1	15.7	0.4	0.0	28.5	20.3	87.0
0.1	20.9	0.5	0.1	0.1	15.7	0.4	0.0	28.1	21.1	87.2
0.1	19.3	0.5	0.2	0.2	15.4	0.8	0.0	30.2	21.5	88.2
0.1	20.7	0.6	0.2	0.2	15.7	0.4	0.0	28.7	21.1	87.6
0.1	20.9	0.5	0.2	0.2	16.0	0.4	0.0	29.0	20.2	87.4
0.1	21.3	0.6	0.1	0.3	15.9	0.8	0.0	29.1	19.8	88.0
0.1	18.8	0.3	0.2	0.2	15.0	0.6	0.0	31.4	22.9	89.5
0.1	18.0	0.3	0.2	0.2	14.4	1.1	0.0	31.3	22.3	87.9
0.2	17.3	0.7	0.1	0.5	12.4	1.2	0.0	34.3	22.0	88.7
0.1	16.5	0.0	0.3	0.3	11.6	2.4	0.0	32.9	22.9	87.1
0.1	12.2	1.2	0.0	0.6	12.1	1.7	0.0	34.3	26.8	89.0
0.1	17.9	1.3	0.0	0.3	13.2	1.7	0.0	32.1	21.0	87.7
0.1	19.2	0.8	0.0	0.2	14.5	0.7	0.0	30.2	22.1	87.9
0.1	21.8	0.6	0.1	0.1	14.9	1.3	0.0	28.9	19.9	87.9
0.3	22.4	0.3	0.1	0.1	15.8	0.4	0.0	29.8	20.1	89.3

0.1	20.7	0.3	0.0	0.3	16.6	1.4	0.0	30.5	19.7	89.6
0.1	19.5	0.5	0.0	0.3	14.9	1.4	0.0	30.4	20.5	87.6
0.1	18.9	0.3	0.0	0.2	15.0	1.5	0.0	31.4	20.9	88.4
0.0	12.6	1.1	0.1	0.5	12.5	1.8	0.0	34.1	25.1	87.8
0.1	18.7	0.2	0.0	0.2	14.3	0.4	0.0	32.1	23.5	89.5
0.1	13.4	0.8	0.2	0.9	10.9	1.8	0.0	36.6	23.6	88.3
0.2	15.2	0.3	0.1	0.7	13.0	1.7	0.0	34.2	24.0	89.4
0.1	15.2	1.0	0.2	0.5	12.6	1.4	0.0	32.0	24.1	87.1
0.1	15.8	0.5	0.3	0.6	12.8	1.3	0.0	32.0	23.9	87.3
0.1	15.6	0.6	0.1	0.7	14.2	1.4	0.0	31.9	23.8	88.5
0.0	14.4	0.4	0.1	0.6	12.3	1.4	0.0	33.8	25.6	88.6
0.1	19.8	1.1	0.1	0.2	16.0	0.0	0.0	30.4	22.0	89.7
0.0	18.5	2.4	0.0	0.1	17.0	0.8	0.1	29.9	20.4	89.2
0.0	20.8	0.5	0.1	0.7	16.1	0.2	1.5	29.6	19.0	88.4
0.1	17.5	0.4	0.2	0.2	15.4	0.5	0.0	32.6	21.8	88.7
0.0	19.0	1.3	0.0	0.8	16.4	1.4	0.0	33.4	17.4	89.8

5.8 Mineral Phases

We analyzed three representative samples from the LAC units in Cilento, specifically selected from the Pioppi shear zone, Punta Telegrafo, and Caprioli outcrops, referred to as PP, PT, and CP. These samples were chosen to investigate the PT conditions and mineralogical assemblages present in the region. To complement the temperature conditions, we conducted Raman spectroscopic studies on the same samples.

For a more comprehensive comparison of the temperature conditions and mineral assemblages, we employed EPMA (Electron Probe Micro-Analysis). Quantitative compositional maps were obtained using the JEOL JXA 8200 electron probe microanalyzer at the National Institute of Geophysics and Volcanology in Rome, following the methodology outlined by (Lanari et al., 2017; Petroccia et al., 2025). The analytical conditions were set at a 15-keV accelerating voltage and a 40-ms dwell time.

Using the EPMA, compositional X-ray maps were generated, identifying minerals such as albite, apatite, rutile, quartz, muscovite, phengite, chlorite, and pyrite. In addition to the compositional maps, several spot analyses were conducted to identify mineral phases outside the mapped areas. The software XMapTools 4 (Lanari et al., 2014; Lanari & Duysterhoeft, 2019; Petroccia et al., 2025) was used for data classification, analytical standardization, and to calculate the mineral structural formulae, as well as to display the relative abundance of cations across the samples.

Across all three maps PP, PT and CP, the dominant mineral phases are muscovite and chlorite are indicative of a low-grade metamorphic environment (Figure 5-37). These minerals form under conditions of moderate pressure and temperature, typical of metapelites. The maps also reveal the presence of accessory minerals such as apatite, rutile, and albite, which occur in smaller quantities. These accessory minerals suggest that additional secondary processes like metamorphic fluid activity or mineral crystallization occurred during the deformation. The overall fabric across the three samples shows foliation, where the mica minerals (especially muscovite) are aligned in parallel planes, a typical feature of metamorphic rocks subjected to directed pressure.

The PP, PT and CP samples all exhibit a similar mineral assemblage, dominated by muscovite and chlorite, with varying amounts of dolomite in the Caprioli sample. The maps reflect a low-grade metamorphic origin, with foliation and fractures indicative of deformation. The presence of dolomite in Caprioli suggests localized alteration, adding complexity to the fabric of the sample.

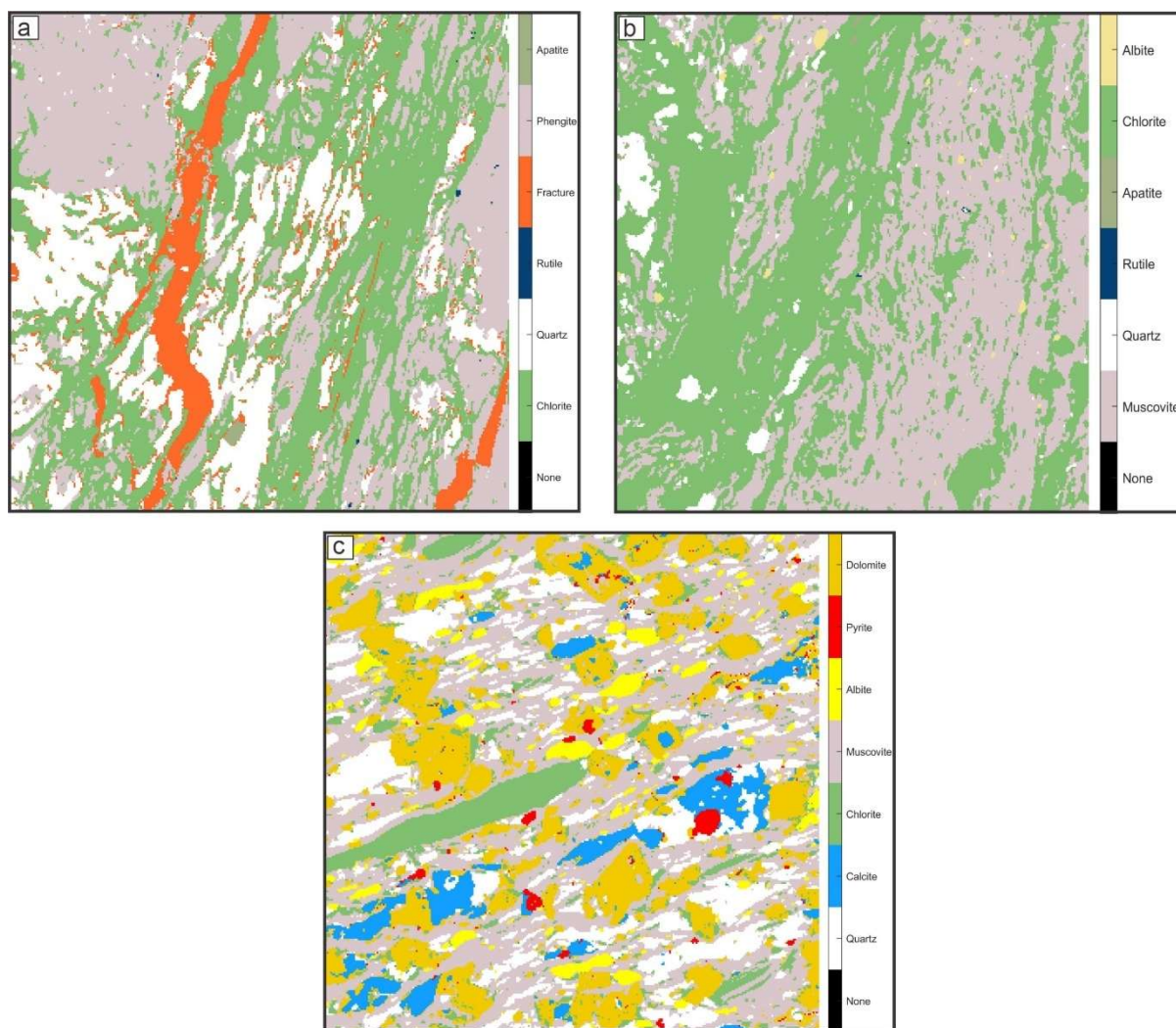


Figure 5-37: EPMA-mineral chemistry based mineral phase maps for three locations (a) Pioppi; (b) Punta Telegrafo; (c) Caprioli.

5.9 Illite crystallinity

The quantitative mineralogical assemblage of the bulk rock is reported in Table 5.3 and Representative XRDP patterns are shown in Figure 5-38. The analytical results show that the samples are mainly characterized by phyllosilicates ranging from 29 to 79 %, subordinately quartz (from 17.5 to 39.5 %) and feldspars (k-feldspar + plagioclase) generally in some percentage units (from 1 to 3.5 %), except in the PP5 sample, where it shows a concentration of 6%. Calcite is present in low percentages (ranging from 0 to 9%) except in PT4, PT5, PT6, and CPP2, which show high amounts (ranging from 23.5 to 43%). Anatase is present in very small amounts (<1%). Phyllosilicates consist of illite-muscovite (10-55 wt.%), chlorite plus mixed-layer chlorite/smectite (C/S) (9-34 wt.%). Paragonite (plus K/Na mica) is present

generally in very low amounts (2-7 %) except in the PT1, PT2, PT3 and CPP3, where it shows higher values. (from 10 to 15%) (Table 5-3; Figure 5-38).

The semi-quantitative mineralogical analysis of the clay fraction shows the main presence of illite-muscovite (from 31 to 74 %). Chlorite follows in abundance, with a value from 10 to 33%. Mixed layers chlorite/smectite (C/S) are detectable in lower amounts only in two samples (PT1 and PT2), while paragonite plus K/Na mica is higher, especially in the samples collected from Punta Telegrafo site (from 12 to 33%) (Table 5.3; Figure 5-39). KI value varies from 0.42 to 0.38 in the Pioppi shales, from 0.33 to 0.30 in the Caprioli area, and intermediate values (from 0.38 to 0.34) are measured in samples collected from the Punta Telegrafo area.

Table 5-3: Qtz, quartz; K-feld, alkali feldspar; Pl, plagioclase; Ant, anatase; Cal, calcite; Ill/Mus, illite-muscovite; Pg, Paragonite; K/Na mica, White K-Na mica; Chl, chlorite; C/S, mixed layers chlorite/smectite. Kubler Index (KI) and temperature estimates are reported. The temperature estimates (in bold).

Sample	Bulk sample										Clay fraction						Kübler Index and Temperature estimates		
	Qtz	Pl	K-feld	Cal	Ant	Ill+Ms	I/S R3	K/Na-Mica + Pag	Chl + C/S	Σ Phy	Ill + Ms	I/S R3	K/Na-Mica + Pag	Chl	C/S	KI (Basel)	T °C	T °C	
PP1	17.5	2.0	1.0	5.0	0.5	55.0	0.0	2.5	16.5	74.0	67	12	5	16	0	0.38	220	210 ± 10	
PP2	22.5	2.0	1.0	9.0	0.5	37.0	0.0	2.0	26.0	65.0	64	8	5	24	0	0.41	200		
PP3	19.5	2.5	0.5	2.5	1.0	52.0	0.0	2.0	20.0	74.0	73	10	5	12	0	0.40	210		
PP4	36.5	2.5	1.0	4.0	0.5	22.0	7.0	2.5	24.0	55.5	63	7	4	25	0	0.42	200		
PP5	28.0	5.5	1.0	1.0	0.5	34.0	0.0	2.0	28.0	64.0	74	4	5	16	0	0.38	220		
PT1	29.5	2.0	tr	8.5	0.8	25.0	0.0	15.0	19.0	59.0	48	0	25	19	8	0.38	220	235 ± 10	
PT2	23.0	2.0	1.0	5.5	1.0	26.5	0.0	13.0	28.0	67.5	33	0	26	33	8	0.36	230		
PT3	23.0	1.0	1.0	3.0	1.0	41.0	0.0	10.0	20.0	71.0	58	0	32	10	0	0.34	240		
PT4	23.5	0.5	0.5	41.5	tr	20.0	0.0	2.0	12.0	34.0	59	0	19	22	0	0.34	240		
PT5	32.0	0.5	0.5	38.0	tr	17.0	0.0	3.0	9.0	29.0	66	0	22	12	0	0.34	240		
PT6	20.0	1.5	0.5	43.0	tr	10.0	4.0	6.0	15.0	35.0	31	6	36	27	0	0.34	240	250 ± 10	
CPP1	23.0	2.5	0.5	3.5	0.5	41.0	0.0	6.0	23.0	70.0	66	0	8	27	0	0.30	260		
CPP2	21.5	2.0	0.5	22.5	0.5	26.0	0.0	7.0	20.0	53.0	73	0	8	19	0	0.31	250		
CPP3	20.5	0.0	0.0	0.0	0.5	22.0	10.0	13.0	34.0	79.0	38	7	43	13	0	0.33	240		

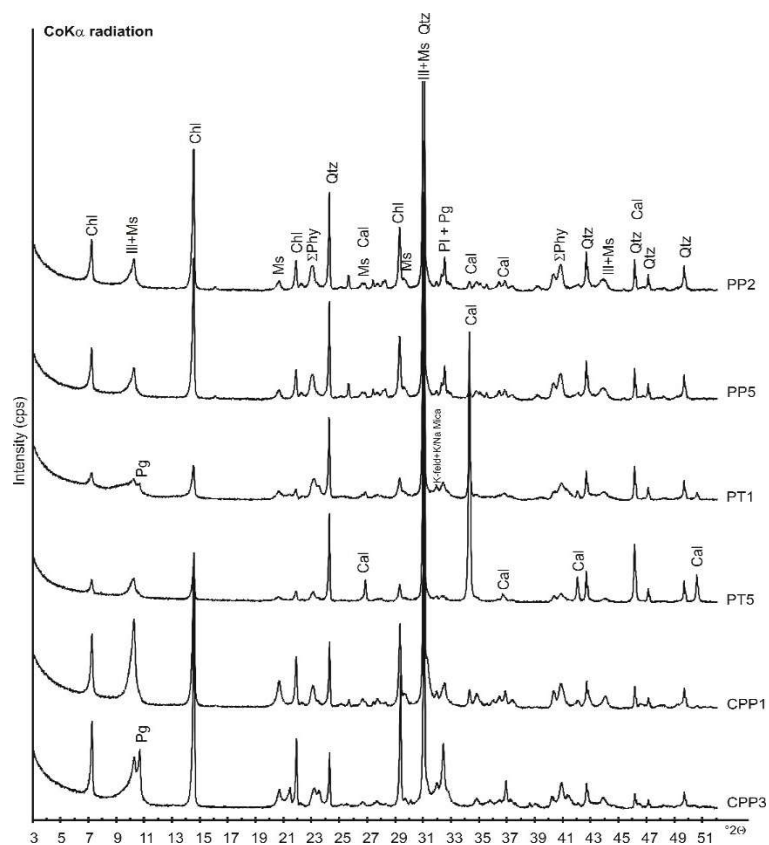


Figure 5-38: Representative XRD patterns of bulk sample. Qtz, quartz; Cal, calcite; Pl, plagioclases; K-felds, k-feldspars; Ill-Ms, illite-muscovite; I/S, mixed layers illite/smectite; Chl, chlorite; C/S, mixed layers chlorite/smectite; Pg, paragonite; K/Na mica, White K-Na mica.

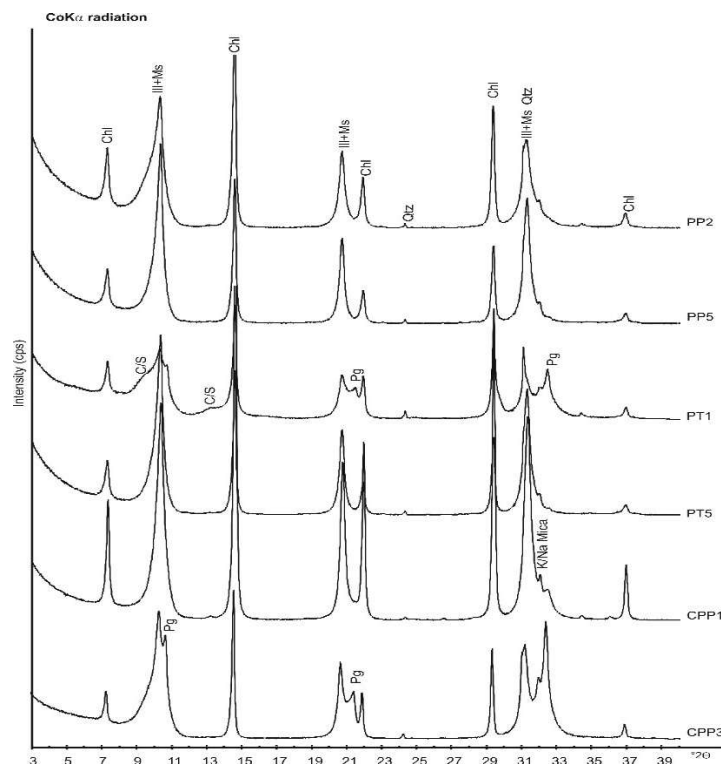


Figure 5-39: Representative XRD patterns of clay fraction. Ill-Ms, illite-muscovite; I/S, mixed layers illite/smectite; Pg, paragonite; K/Na mica, White K-Na mica Chl, chlorite; C/S, mixed layers chlorite/smectite.

5.10 Raman spectroscopy

Four samples (PP1X, PP2X, PP4 and PP5) of Crete Nere slates were analysed. Results (Figure 5-40) indicate that the mean temperatures T1 and T2 range between 238 (PP2X) and 263 (PP1X), and between 231 (PP2X) and 256 (PP1X), respectively. Data are generally dispersed, but the average values of temperatures are consistent between them. On each box, the central mark indicates the median, and the bottom and top edges of the box indicate the 25th and 75th percentiles, respectively.

The whiskers extend to the most extreme data points, excluding those considered outliers, and the outliers are plotted individually using the '+' marker symbol.

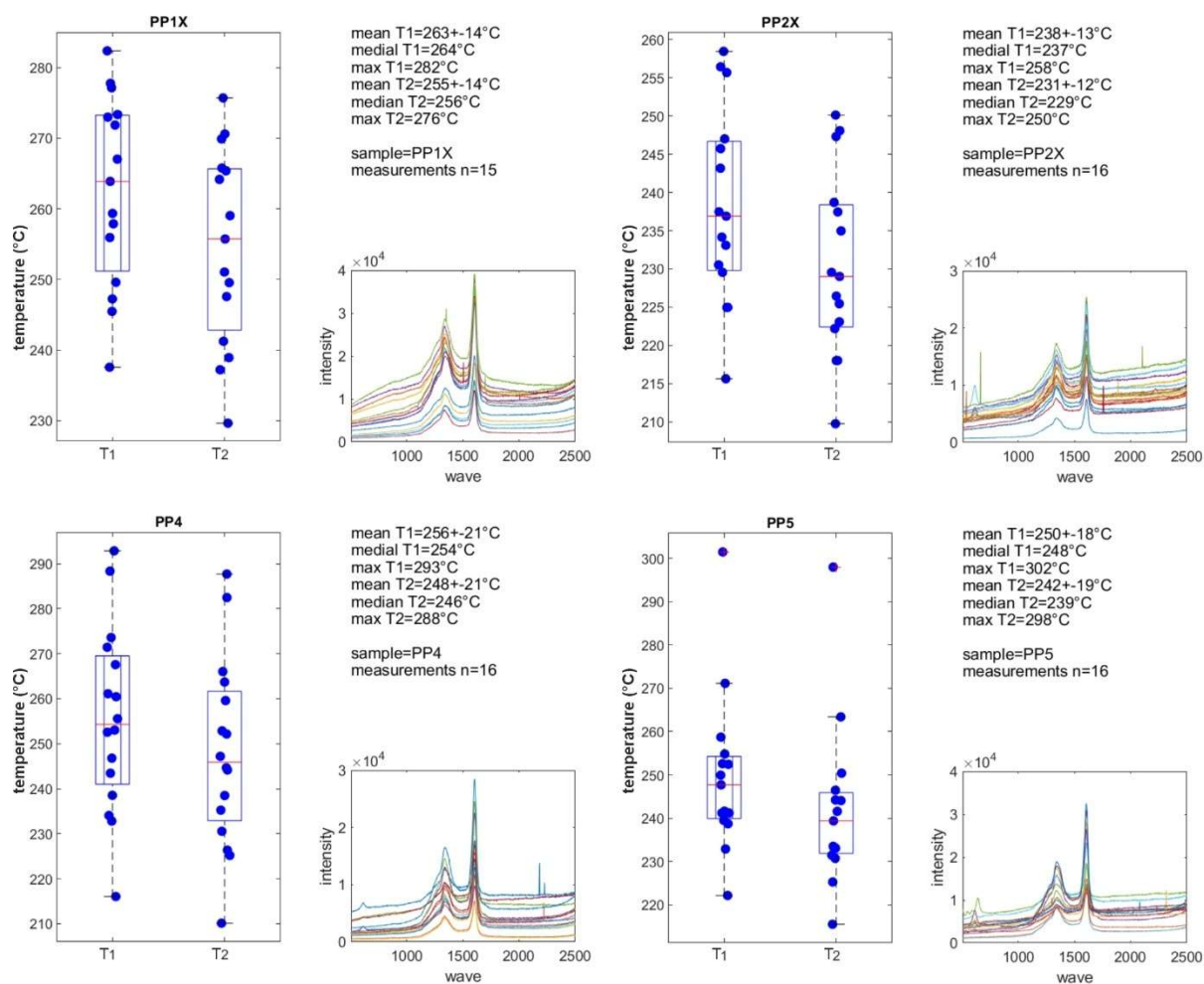


Figure 5-40: Temperature estimates (T1 and T2) from four samples of the Crete Nere slates (PP1X, PP2X, PP4, PP5). Mean values range from 238°C (T1, PP2X) to 263°C (T1, PP1X) and from 231°C (T2, PP2X) to 256°C (T2, PP1X).

A comparative statistical analysis was carried out for the four analysed samples, as shown in Figure 5-41 a. The error bars overlap, suggesting data consistency. Data were combined, and frequency histograms were prepared for T1 and T2 (Figure 5-41, b). The mean values calculated for all data indicate temperatures of 252 ± 19 °C (T1) and 244 ± 19 °C (T2). It is worth noting that both histograms show a Gaussian distribution.

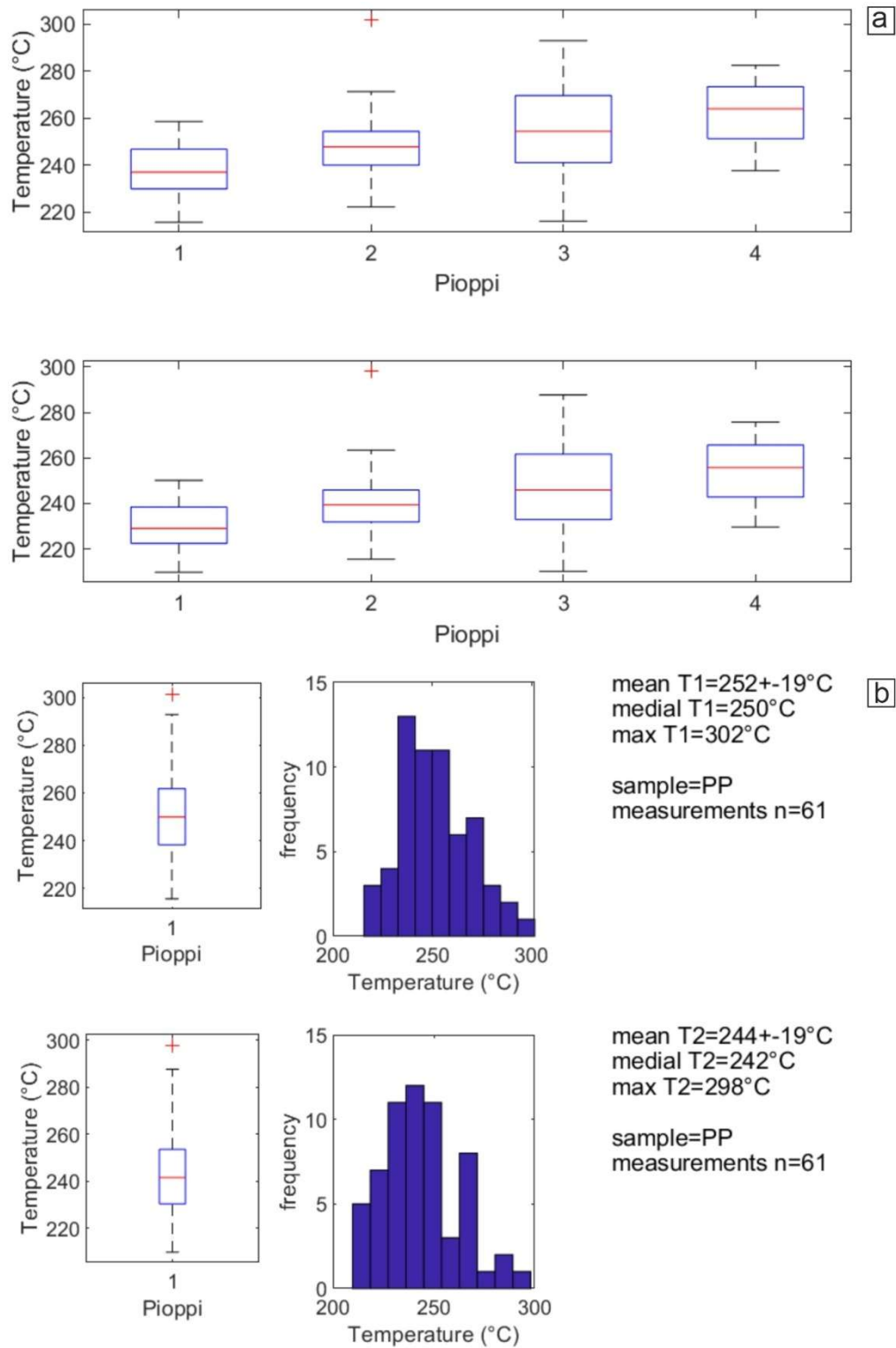


Figure 5-41: (a) Statistical comparison of T1 and T2 for four analysed samples, with overlapping error bars indicating data consistency. (b) Frequency histograms of the combined dataset, both showing near-Gaussian distributions with mean values of 252 ± 19 °C (T1) and 244 ± 19 °C (T2).

5.10.2 Punta Telegrafo

It was analysed seven samples (PT1, PT1X, PT2X, PT3, PT4, PT5 and PT6) of Crete Nere Formation. Results (Figure 5-42 and Figure 5-43) indicate mean temperatures T1 and T2 ranging between 252 (PT4) to 277 (PT1X) and 269 (PPT4) and 245 (PT1X), respectively.

Data are generally dispersed, but the average temperature values are consistent across them. On each box, the central mark indicates the median, and the bottom and top edges of the box indicate the 25th and 75th percentiles, respectively.

The whiskers extend to the most extreme data points, excluding outliers, and the outliers are plotted individually as '+' markers.

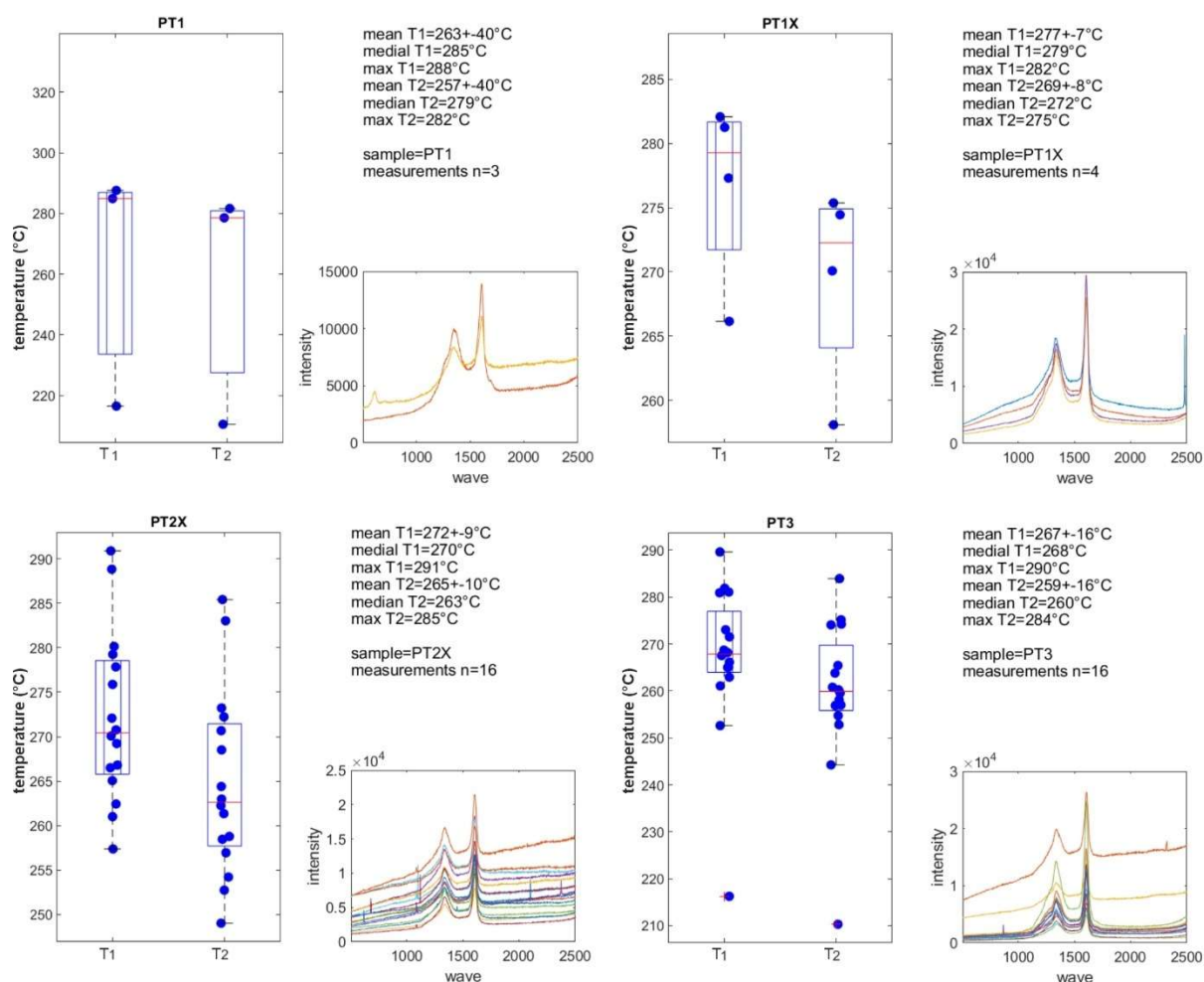


Figure 5-42: T1 temperatures from seven samples of the Crete Nere Formation (PT1, PT1X, PT2X, PT3).

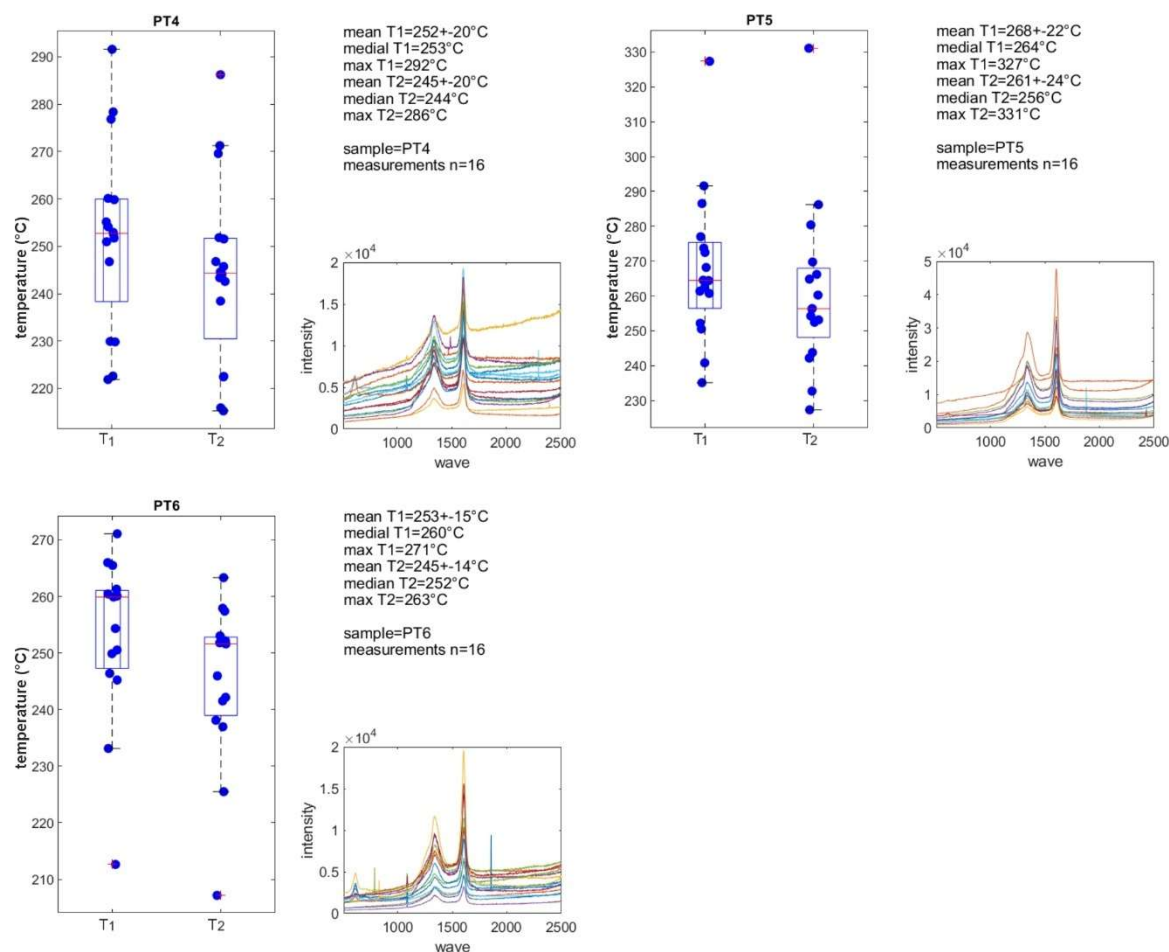


Figure 5-43: T2 temperatures from the same seven samples of the Crete Nere Formation.

Figure 5-44 a, shows a comparison between statistical data for the four samples analysed. The error bars overlap, suggesting data consistency.

Data were joined altogether, and frequency histograms were prepared for T1 and T2 (Figure 5-44 b). The mean values calculated for all data indicate temperatures of 263 ± 18 $^{\circ}$ C (T1) and 255 ± 19 $^{\circ}$ C (T2). It is worth noting that both histograms show a Gaussian distribution.

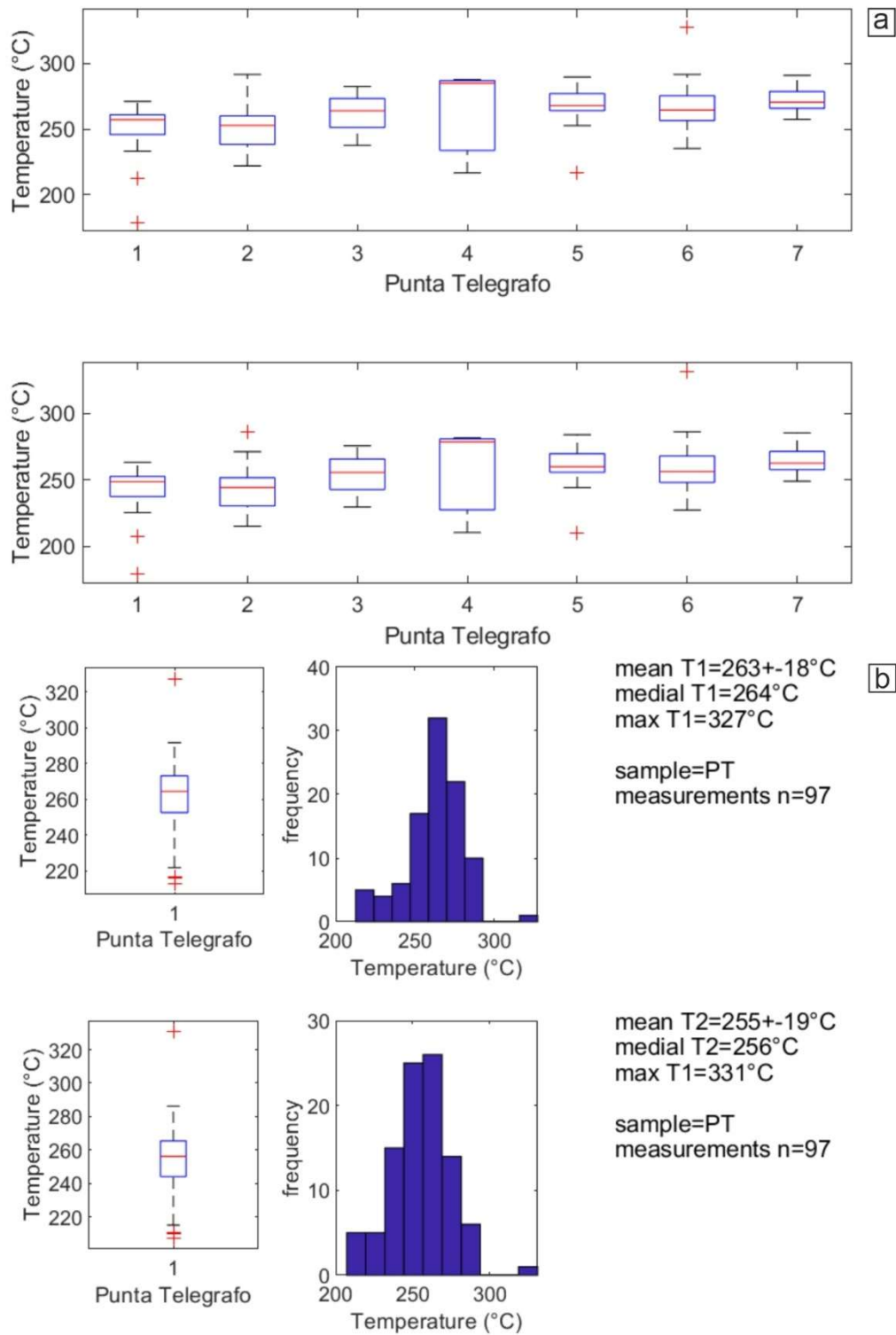


Figure 5-44: (a) Statistical comparison of T1 and T2 for four analysed samples, with overlapping error bars indicating consistent results; (b) Frequency histograms of the combined dataset, both displaying near-Gaussian distributions with mean values of 263 ± 18 °C (T1) and 255 ± 19 °C (T2).

Four samples (CP1X, CP2, CP2X, CP3X) of the Crete Nere Formation were analysed. Results Figure 5-45 indicate that the mean temperatures T1 and T2 range between 289 (CP3X) and 256 (CP2), and between 283 (CP3X) and 249 (CP2), respectively.

Data are generally dispersed, but the average values of temperatures are consistent between them. On each box, the central mark indicates the median, and the bottom and top edges of the box indicate the 25th and 75th percentiles, respectively.

The whiskers extend to the most extreme data points, excluding those considered outliers, and the outliers are plotted individually using the '+' marker symbol.

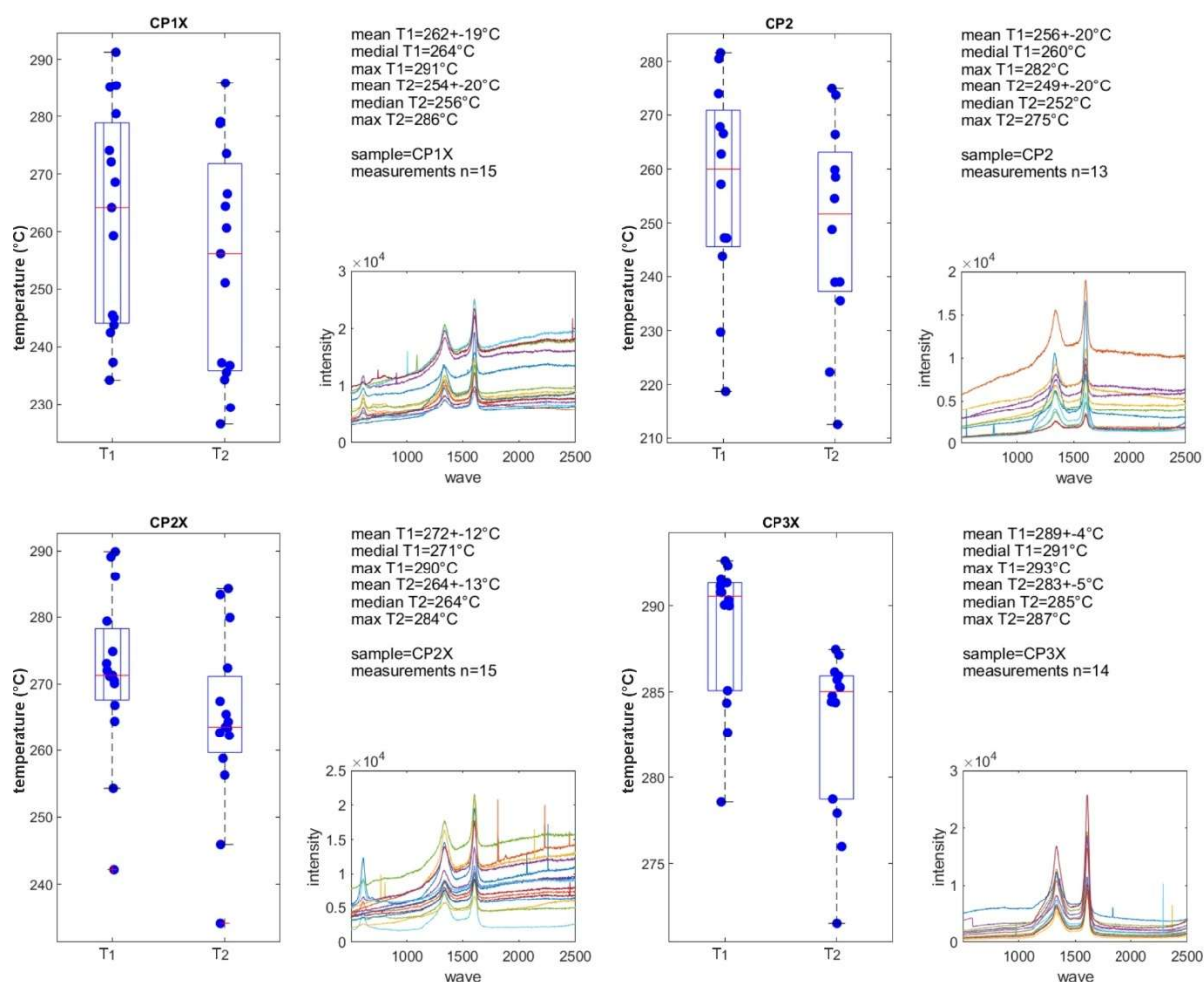


Figure 5-45: Temperature estimates (T1 and T2) from four samples of the Crete Nere Formation (CP1X, CP2, CP2X, CP3X). Mean values range from 289 $^{\circ}$ C (T1, CP3X) to 256 $^{\circ}$ C (T1, CP2) and from 283 $^{\circ}$ C (T2, CP3X) to 249 $^{\circ}$ C (T2, CP2).

Figure 5-46 (a) shows a comparison between statistical data for the four samples analysed. The error bars overlap, suggesting data consistency.

Data were joined altogether, and frequency histograms were prepared for T1 and T2 (Figure 5-46 (b)). The mean values calculated for all data indicate temperatures of 270 ± 19 °C (T1) and 263 ± 20 °C (T2). It is worth noting that both histograms show a Gaussian distribution.

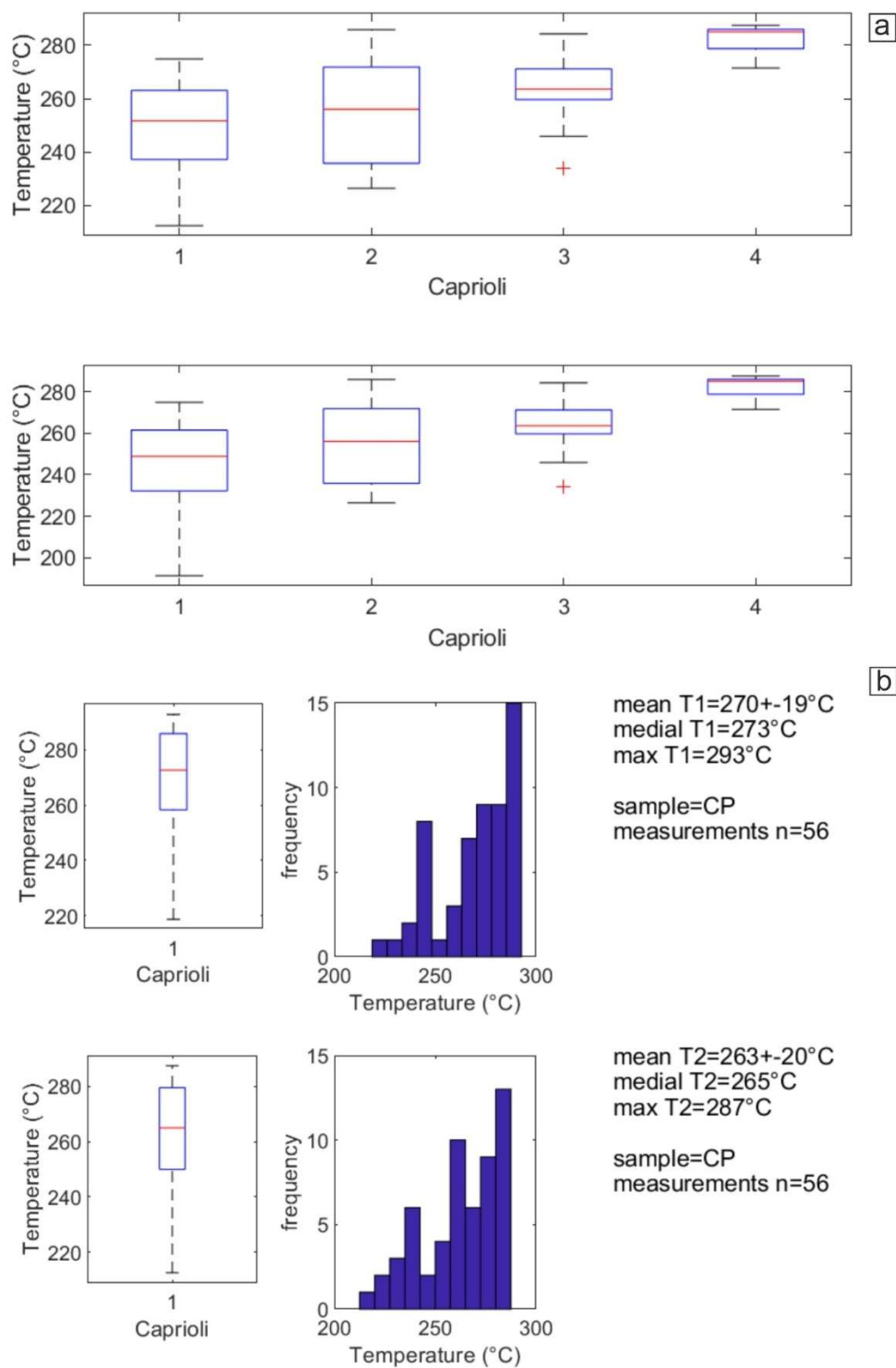


Figure 5-46: (a) Statistical comparison of T1 and T2 for four samples, showing overlapping error bars; (b) Frequency histograms of the combined data, with Gaussian distributions and mean values of 270 ± 19 °C (T1) and 263 ± 20 °C (T2).

Several samples were analysed and collected from three different locations (Pioppi, Punta Telegrafo, and Caprioli) from the Crete Nere Formation, as shown in Figure 5-47. The results indicate mean temperatures (T1) ranging from 252 °C (Pioppi) to 270 °C (Caprioli), and mean temperatures (T2) ranging from 244 °C (Pioppi) to 263 °C (Punta Telegrafo), respectively. The data are dispersed, though the average temperature values remain consistent across the samples. In each box plot, the central red line represents the median, while the bottom and top edges of the box represent the 25th and 75th percentiles. The whiskers extend to the most extreme data points that are not considered outliers, while individual outliers are marked with a '+' symbol.

Figure 5-46 and Figure 5-47 show a comparison of statistical data across the analysed samples, where the overlap of error bars suggests consistency in the measurements. To visualise the data distribution, frequency histograms for T1 and T2 were prepared, as seen in Figure 5-47 and Figure 5-48. The mean values for all data indicate temperatures of 270 °C (T1) and 263 °C (T2). Notably, both histograms exhibit a Gaussian distribution.

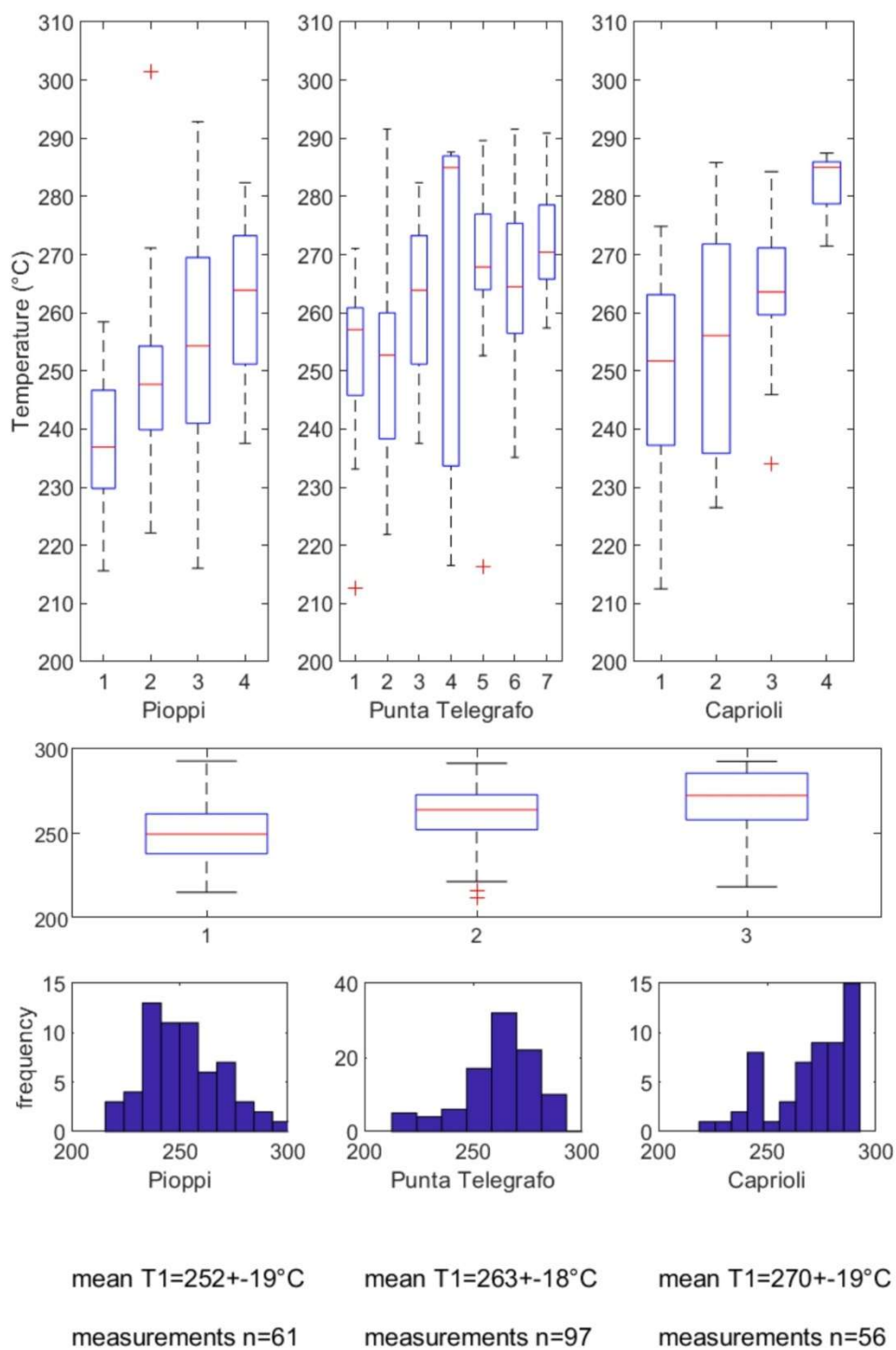


Figure 5-47: Temperature estimates (T1 and T2) from samples of the Crete Nere Formation at Pioppi, Punta Telegrafo, and Caprioli.

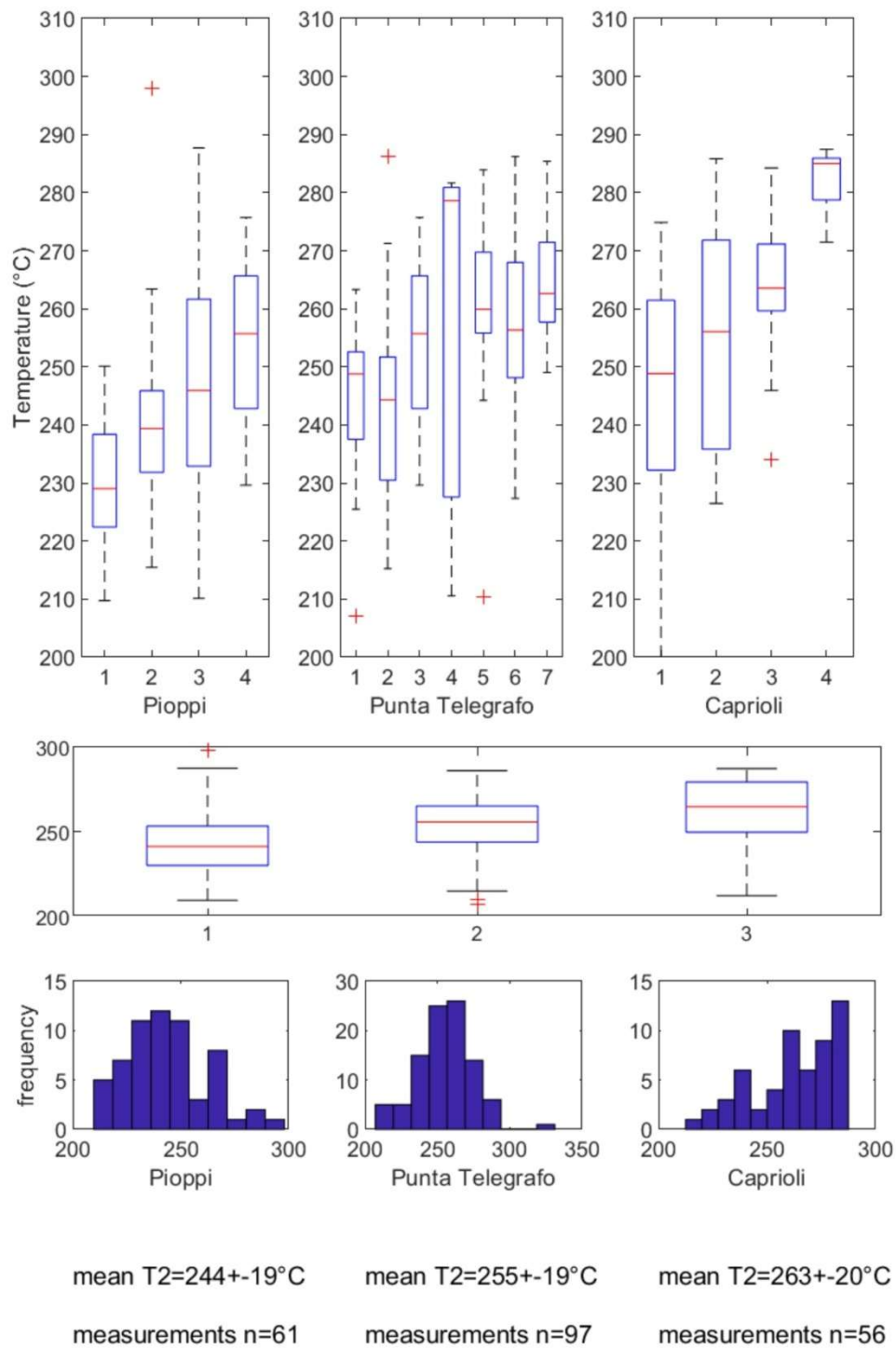


Figure 5-48: Mean temperature estimates (T2), with overall mean values of 270 °C (T1) and 263 °C (T2).

5.11 Comparison between the temperatures estimated by Raman and the Kubler index

To compare the temperature estimated by Raman (T1 and T2) of each sample and the temperature estimated by Kubler index (Tk), these values have been plotted in the diagram of Figure 5-49 a. It is possible to observe that the T1 and T2 are generally higher than those from the K index. The points are scattered, but a broad trend exists across the three localities, with increasing temperatures from Pioppi to Punta Telegrafo to Caprioli. This is well marked in Figure 5-49 b, where average values for the three localities are reported. The average Raman temperatures (Table 5.4) in Pioppi range from 244 to 252 °C, increase to 257-264 °C in Punta Telegrafo, and increase again to 263-270 °C in Caprioli (Figure 5-49).

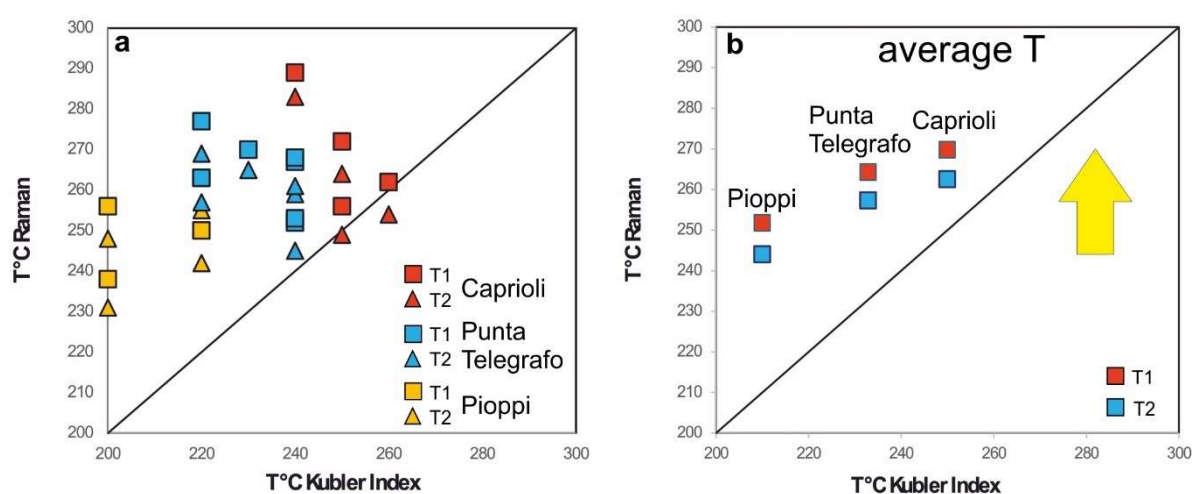


Figure 5-49: Temperature estimated by Raman vs Temperature estimated by K index: (a) mean values of each sample; (b) average values of the three study areas.

Table 5-4: Estimated temperatures by Raman and K-index.

Sample	T K-index	T1 Raman	T2 Raman	Average Tk	Average T1	Average T2
PP1	220	263	255			
PP2	200	238	231			
PP3	210					
PP4	200	256	248			
PP5	220	250	242	210	252	244
PT1	220	263	257			
PT1	220	277	269			
PT2	230	270	265			
PT3	240	267	259			
PT4	240	252	245			
PT5	240	268	261			
PT6	240	253	245	233	264	257
CP1	260	262	254			
CP2	250	272	264			
CP2	250	256	249			

CP3	240	289	283	250	270	263
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5.12 Thermobarometry

Based on the integrated analysis (Raman spectroscopy, White mica barometry and Chlorite thermometry), the thermobarometric characterisation of the Nord-Calabrese Unit in the three localities has been obtained.

For the sample PP, the white mica P-T plot (Figure 5-50 a) consists of a set of lines that are more tightly grouped compared to sample PT. The distribution is concentrated between ~280 °C and 420 °C. The chlorite P-T histogram (Figure 5-50 b) supports this result, displaying a sharp distribution restricted to ~280–300 °C.

The P-T plot of white mica of the sample PT (Figure 5-50 c) is depicted by a series of lines. The temperature axis extends from 200 to 600 °C and the pressure axis ranges from 2 to 14 GPa. The lines show a broad distribution, with values spanning between ~250 °C and 500 °C. The accompanying chlorite P-T diagram (Figure 5-50 d) shows a restricted temperature distribution, with values concentrated between 200 and 300 °C.

The white mica P-T plot for the sample CP (Figure 5-50 e) shows lines with an even narrower distribution than in the previous two samples. The lines span between ~260 °C and 380 °C, with minimal scatter. The related chlorite P-T histogram (Figure 5-50 f) reveals the narrowest temperature distribution of all three samples, confined to ~250–280 °C.

Joining all diagrams together (Figure 5-51), The P-T values obtained by intersecting the Wm lines with Chlorite histograms, furnish a common value of T=250-280 °C and P=0.3-0.4 GPa.

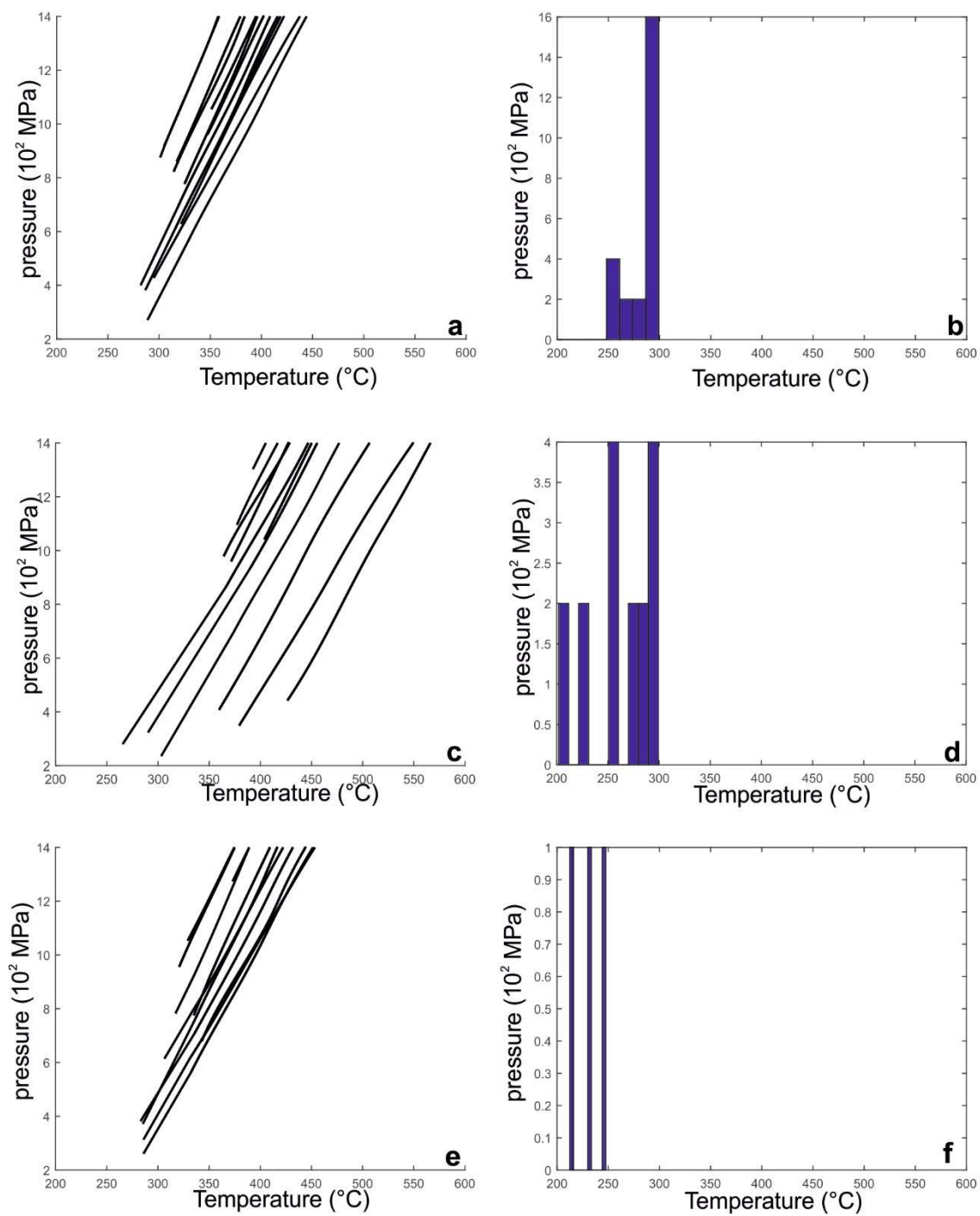


Figure 5-50: White mica P-T plot (a, c, e) and chlorite P-T histogram (b, d, f) for samples PP, PT and CP.

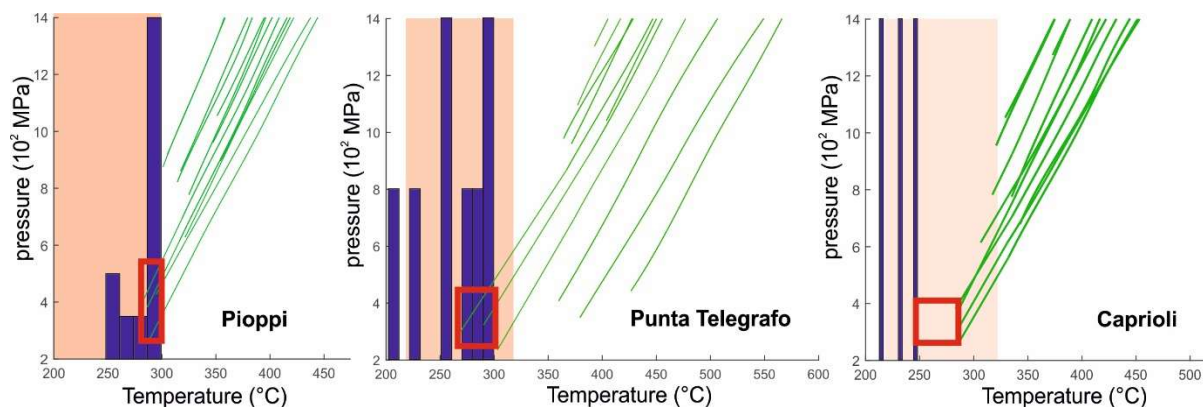


Figure 5-51: Joined P-T plots for the PP (Pioppi), PT (Punta Telegrafo) and CP (Caprioli) samples.

5.13 Discussion

5.13.1 Deformation structures and strain

This study focused on understanding the deformational evolution of the Ligurian successions exposed at three key locations: Pioppi, Punta Telegrafo, and Caprioli. Although previous works have proposed a similar study, some aspects were analyzed in greater depth. These included the characterization of the deformation mechanisms through which the Ligurian accretionary wedge formed.

First, this work confirms the complex deformation that these terrains underwent during the construction of the oceanic accretionary prism, which occurred in the early stage of the Apennine orogen. Mesoscale analysis, along with the orientations and kinematics of the main compressional structures, such as folds and thrusts, has highlighted how crustal thickening occurred with intense initial folding that generated, in the three locations, isoclinal to tight folds, often completely unrooted in ductile shear zones where both the folds and the more competent strata are completely dismembered and affected by boudinage. Locally, the D1 fold is preceded by a bed-parallel shortening with the formation of pre-buckle thrusts, which have sometimes been stretched and affected by boudinage.

At the mesoscale, this folding produces an axial plane cleavage, often parallel to the original layering. SEM investigations clearly show that this foliation in more pelitic layers (black shales) is characterised by alternating cleavage domains with white micas and microlithons dominated by quartz or carbonate minerals. It is classified as a spaced disjunctive cleavage.

At Punta Telegrafo, where a succession of calcareous turbidites is exposed, consisting of alternating limestone and pelitic layers (Punta Telegrafo member), the folding is accompanied

by intense vein development, especially during the stretching and thus boudinage phase, which frequently manifests as chocolate tablet boudinage, characterised by two orthogonal stretching directions. These veins indicate the presence of substantial amounts of fluid, which primarily accompanied the orogenic burial and the initial phase of folding.

The study of the main sets of veins in the Punta Telegrafo area revealed that the two main sets, V_{B1} and V_{B2} (Figure 5-56 a, e), orthogonal to the bedding, when restored, do not change much their orientation, suggesting that they formed before the isoclinal folding stage.

The second folding phase (D2) produced isoclinal to open folds (F2). In the Pioppi area, the F2 folds exhibit sub-horizontal axes and axial planes that dip to the southwest and northeast. In Punta Telegrafo, F2 folds exhibit horizontal axes and axial planes dipping to the northwest and secondarily to the southeast. In Caprioli, the few F2 folds show sub-horizontal axes and axial planes dipping to E/NE. The vergence of these folds varies between the three areas. In Pioppi, the F2 vergence is top-to-the NE, in Punta Telegrafo and Caprioli is top-to-the W/SW.

The two folds form interference patterns between types 2 and 3 of Ramsay's classification, meaning that the axes are generally parallel (type 3) and locally at a low angle (intermediate between types 2 and 3). Verticalized F1 folds are frequently observed due to this non-coaxiality between the two folds. One of the most spectacular places to see this superposition is at Punta Telegrafo, where ptigmatic F1 veins are folded non-coaxially by F2 folds, forming tortile structures. In Caprioli, a third fold set (F3) is present, but it is represented by crenulation.

Veins also developed during the second folding, but only secondarily. At the thin-section scale, veins formed during the previous folding phase are observed, which are shortened by subsequent shortening through stylolitic surfaces. This evidence is consistent with some boudinaged layers that were subsequently shortened, suggesting that some highly deformed zones (shear zones and stretched fold limbs) were successively affected by shortening (Figure 5-56 g).

No marked boudinage is observed associated with the second folding. Instead, numerous thrusts with associated folds are observed in the hanging wall and footwall. Generally, the vergence of these thrusts, which is derived from the presence of fibres and steps along the shear planes, is quite variable. However, associated with these thrusts are S-C, S-C' (Riedle shear) and S-C'' (anti-Riedle shear) structures, which generally indicate a vergence top-to-the NE in the Pioppi area and top-to-the SE in Caprioli. The S2 foliation, which sometimes represents the main foliation, appears under the electron microscope as a crenulation cleavage. Some rock

volumes, especially pelitic ones, are affected by varying degrees of transposition. Locally, the previous S1 foliation is completely transposed and can only be observed as intrafolial folds in the microlithons of the S2 crenulation cleavage.

The scheme in Figure 5-52 illustrates the general evolution of deformation in the studied successions. An early bed-parallel shortening predates the F1 folding stage, recorded by pre-buckle thrusts (Figure 5-52 a). Subsequently, the F1 deformation (Figure 5-52 b) folds the successions with variable intensity. Locally, the deformation focuses and produces severe stretching, as evidenced by intrafolial folds and boudinage (Figure 5-52 c). The successive F2 folding stage is again predated by a bed-parallel shortening (Figure 5-52 d) with the formation of some minor thrust faults that frequently short the previously boudinaged stiff layers. The F2 folding stage (Figure 5-52 e) refolds the previous structures. The late thrust faults crosscut the whole polydeformed succession associated with a crenulation F3 (Figure 5-52 f).

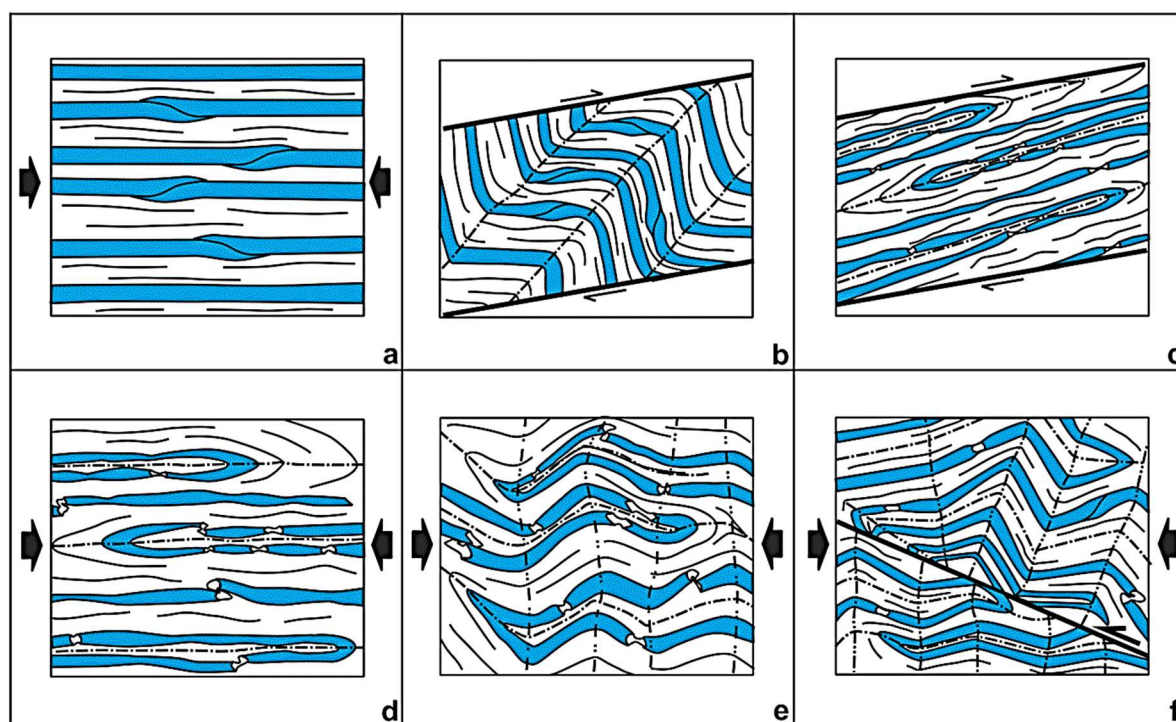


Figure 5-52: Schematic diagrams showing deformation stages of Nord-Calabrese Unit. a) Initial parallel shortening and pre-buckle thrusting; b) Fold amplification; c) Development of rootless folds and boudinage; d) Development of folded and deformed boudins. e) F2 folding; f) Meso- and macroscale final thrusting (Vitale et al., 2013).

A tardive spectacular shear structure is present in Pioppi. Here, a meter-thick ductile shear zone occurs. This structure likely formed during the last stages of the F2 folding event, and it could represent one of the major detachments within the oceanic wedge. The Pioppi Shear Zone is

characterised by highly deformed rocks that appear as a tectonic *mélange*. At the mesoscale, it is made by a main mylonitic foliation (S2) embedding calcareous and arenaceous porphyroclasts, crosscut by Riedle and Anti-Riedle shears and S-C structures, all indicating a NE vergence. The vorticity kinematic analysis indicates that the strain was characterised by a general shear, dominated by simple shear with a secondary component of pure shear (20-30%). This is in line with the kinematic vorticity estimated for the Tempa Roccarossa *mélange* (see chapter 3). Compared with W_k values obtained in various tectonic contexts worldwide (Figure 5-53), the estimated values are generally higher.

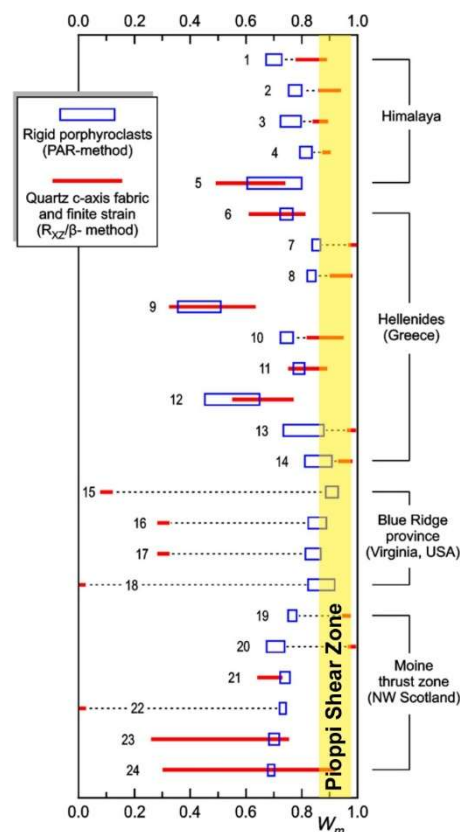


Figure 5-53: Comparative diagram showing the range of W_m values obtained by applying both rigid porphyroclast and RXZ/b methods of vorticity analysis (from Xypolias (2010)). The yellow area indicates the range of the W_k values for the Pioppi Shear Zone.

The strain analysis conducted on SEM images of the Pioppi Shear Zone rocks indicates that the strain was oblate. Compared with other FSEs from the southern Apennines (Figure 5-54 a), the PSZ ellipsoids indicate a lower strain intensity; however, they overlap with shear zones associated with thrust faults in the Pollino-Ciagola Unit (Vitale & Mazzoli, 2008) and Lagonegro Units (Mazzoli, Barkham, et al., 2001). When compared with FSEs from different tectonic settings in the world (Figure 5-54 b), they overlap with the lower part of the oblate field.

Finally, joining information about kinematics, strain and strain partitioning, the Pioppi Shear Zone can be classified as a widening-thinning shear zone (*sensu* Tikoff and Fossen, 1999) (Figure 5-55). Hence, this shear zone is defined by a dominant simple shear and a secondary pure shear, the latter deformation is also responsible for a stretching along the Y-axis that generates an oblate strain and a dominance of foliation (S-tectonite). Structures with a monocline symmetry (rotated porphyroclasts and Riedle-shears) that furnish information about the sense of shear (and the orogenic vergence) are observable only along the XZ plane,

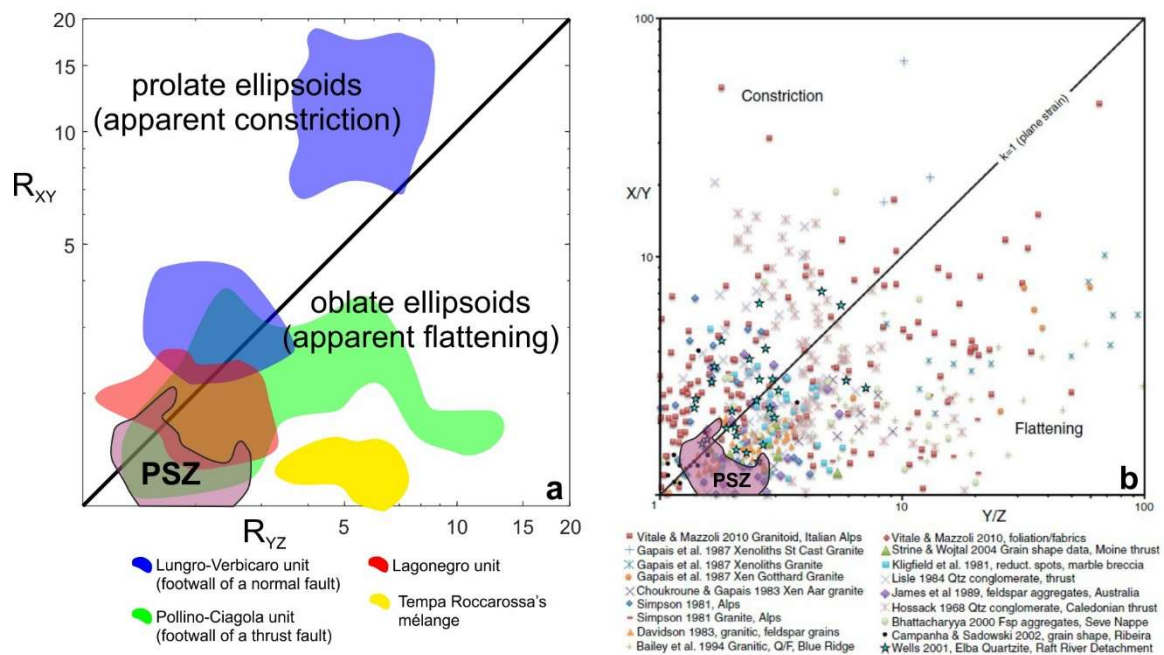


Figure 5-54: Ramsay diagram for the Finite Strain Ellipsoid (FSE). (a) Comparison between FSEs of the Pioppi Shear Zone (PSZ) and southern Apennine shear zones. (b) Comparison between FSEs of the Pioppi Shear Zone (PSZ) and world shear zones (from Fossen and Cavalcante (2017)).

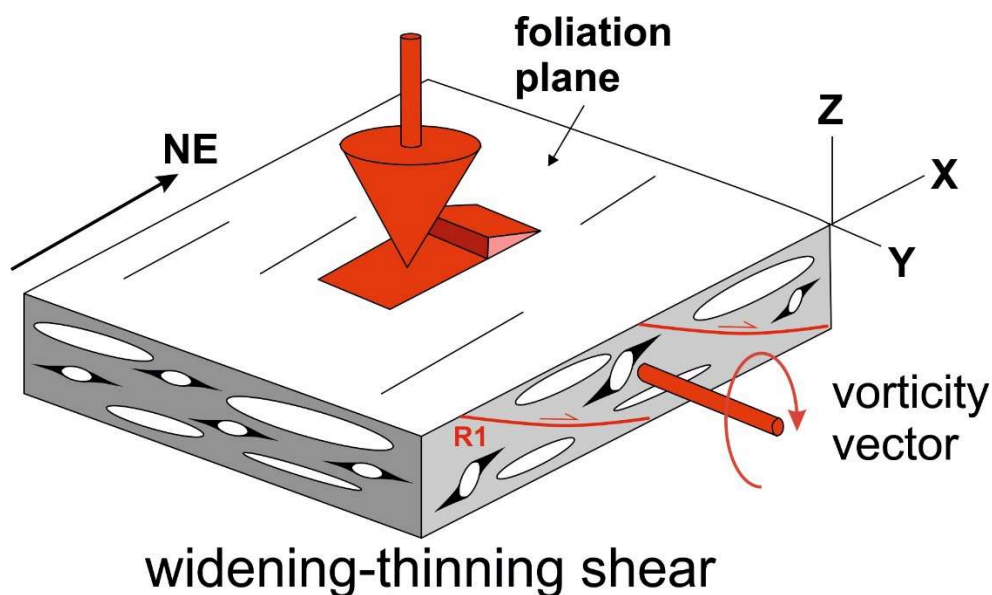


Figure 5-55: strain and kinematic model for the Pioppi Shear Zone.

In summary, considering all deformation stages and related fold development, a conceptual model of the veining process is presented in Figure 5-56.

During the orogenic burial of the oceanic succession, an intense compaction of sediments occurred (Figure 5-56 a) with the formation of some sets of veins orthogonal or with a high angle to the bedding and the formation of stylolites (ST_B) parallel to the bedding. Veins, frequently formed by elongate crystals of calcite, have an antitaxial growth (Figure 5-56 b). Fluids were fed by the oceanic, OCT or thinned crust sediments located below the main wedge decollement (Figure 5-56 a) and subsequently by the chemical reaction of clay minerals affected by pressure and thermal increase (Figure 5-56 c).

This process is defined by the release of water due to chemical reactions that convert (dioctahedral) smectite into illite via mixed illite/smectite layers, and eventually into muscovite mica (Elliott & Matisoff, 1996; Środoń, 1999; van de Kamp, 2008; Cavalcante et al., 2023). It also includes the transformation of (trioctahedral) smectite into chlorite through mixed layers of chlorite/smectite (e.g., Robinson et al., 2002; Beaufort et al., 2015). This reaction generates H_2O , silica, and other cations, which contribute to the final stages of quartz crystal growth in antitaxial veins (Figure 5-56 d).

Subsequently, with the involvement of these rocks in the crustal shortening, a new set of veins formed (V_{T1}) parallel to the bedding associated with a bed-parallel shortening deformation. These veins frequently show evidence of an inversion of stylolites (ST_B) and new stylolites

formed (Figure 5-56 e) orthogonal to the bedding. Veins V_B recorded this shortening by mechanical twinning in the large crystals of calcite (Figure 5-56 e). All described structures have been formed before the formations of F1-F2 folds. Hence, they formed during the burial of this succession and first stages of the shortening. Finally, with the formation of folds (Figure 5-56 f), veins V_B and V_{T1} were deformed, also as ptigmatic folds for the veins embedded in the pelitic matrix. However, during the folding process, other vein sets also formed: (i) veins V_{T2} parallel to the shortening direction, (ii) veins V_{T3} within the necks of boudins and (iii) veins V_{T4} at the outer arc of folded stiff layers. Veins V_B are further deformed within ductile shear zones and depending on their orientation with respect to the shear, they were stretched or folded.

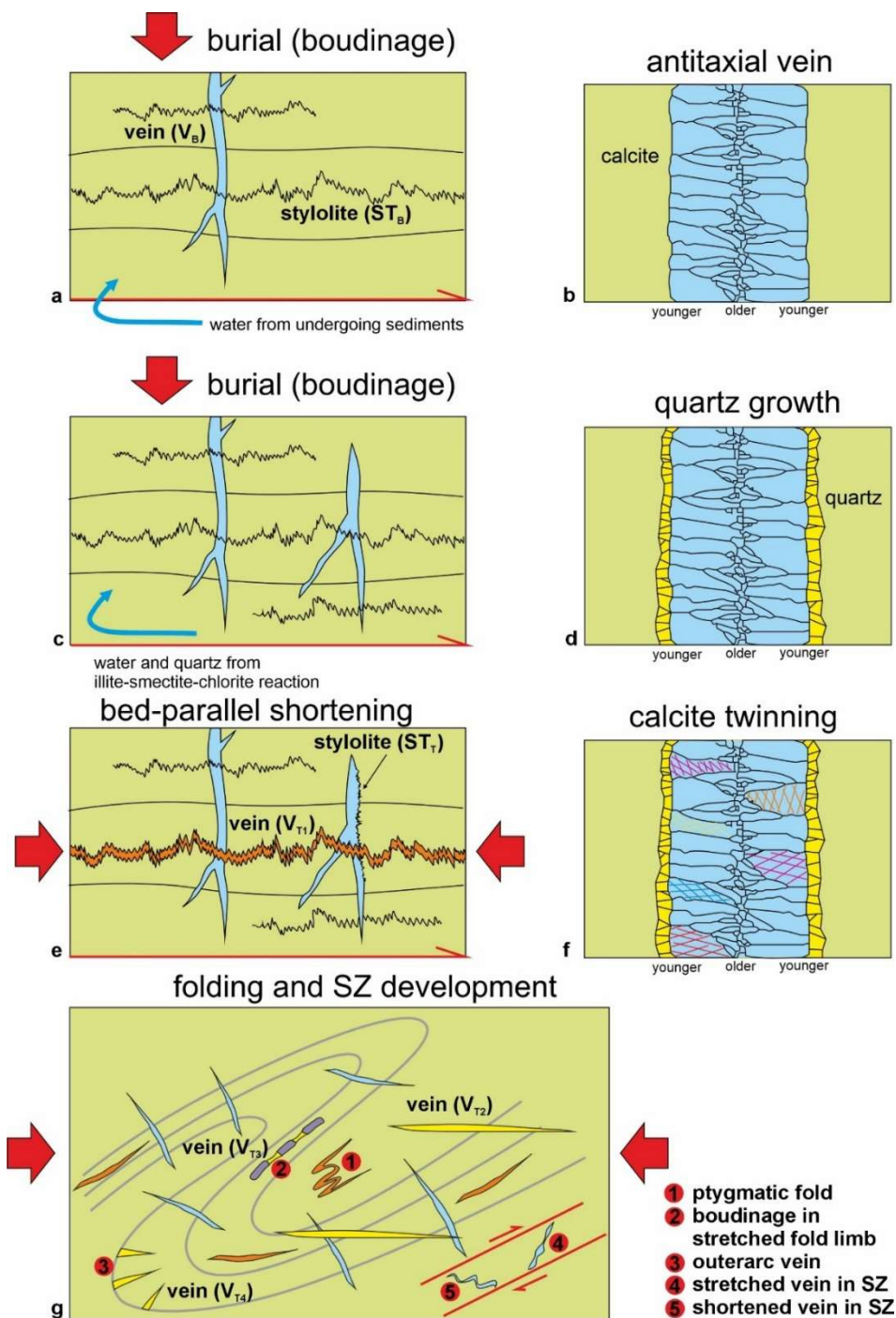


Figure 5-56: (a) The formation of veins (V_B) and stylolite's (ST_B) due to water influx from undergoing sediments; (b) Formation of an antitaxial vein with calcite growth, where younger mineral layers alternate with older ones; (c) Burial (boudinage) continuing with water and quartz generated from the illite-smectite-chlorite reaction; (d) Quartz growth in an antitaxial vein; (e) Bed-parallel shortening leading to the development of stylolites (ST_T) and veins (V_{T1}); (f) Calcite twinning within the vein, showing complex twinning patterns in alternating older and younger layers; (g) Folding and shear zone (SZ) development: Multiple vein generations (V_{T2} - V_{T4}) are shown with different structural features such as ptynmatic folds, stretched veins, and shortening in the shear zone.

5.13.2 Geothermo-barometry

The estimations of temperature by Raman (T1 and T2) on the organic matter and the Kubler index (Tk) by clay minerals furnish some food for thought. Firstly, T1/T2 and Tk show consistent ranges across the three localities. T1/T2 generally show higher values than Tk, typically 1.20-1.05 times higher. Furthermore, when average temperatures are compared between the three localities, a small thermal gradient exists from Pioppi to Punta Telegrafo to Caprioli. For temperatures estimated by the K index, Tk ranges from 210 to 233 to 250 °C. In the case of Tmean temperature, T1/T2 passes from 240/252 to 257/264 to 268/270. Hence, a thermal gradient occurs from NW (Pioppi) to SE (Caprioli).

These temperatures indicate that the Nord-Calabrese Unit underwent anchimetamorphism during the construction of the orogenic wedge. Only one previous study exists on illite crystallinity in this area. Mauro and Schiattarella (1988) provided a study on three areas: Torre di Caleo, Punta Telegrafo and Pisciotta (Figure 5-57). The first and last localities are very close to Pioppi and Caprioli, respectively, so the results of this study can be comparable to those of the current work. Figure 5-57, shows the diagram of the Kubler index for the samples from the three localities. Only the Pisciotta succession is characterised by anchimetamorphism, whereas Punta Telegrafo is not metamorphic, and Torre di Caleo is at the boundary between anchimetamorphism and high diagenesis. Hence, the study presented in this thesis represents an increase in knowledge of the thermal conditions of the Nord-Calabrese Unit in Cilento compared with previous work.

Furthermore, the anchimetamorphic nature of the rocks at the three sites is confirmed by P-T estimates based on white mica and chlorite mineralogy, yielding average temperatures of 250-280 °C and pressures of 0.3-0.4 GPa.

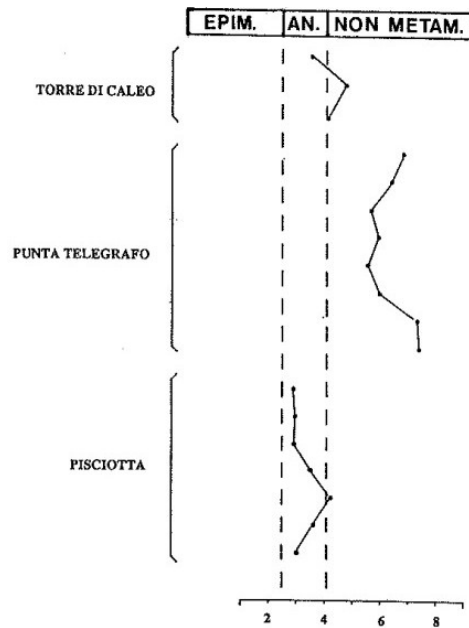


Figure 5-57: Kubler index from Illite Crystallinity for samples from Torre di Caleo, Punta Telegrafo and Pisciotta (from Mauro and Schiattarella, 1988).

These values can help to estimate the burial depth and the geothermal gradient. Considering a geobaric gradient of 0.027 GPa/km and pressures of 0.3-0.4 GPa, a depth range of 11-15 km is obtained. Hence, with temperatures ranging from 250 to 280 °C, a geothermal gradient of 17-25 °C/km is inferred. For the proposed model, average values are considered, hence $T_m=265$ °C, $P_m=0.35$ GPa, burial depth $BD_m=13$ km and a geothermal gradient of ca. 20 °C/km,

5.13.3 Insights into Ligurian accretionary wedge evolution

The occurrence of well-developed veining accompanying the first stages of prism accretion, and the presence of shear zones and décollements, are common in oceanic orogenic wedges (e.g., Maury et al., 1991; Moore & Vrolijk, 1992; Dal Zilio et al., 2020). Here, the prism materials lose fluid through both tectonically induced consolidation and thermally induced dehydration. Sediment that enters the subduction zone on the oceanic plate may either be quickly integrated into the accretionary prism (frontal accretion) or tectonically moved to greater depths along with the subducting oceanic crust (Figure 5-58). The accretionary prism forms directly from the material scraped off at the thrust front, which is the outermost sign of sediment deformation (Figure 5-58). The décollement zone, which acts as the plate boundary, distinguishes the unaltered underthrust sediment from the deformed accretionary prism. The accretionary prism becomes thicker as it extends landward, either through internal deformation of the scraped sediment or via thrusting, which refers to the movement of underthrust material to the bottom of the accretionary prism. Large quantities of fluid movement through restricted, highly permeable volumes are necessary for the formation of extensive carbonate and quartz veins found in accretionary prisms (Vrolijk et al., 1988; Byrne & Fisher, 1990; Orange & Breen, 1992). Moreover, significant dissolution occurring within subduction zones is linked to the leaching of silica from highly deformed rocks (Hammond, 1987).

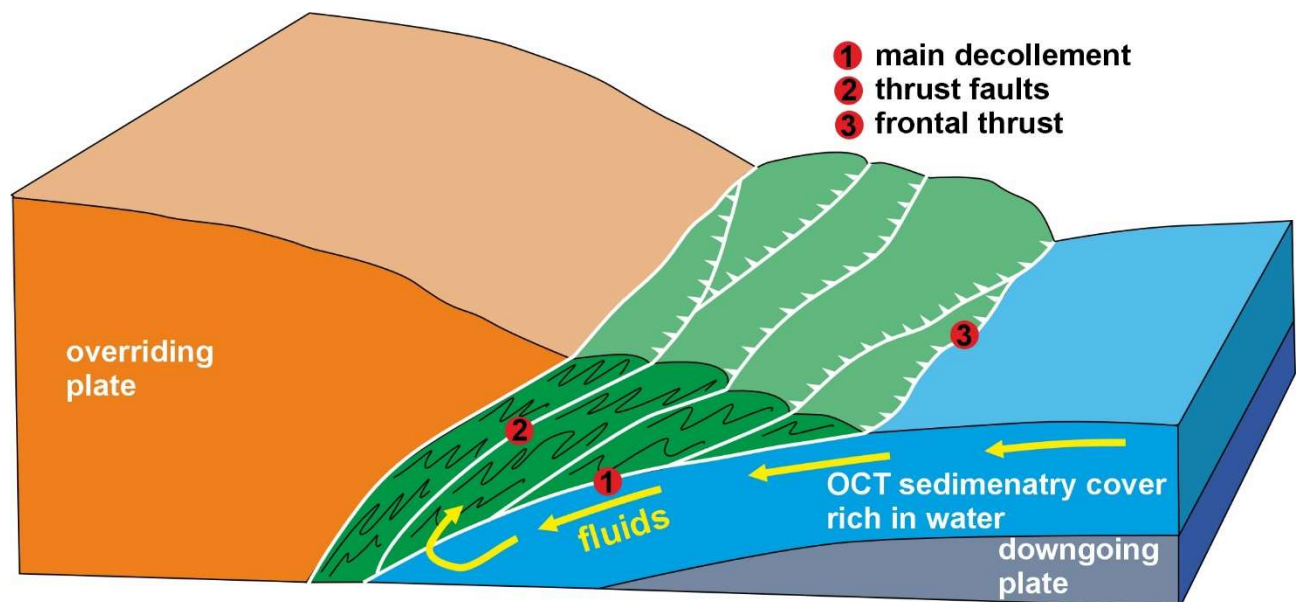


Figure 5-58: Schematic 3D model of an accretionary prism showing major tectonic elements and features controlling fluid emplacement and expulsion.

The significance of elevated fluid pressure and diminished effective stress in promoting low-angle thrust faulting has been acknowledged for a long time (Rubey & King Hubbert, 1959)

and has more recently been related to accretionary prisms affected by extensional décollements (Aoki et al., 1982; Westbrook, 1982). Accretionary prisms exhibit a structural style characterised by significant stratal disruption (Cowan, 1985); a considerable portion of this style can be attributed to the deformation of fluid-saturated, differently consolidated sediments within a high-strain-rate setting.

Fluids in accretionary wedges originate primarily from (i) pore-water expulsion during the compaction of sediments under increasing overburden, squeezing out interstitial water, and (ii) the dehydration of buried sediments such as clay minerals (e.g., smectite → illite → chlorite) that release water during progressive burial and heating (Emerman & Turcotte, 1983; Curtis, 1985; Westbrook, 1991; Moore & Vrolijk, 1992). Other secondary sources can be associated with (iii) hydrocarbon generation from organic-rich sediments that release fluids (methane, CO₂) during diagenesis and low-grade metamorphism (Ussler III & Paull, 2008); (iv) fluids released from the subduction channel, where hydrous minerals in altered oceanic crust (e.g., amphibole, lawsonite, serpentinite) break down at increasing pressure-temperature conditions (Prouteau et al., 1999; Zheng & Hermann, 2014); (v) Magmatic or mantle-derived inputs that generate fluids ascending from deeper magmatic or mantle sources, especially near forearc or backarc regions (Hamilton, 1985).

Generally, fluid pathways are controlled primarily by fractures, and fault zones act as preferential conduits for fluid flow. However, also permeability anisotropy (e.g., along bedding or thrust faults) governs flow direction (Zhu et al., 2021). When fluids migrate, they can reactivate preexisting faults due to their overpressure or can promote fault propagation. Moreover, fluids can induce rock brecciation and sediment fluidisation, triggering episodic fluid release events such as mud volcanoes and seeps (Dimitrov, 2002).

Finally, fluids can react with rocks, especially along fault-related damage zones, altering mineral assemblages (Wintsch et al., 1995) (e.g., carbonate precipitation, silica redistribution, clay transformation). This process is generally marked by (i) isotopic and elemental signatures (e.g., $\delta^{18}\text{O}$, Cl⁻, B, Sr, Li) in pore waters or vein minerals that trace fluid sources and flow histories (Wei, 2007; Wang & Xiao, 2018), and (ii) cementation and diagenesis affecting wedge strength and porosity (Saffer & Bekins, 2006).

Generally, oceanic accretionary wedges show a variable thickness. Figure 5-59, Figure 5-60, Figure 5-61, and Figure 5-62 show interpretative models of the Hikurangi margin (North Island, New Zealand) (Ghisetti et al., 2016), the Nankai Trough (Moore, 2014), the

Southwestern Alaskan margin (Li et al., 2018), the Northern Barbados margin (Moore et al., 1995), Southern Caribbean Deformation Belt (Mantilla-Pimiento et al., 2009; Rodríguez Álvarez et al., 2021) Makran margin (Kopp et al., 2000; Grando & McClay, 2007; Burg et al., 2012; Burg, 2018; Ruh et al., 2018). A more recent example from an active accretionary prism is the Makran; a detailed study by Pararas-Carayannis (2006) shows a maximum thickness of 15 km.

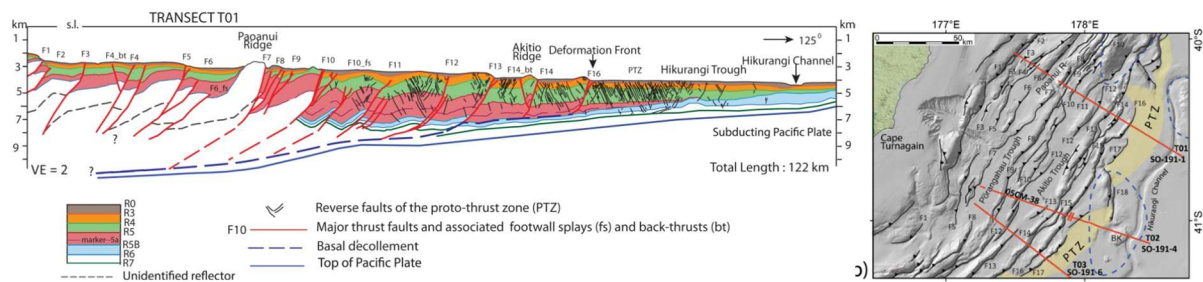


Figure 5-59: Geological interpretation of a seismic profile across the last 2 Myr of accretionary wedge construction in the central Hikurangi margin (North Island, New Zealand) (Ghisetti et al., 2016).

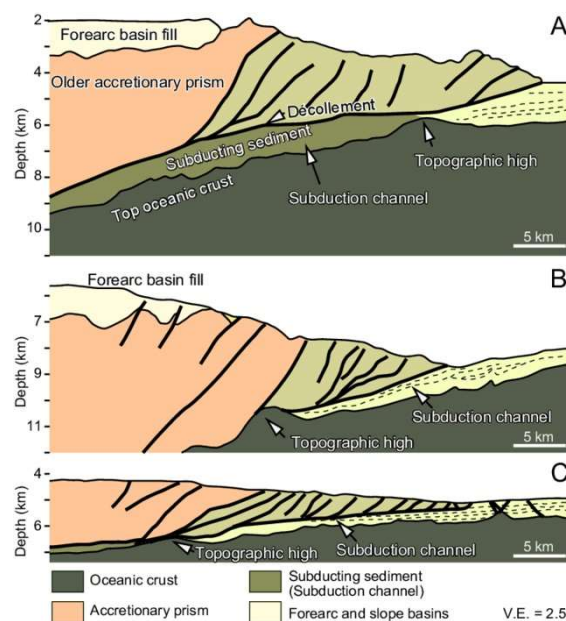


Figure 5-60: Representative cross sections of accretionary margins with topographic highs. (A) Nankai Trough (Moore, 2014). (B) Southwestern Alaskan margin (Li et al., 2018). (C) Northern Barbados margin (Moore et al., 1995). All drawings come from seismic profile interpretations. Vertical exaggeration (V.E.) is 2.5 for all cross sections. (From Noda et al. (2020).

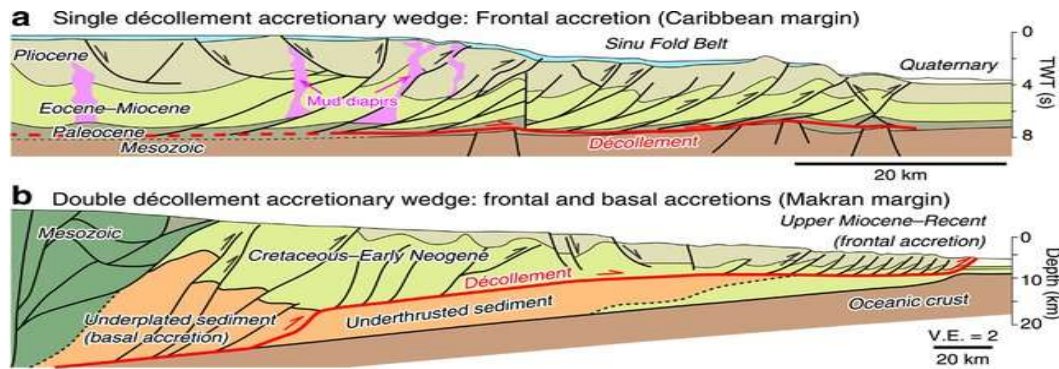


Figure 5-61: Natural examples of modern accretionary wedges with various décollement features; (a) Single décollement in the accretionary wedge of the Southern Caribbean Deformation Belt (Mantilla-Pimiento et al., 2009; Rodríguez et al., 2021); (b) Accretionary wedge dominated by both frontal and basal accretion (underplating) in association with landward stepping down of the décollement level at the Makran margin (Kopp et al., 2000; Grando & McClay, 2007; Burg et al., 2012; Burg, 2018), (From Noda et al. (2020).

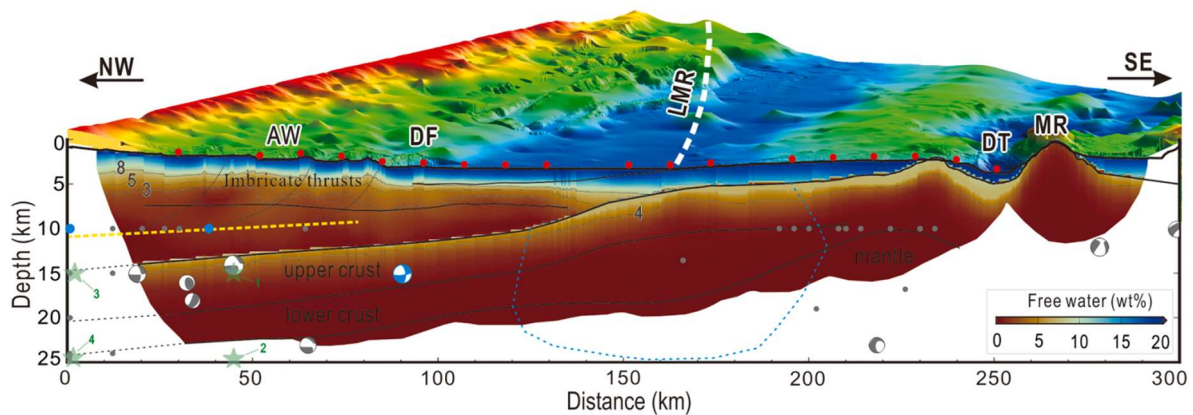


Figure 5-62: A schematic model of the Makran subduction zone illustrating the structure, seismicity, and water content of MSZ based on the velocity model. Black lines denote the interfaces of the oceanic subducting slab, including the sedimentary unconformity, basement, upper-lower crustal interface, and Moho discontinuity. The yellow dashed line and the sloping black dotted line above it represent the décollement and imbricate thrust faults, respectively, based on the results of (Kopp et al., 2000; Smith et al., 2012).. From Yu et al. (2024).

All these examples show that a depth of 13 km (such as inferred by the current study) would indicate a very thickened wedge, and that the exposure of these anchimetamorphic rocks requires severe tectonic exhumation.

Vitale et al. (2011) proposed a model for the evolution of the Ligurian Accretionary Complex (LAC) (Figure 5-63) where the LAC overthickened following the docking against the Apennine Platform domain and collapsed by means of low-angle normal faults that thinned the wedge and exhumed rocks located at depth. The collapsed wedge was subsequently filled by the Cilento Group sediments (see the next chapter).

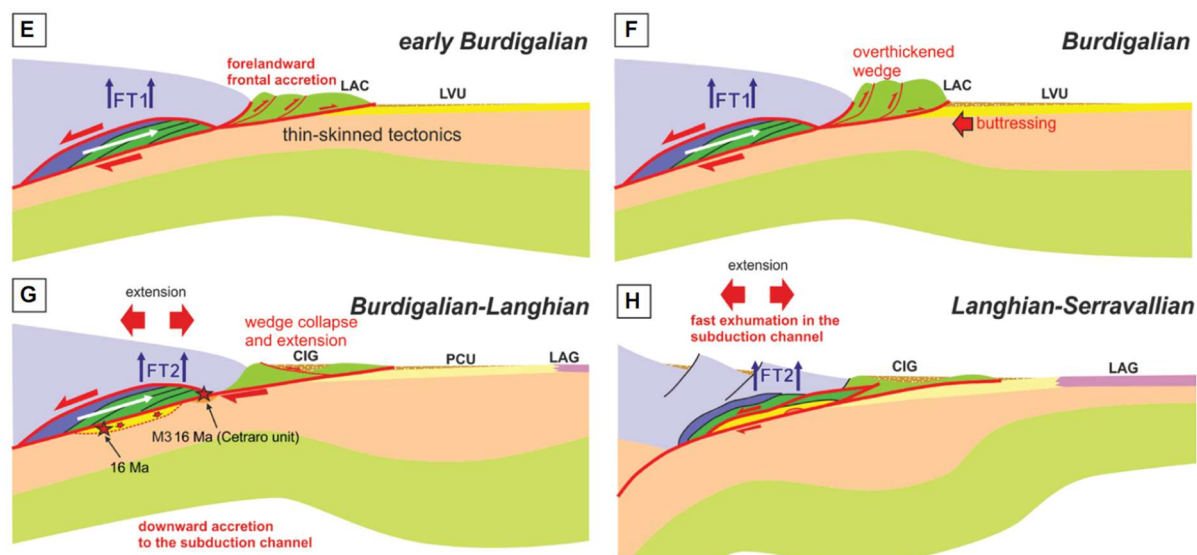


Figure 5-63: Cartoon showing the tectonic evolution of the Ligurian Accretionary Complex from the Burdigalian to Serravallian. Not to scale. AMA—Amantea; CIG—Cilento Group; CPT—Calabria-Peloritani terrane; CVTG—Castelvetero Group; FRI—Frido Unit; LAC—Ligurian accretionary complex; LAG—Lagonegro-Molise Units; LVU—Lungro-Verbicaro Unit; NCA—Nord-Calabrese Unit; PAS—Parasicilide Unit; PCU—Pollino-Ciagola Unit; SIC—Sicilide Unit. From Ciarcia and Vitale (2025).

This hypothesis is supported by the work of Corrado et al. (2019), who studied the thermal maturity of both organic and inorganic components of fine sediments from the Cilento Group. Their findings reveal two rapid phases of sediment accumulation during clastic deposition. The first phase dates to 16-13 Ma (uppermost Burdigalian–Langhian), and the notable accumulation rate of 2.7 mm/yr aligns with rapid subsidence, indicative of an extensional rather than a compressional environment.

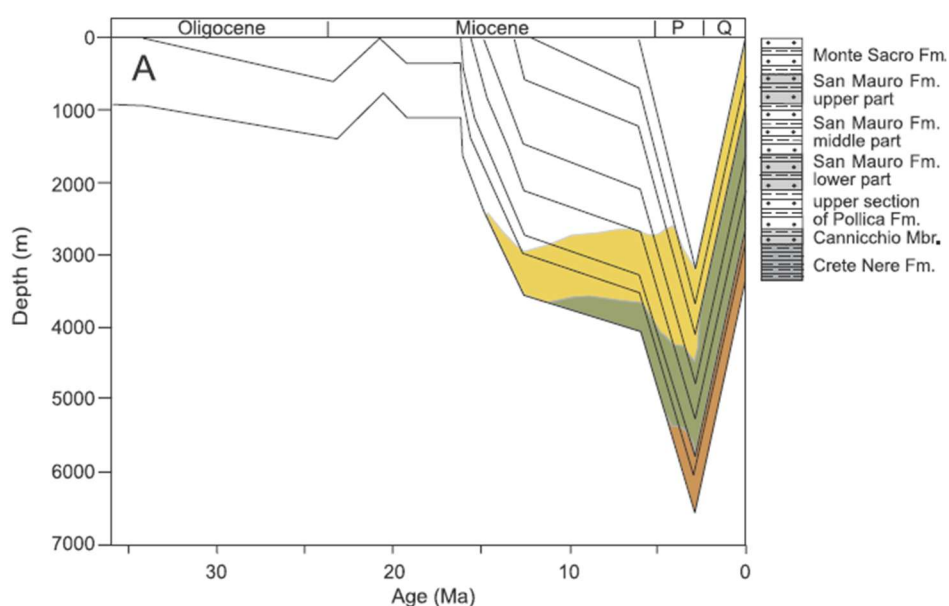


Figure 5-64: Representative one-dimensional burial and thermal model for the Cilento Group and its substratum. A: Burial history and hydrocarbon generation windows through time.

An alternative interpretation is to consider the anchimetamorphic rocks of Nord-Calabrese Unit in Cilento as being buried and partially overthrust by the overriding plate (Calabria-Peloritani Terrain, CPT) during frontal accretion. In this case, these rocks entered the upper part of the subduction channel; however, they never reached deeper levels where metamorphism can occur. However, this hypothesis requires that the overlying crystalline layer has been eroded and dismantled, as no evidence of tectonic superposition currently exists in Cilento.

Evidence of a severe erosion is witnessed by the Cilento Group (CiG) deposits made of sediments derived from the erosion of the LAC and the CPT (Figure 5-64; see also the next chapter).

5.14 Conclusions

The integrated structural, metamorphic, and petrographic analyses carried out in the three localities of Pioppi, Punta Telegrafo, and Caprioli collectively demonstrate that the lower portion of the Nord-Calabrese succession (Crete Nere Fm and Punta Telegrafo member) in Cilento underwent a coherent and polyphase deformation history linked to the Miocene closure of the Ligurian Ocean and the progressive frontal accretion of the Ligurian Accretionary Complex. Field structures, microstructures, and metamorphic indicators converge to define a burial-related deformational evolution characterised initially by orthogonal antitaxial veining and bedding-parallel stylolites, followed by early bed-parallel shortening that produced bedding-parallel veins, inversion of stylolites, and the onset of pre-buckle thrusting. Subsequent deformation was accommodated by at least two folding phases that generated S1 slaty cleavage and later S2 crenulation cleavage, with the Caprioli area preserving an additional S3 crenulation stage and pervasive crenulation lineation. The Pioppi Shear Zone further records a significant component of non-coaxial deformation, where Riedel shears, monocline-symmetry porphyroclasts, and vorticity estimates indicate a predominantly oblate finite strain ellipsoid with 20–30% pure shear and a consistent top-to-the-NE shear sense.

The metamorphic conditions reconstructed from Kubler index, Raman spectroscopy, and white mica–chlorite thermobarometry provide convergent evidence that the succession reached anchizonal temperatures of ca. 250–280 °C under pressures of 0.3–0.4 GPa, corresponding to burial depths of approximately 11–15 km and geothermal gradients of 17–25 °C/km. A clear thermal gradient from NW (Pioppi) to SE (Caprioli) parallels the structural variations, suggesting progressive deepening and higher strain intensity toward the SE sector. Petrographic

observations further support this interpretation, highlighting the consistent mineral assemblage of white mica, chlorite, quartz, calcite, and dolomite that records burial-related deformation and pervasive fluid-rock interaction. The widespread antitaxial veining, stylolitization, and late-stage quartz precipitation indicate the continuous presence and migration of fluids released from compaction and dehydration reactions (smectite $\rightarrow$ illite $\rightarrow$ chlorite/muscovite), which in turn facilitated deformation partitioning, thrust development, and veining episodes through all ductile stages.

Taken together, these structural, metamorphic, and tectonic constraints outline a coherent evolutionary model for the Ligurian Accretionary Complex in this sector. The rocks were incorporated into the frontal part of the orogenic prism during the Burdigalian, where they were progressively buried to ca. 13 km, shortened, and thickened through a combination of thrusting, folding, and shear zone development. Fluid-assisted deformation played a key role in enabling both early compaction-related veining and later thrust emplacement. The final stages of the prism evolution, following collision with the Apennine Platform and crustal overthickening, were characterised by gravitational collapse accommodated by low-angle normal faults, which exhumed the deepest portions of the complex now exposed at the surface. Thus, the integrated dataset supports a unified tectono-metamorphic evolution in which deformation, metamorphism, and fluid processes operated synchronously during prism growth and subsequent extensional unroofing.

Chapter 6 The Wedge-Top Basin of Cilento Group

6.1 Introduction

Wedge-top (or thrust-top) basins form above the active orogenic wedge, and together with foredeep, forebulge and back-bulge depozones, located ahead the thrust front, compose the Foreland Basin System (DeCelles, 2011). Wedge-top and foredeep basins are the places where thick piles of sediments accumulate, eroded from the allochthonous thrust sheets and the surrounding rocks, including the overriding and downgoing plates. The detrital modes of foreland basins are primarily influenced by tectonics, climate and eustasy (e.g., Dickinson, 1974; DeCelles, 2011). The standard stratigraphic succession consists of upward coarsening sequences unconformable onto a deformed basement, frequently sealing tectonic contacts between major thrust sheets (DeCelles, 2011). Wedge-top deposits have varying characteristics, with significant lateral facies variations, multiple local unconformities and complex paleogeographic features due to the neighbouring topographic reliefs that developed with the advancing front of the fold–thrust belt (DeCelles, 2011). Generally, wedge-top basins form in morphological depressions, frequently originating from tectonic structures, like synclines (e.g., Tinterri & Tagliaferri, 2015) or graben (e.g., Pascucci et al. (2006). Understanding the stratigraphy and the sedimentary features of the Foreland Basin System is crucial for constructing the tectonic evolution of the orogenic chains. By determining the ages of the deposits in the foredeep basin and the first wedge-top basin, we can estimate when the orogenic pulses occurred (Vitale & Ciarcia, 2022). Additionally, studying the detrital modes and their changes over time can provide valuable information about the source areas, aiding in paleogeographic reconstructions. Generally, wedge-top basin deposits are vulnerable to erosion because of their high structural elevation, and usually, they are preserved in post-orogenic structural depressions, such as the Cilento area in the southern Apennines (Figure 6-1) along the Tyrrhenian side of the orogenic chain. Here, the Cilento Group (CG) deposits (Amore et al., 1988), the subject of this study, are exposed.

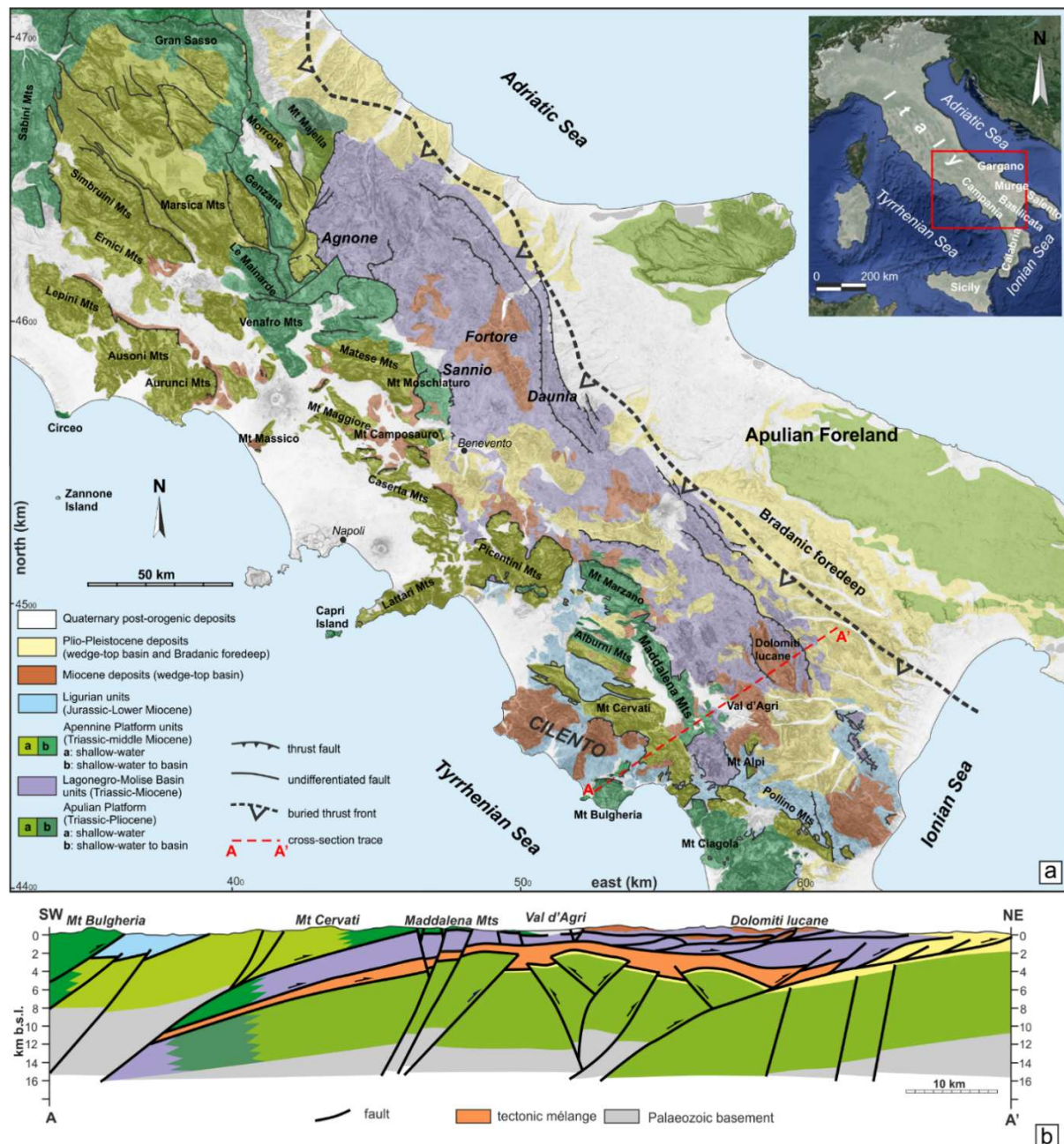


Figure 6-1: (a) Geological scheme of the southern Apennines (modified after Vitale and Ciarcia (2022)). (b) Geological cross-section (modified after Mazzoli et al. (2008)).

6.2 Cilento Group

The CG is one of the best-preserved wedge-top basin deposits in the southern Apennines, recording the first stages of orogenic construction. It consists of widespread outcrops of turbiditic successions, generally unconformably covering the oceanic Ligurian Accretionary Complex (LAC) formed by the Nord-Calabrese, Parasicilide and Sicilide thrust sheets, including Cretaceous–Lower Miocene successions affected by a complex polyphasic deformation (Vitale et al., 2011; Ciarcia et al., 2012). In the southern sector of the southern

Apennines (Basilicata region), the CG, at places, covers the Meso–Cenozoic successions of the Apennine Platform domain (ISPRA, 2023b), Foglio 503). The uppermost Burdigalian–lowermost Tortonian CG (Amore et al., 1988; Bonardi et al., 1988a; Russo et al., 1995), is characterized by siliciclastic, calciclastic and volcanoclastic deposits. In the Cilento area (ISPRA, 2023a, Foglio 519), the CG includes two turbiditic units: (1) the Pollica Fm, encompassing thin beds of siltstones and sandstones at the base, and prevailing sandstones in the middle-upper part; and (2) the San Mauro Fm, formed by sandstones and conglomerates with some characteristic levels of whitish marls (“Fogliarina” Auctt.). In the southern sector of the Cilento area, both formations laterally evolve into coarse clastic deposits, frequently hosting conglomerates (Torrente Bruca facies; (Guida et al., 1988). More southward, the CG is replaced by the undifferentiated Albidona Fm, consisting of siliciclastic turbidites with intercalations of calciclastic megaturbidites (Amore et al., 1988; Cesarano et al., 2002; Vitale et al., 2010).

6.3 Stratigraphy

The wedge-top basin deposits of CG comprise large exposure of turbiditic sediments with ages ranging from the uppermost Burdigalian to the lowermost Tortonian (Amore et al., 1988; Vitale et al., 2011; Mazzoli et al., 2013). It presents thicknesses ranging from 1200 to 2400 m and crops out extensively in the Cilento area (Figure 6-2). Here, this coarsening-upward megasequence unconformably covers the LAC units and, in turn, is unconformably covered by the Monte Sacro Fm (Castelvetere Group; Vitale et al., 2018) of the upper Tortonian–lower Messinian age. The sedimentary facies associations of the CG suggest a paleoenvironment characterized by submarine fan successions made of turbidite sequences (Hansen, 1962; Zuppetta & Mazzoli, 1997) formed by sediment gravity flows (Walker, 1978; Bouma, 1985). In the northwestern Cilento area, the CG comprises different turbidite depositional systems grouped into two sedimentary units: the Pollica and San Mauro Fms.

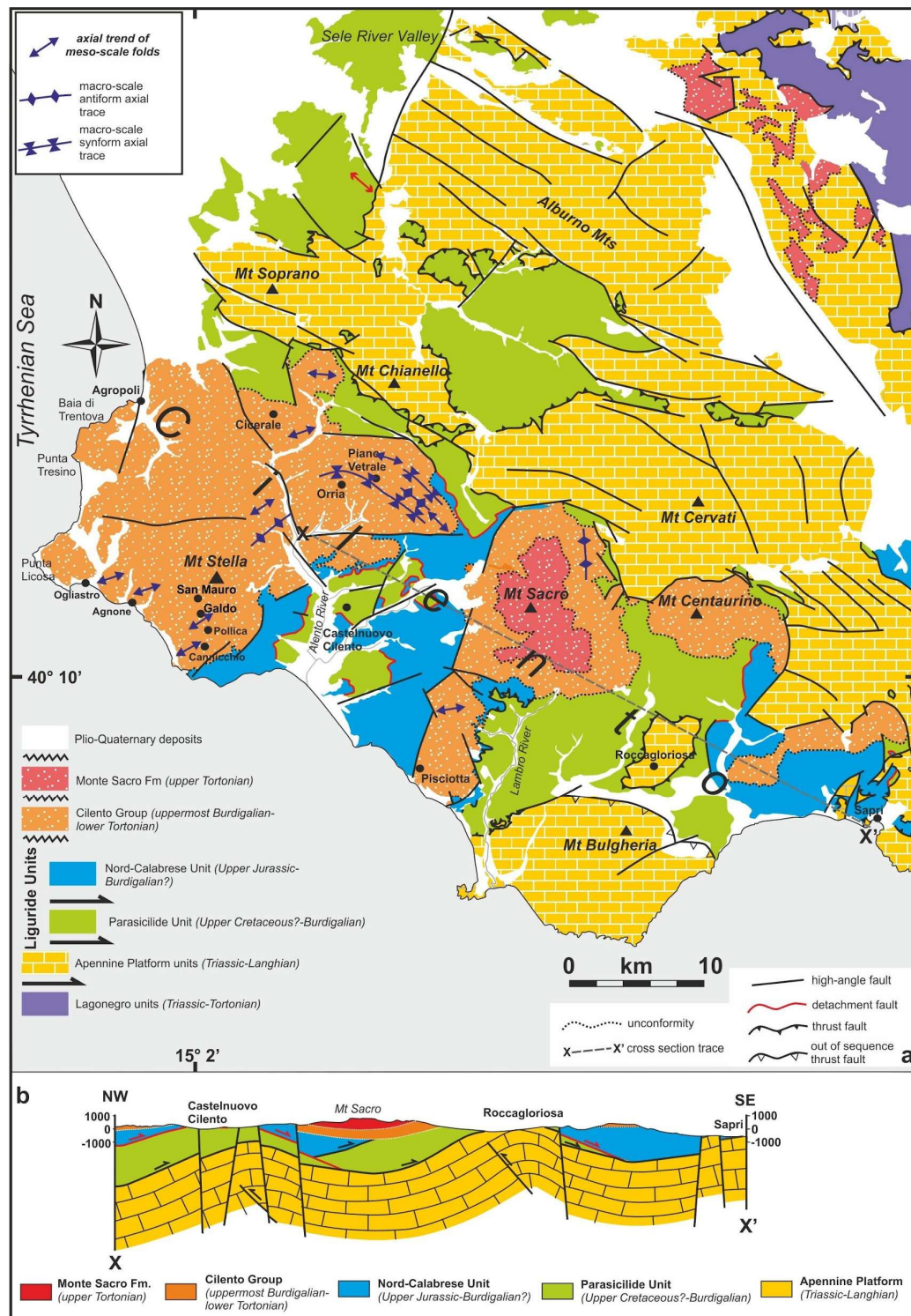


Figure 6-2: (a) Geological map of Cilento; (b) geological cross section (modified after (Vitale et al., 2011)).

The Pollica Fm (Figure 6-3), ~800 m thick (Amore et al., 1988), is characterized, at the base, by thinbedded shaly fine-grained arenaceous distal turbidites (Cannicchio Mb), followed by more proximal turbidites consisting of coarse-grained sandstones locally interbedded with conglomerates (Mutti et al., 2009). The Cannicchio Mb is characterized by regular and laterally continuous bedding, as well as the presence of thin hemipelagic intercalations. These features

are typical of the basin plain thin-bedded turbidites (Critelli & Le Pera, 1990) or lobe fringe (Mutti et al., 2009). The remaining part of Pollica Fm sediments has been deposited in a transitional and/or marginal area at the distributary channels and the suprafan lobes (Mutti, 1991) or at the middle-outer fan sequences (Normark, 1970; Mutti & Ricci Lucchi, 1972). The base of the strata is often erosional, and chaotic levels, which are associated with intraformational landslides, are frequent (Mutti et al., 2009).

The San Mauro Fm (Figure 6-3), ~1600 m thick (Amore et al., 1988), was formed, from bottom to top, by a thick sedimentary succession consisting of pelitic–arenaceous and arenaceous–pelitic layers with frequent carbonate intercalations, even in megabeds. In the higher sections, there are still intervals of sandy and muddy sediments, but there is an increase in conglomerate–sandstone and conglomerate fractions. The vertical evolution of the sedimentary facies highlights progressively more proximal features, showing a progradational sequence of the thickening and coarsening-up type (Mutti & Ricci, 1975), passing from outer fan deposits (fan fringe and depositional lobes) to inner fan deposits (channelized deposits and mouth and of channel embankment). The sedimentary succession also includes two carbonate–clastic megabeds (“Fogliarina” layers), respectively ~65 and ~35 m thick, and others of a lower thickness (from 10 m to 50 cm), exposed in the lower part of the sequence. A thick conglomerate–sandstone portion caps the succession (Ricci-Lucchi, 1975).

In the southeastern sector (from Pisciotta to Mt Sacro and Mt Centaurino, Figure 6-2 a), the sedimentary facies of both formations become coarser with the dominance of conglomerates (Torrente Bruca facies of Guida et al. (1988), consisting of a few tens of meters of thinly bedded sandstones and micaceous siltites followed by an arenaceous–conglomeratic succession with frequent thick beds of rudites made of crystalline clasts and frequently eroded clay balls and chips, enclosed in an arenaceous matrix. Some marly beds and megabeds are present.

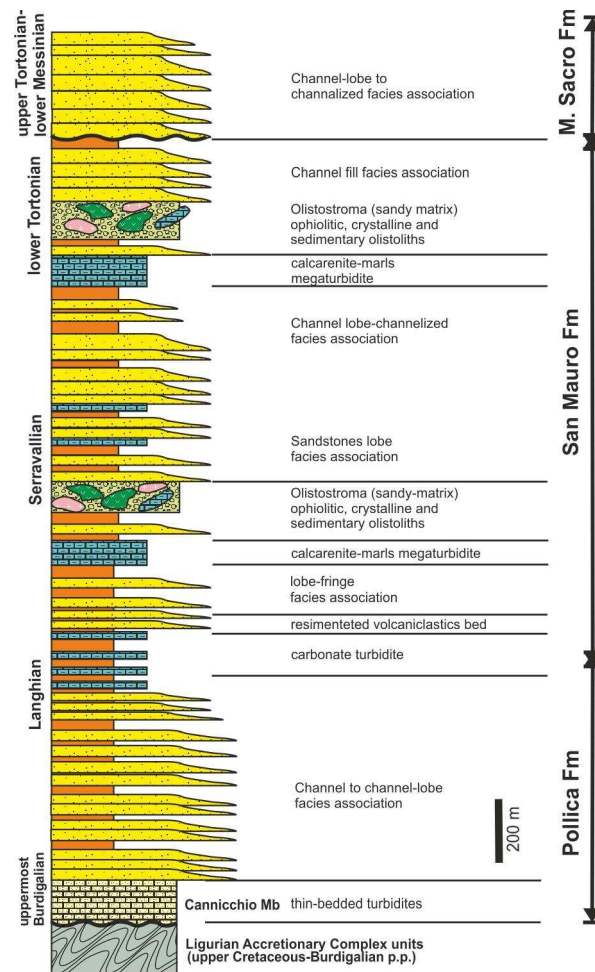


Figure 6-3: Schematic stratigraphic log of the Cilento Group succession (modified after Critelli (1999)).

6.4 Problematics and Goals

The aim of this work is to reconstruct, through the analysis of paleocurrents in the CG turbidites, the geometry of this wedge-top basin and the source areas that fed this basin, shedding light on the early geodynamic evolution of the southern Apennines. Paleocurrent analysis in the Cilento area has already been the objective of several studies in the past few decades (Pescatore, 1966; Cocco et al., 1978). However, these works analyzed only a small area of the Cilento, not differentiating the LAC from the wedge-top basin deposits. Hence, in the present study, we analyzed the CG deposits in a broader area of Cilento, reconstructing the original paleocurrent attitudes and analyzing the different formations.

For the assessment of turbidity paleocurrent directions, 74 different locations in Cilento were analyzed, spanning from the southern Pisciotta–Mt Sacro ridge to the northern town of Agropoli (Figure 6-2 a), resulting in the collection of a total of 338 paleoflow directions. An average of 4–5 paleocurrent indicators per site were measured, with data distinguished

according to their formations. As the succession was heavily deformed, especially in the lower part, for every paleocurrent orientation, the bedding attitude was recorded to allow the retro-deformation of the paleocurrent lineation, restoring the bedding in the horizontal attitude, assuming that the original paleocurrent plunge, which is lesser or equal to the turbidite bed dip angle, measures a few degrees. The retro-deformation of the paleocurrent and plotting of bedding orientation data was performed using the software TectonicsFP© 1.7.9 (Tozzi et al., 1996). Subsequently, the retro-deformed data were analyzed as rose diagrams for the final evaluation of the mean azimuth. Finally, paleocurrent lineations were plotted in a geological map of Cilento as arrows for each site, according to their original turbidity paleocurrent directions.

6.5 Geological Setting

The Cilento area is located in the southern Apennines (Figure 6-1), a segment of the Alpine circum-Mediterranean chains (e.g., Faccenna et al., 2001a; Carminati et al., 2012; Vitale & Ciarcia, 2013b; Van Hinsbergen et al., 2020). Currently, four first-order geological elements may be distinguished in the central Mediterranean area (Patacca et al., 1990): (1) the Tyrrhenian Sea, characterized by oceanic and thinned continental crust, representing the back-arc basin developed at the rear of the Apennine–Calabrian Arc system since the Serravallian time; (2) the southern Apennine chain, formed by the superposition of several thrust sheets, including the oceanic (LAC) and platform-to-pelagic basin successions of the Adria Plate of Lower Miocene to Middle Pleistocene in age; (3) the Bradanic foredeep basin, corresponding to the youngest (Pliocene–Pleistocene) flexural depression developed at the front of the thrust belt; and finally (4) the Adriatic–Apulia foreland, constituted of exposed Mesozoic–Tertiary carbonates. The fold-and-thrust belt of the southern Apennines (Figure 6-1 a) consists of a tectonic pile of thrust sheets (Figure 6-1 b) referring to different paleogeographic domains (Faccenna et al., 2001a; Mazzoli et al., 2008; Vitale et al., 2011; Carminati & Doglioni, 2012; Ciarcia et al., 2012; Vitale & Ciarcia, 2013b; Van Hinsbergen et al., 2020; Vitale & Ciarcia, 2022): (1) the Ligurian Ocean (2); Apennine Platform; (3) Lagonegro–Molise Basin; and (4) Apulian Platform.

The pre-orogenic stratigraphic architecture of the Adria Plate is made by deposits formed in shallow-water to deep-basin passive margin environments affected by different extensional events (Patacca & Scandone, 2007; Vitale & Ciarcia, 2013b; Vitale et al., 2018). These include

the Late Triassic–Lower Jurassic rifting stage that led to the opening of the Ligurian Ocean and the dismembering of the shallow-water carbonate domain, with the formation of the Apennine and Apulian platforms partially separated by the pelagic Lagonegro–Molise Basin. Subsequently, the Late Cretaceous–Eocene abortive rifting affected these realms, further dismantling the carbonate platforms. According to this pre-orogenic paleogeographic model (Ciarcia et al., 2009a; Vitale et al., 2011; Ciarcia et al., 2012; Vitale & Ciarcia, 2013b, 2022), the LAC includes remnants of the Ligurian Ocean (Nord-Calabrese, Parasicilide and Sicilide nappes), which separated the continental part of the Adria Plate (Apennine and Apulian Platform units and the intervening Lagonegro–Molise Basin) from the Europe/Calabria Peloritani Terrane (CPT) (Scrocca, 2010; Vitale & Ciarcia, 2013b). The orogenic crustal shortening started with the subduction of the Ligurian lithosphere under the Europe/CPT in the Paleocene/Eocene (Vitale & Ciarcia, 2022), with the complete closure of the Ligurian Ocean in the Early Miocene, and the onset of the LAC formed through frontal accretion of the easternmost oceanic successions. Subsequently, in the Middle-to-Late Miocene, the tectonic prism migrated to the E/NE with the inclusion of several thrust sheets composed of successions belonging to the Apennine Platform and the Lagonegro–Molise Basin. The orogenic stacking of the latter successions was mainly ruled by thin-skinned tectonics characterized by dominant flat lying thrust faults. Following the docking of the allochthonous wedge with the Apulian Platform in the Early Pliocene, the orogenic wedge superposed onto the Apulian carbonates. After that, the Pliocene–Pleistocene orogenic tectonic style was controlled by ramp-dominated thrust faults enucleated at depth within the Apulian succession, which presently forms a buried para-autochthon wedge (Mazzoli et al., 2008; Vitale & Ciarcia, 2022). During the orogenic evolution, several unconformable clastic deposits formed on top of the allochthonous units (wedge-top basin deposits), including the Lower-to-Middle Miocene CG (Figure 6-2), mainly covering the LAC units, the Upper Miocene Castelvete Group, which is widespread in the whole chain, the upper Messinian–lowermost Zanclean Altavilla Group and Pliocene Baronia and Sferracavallo Fms, which are exposed primarily on the external sector of the chain (Patacca & Scandone, 2007; Vitale & Ciarcia, 2018, 2022). In the Pliocene–Pleistocene time, the crustal shortening produced an out-of-sequence thrusting that involved the tectonic prism, including the wedge-top basin deposits (Vitale et al., 2017; Vitale et al., 2019; Vitale et al., 2020b). From the upper part of the Early Pleistocene, the orogenic chain was affected by the postorogenic extension, starting from the Tyrrhenian side and migrating to the East, with the formation of several intramontane basins and coastal plains, including the Campania Plain, where extensive back-arc volcanism occurred (Savelli, 2002).

6.6 Detrital Evolution

CG sandstones are quartzolithic, volcanolithic and quartzofeldspathic (Critelli & Le Pera, 1994). In addition, the upper part (San Mauro Fm) contains numerous calcarenites and marly megabeds and some olistostromes with thicknesses ranging from a few tens to hundreds of meters. Plutonic, metamorphic, and volcanic debris are found in the upper part of the Pollica Fm and the lower part of the San Mauro Fm. The latter deposits include abundant felsic calc-alkaline clasts, including rhyodacites and rhyolites (Critelli & Le Pera, 1994). The Miocene sandstones show a detrital evolution (Figure 6-4 a), indicating a recycled-orogen provenance related to the emplacement of CPT and LAC thrust sheets. In contrast, volcanoclastic sandstones in the median part of the CG (Figure 6-4 a) show a magmatic-arc provenance, such as the synchronous orogenic volcanism that occurred in the Sardinia block and Tyrrhenian basin, (Critelli & Le Pera, 1994). The debris deriving from the LAC appears only in the upper middle part of the CG (Figure 6-3 and Figure 6-4 a), consisting of sandstones and olistostromes of pillow lavas and gabbros (Mt Centaurino, Figure 6-2 a) and varicoloured argillites and radiolarites (Mt Sacro, Figure 6-2 a). However, the detrital-mode evolution of the LAC quartzolithic sandstones (Figure 6-4 a) is also related to the erosion of the CPT rocks with andesitic volcano-lithic supply (Cavazza, 1989; Critelli, 1991). On the contrary, carbonate-clastic mega layers of “Fogliarina” (Pescatore, 1966; Cocco & Pescatore, 1968) are well-distinguished by large volumes of calcareous sands and muds, likely deriving from the orogenic deformation of Apennine Platform carbonates located to the east (Colella & Zuffa, 1988; Critelli & Le Pera, 1994; Cieszkowski et al., 1995). Recently Corrado et al. (2019) analyzed the fine-grained portion of the CG to determine paleo-weathering, provenance, and recycling information. The authors state that the source area primarily comprises felsic rocks with a smaller number of mafic rocks, mostly corresponding to an upper crust composition evidenced by the Ni-La*4-V ternary plot (Figure 6-4 b). Furthermore, the Chemical Index of Alteration (CIA) estimations suggest that moderate chemical paleoweathering processes accompany erosion. Finally, the analyzed sediments indicate poor sorting and rapid deposition.

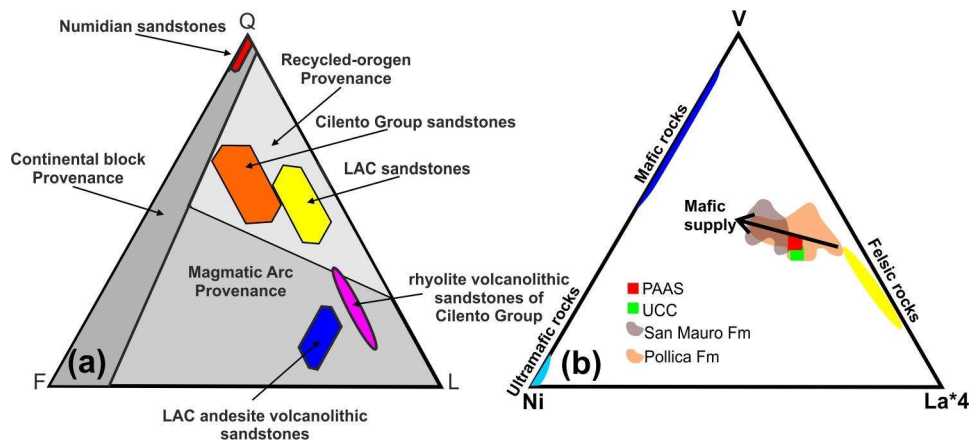


Figure 6-4: (a) QFL triangular plot, showing detrital-mode evolution of the CG, LAC and Numidian sandstones of the southern Apennines. Data sources (Critelli, 1991; Critelli, 1993; Critelli & Le Pera, 1994). (b) Provenance plot based on the V-Ni-La*4 relationships, showing fields representative of felsic, mafic, and ultramafic rocks plot separately. Modified after Corrado et al. (2019).

6.7 Field Observations

6.7.1 Deformation

The CG succession recorded the out-of-sequence deformation stage that affected the orogenic chain in the lower Pliocene (Vitale et al., 2020a; Vitale et al., 2020b). The deep structure of the Cilento area is defined by some out-of-sequence ramp-dominated blind and emerging thrust faults that formed the tectonic windows of Castel Nuovo Cilento and Roccagloriosa (Figure 6-2 b; (Vitale et al., 2011) and the overthrusting of Mt Bulgheria onto the LAC (Tozzi et al., 1996) (Figure 6-2 a). The CG deposits host several structures, including thrust faults and folds indicating a roughly NW–SE shortening and a vergence toward SE, however in the Orria-Piano Vetrare area (Figure 6-2 a), an SW verging overturned synclinal occurs (Zuppetta & Mazzoli, 1997).

Generally, the mesoscale folds are tight to open with a kink shape, such as in the Cannicchio Mb (Figure 6-5 a) and the middle-upper part of the Pollica Fm (Figure 6-5 c). Brittle shortening structures appear as pre-buckle thrust (Figure 6-5 b) and thrust faults (Figure 6-5 d, e, f) with centimetric to metric displacements and associated drag folds.

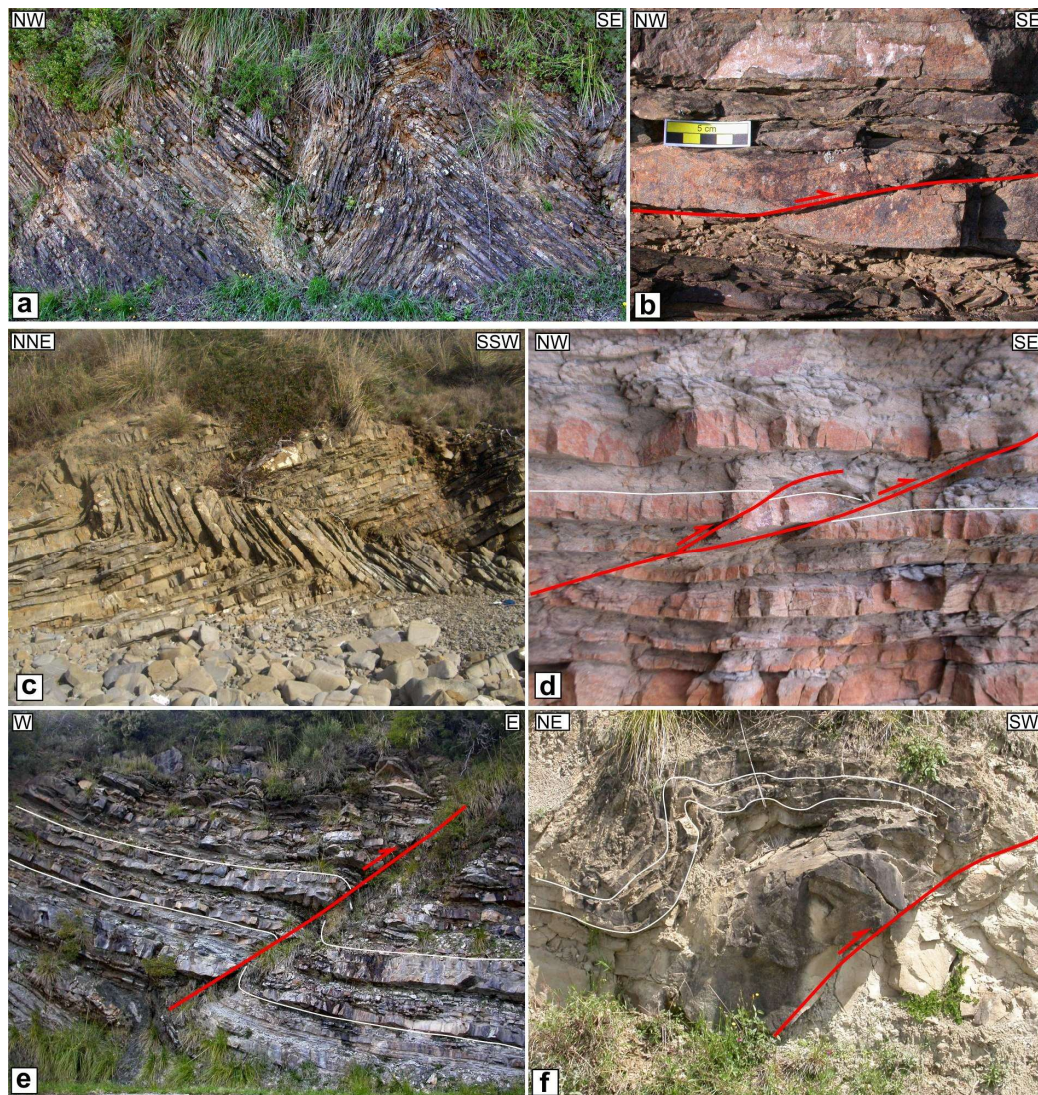


Figure 6-5: Examples of deformation of the CG deposits. Cannicchio Mb: (a) kink folds (Cannicchio); (b) pre-buckle thrust (Cannicchio). Pollica Fm: (c) overturned kink folds (Agnone locality); (d) thrust verging to SE (Punta Licosa). San Mauro Fm: (e) thrust verging to E; (f) thrust and related folds verging to SW (Piano Vetrale).

6.7.2 Paleocurrent Indicators

Various paleocurrent indicators (Allen, 1982; Reiter, 2023) were identified in the field (Figure 6-6 and Figure 6-7). These structures mainly consist of flute casts (Figure 6-6 and Figure 6-7 Figure 6-6 e, f and Figure 6-7 a) and, to a lesser extent, groove casts (Figure 6-7 b), flame structures (Figure 6-7 c) and ripple marks (Figure 6-7 d). Most of these structures were preserved within the more resistant lithologies, such as sandstones, especially those with larger thicknesses. Flute casts are structures formed by a paleocurrent flow that erodes and widens downstream from a specific point known as the “nose”.

These structures suddenly deepen before gradually becoming shallower toward their end (Allen, 1982). Flute casts form when a substrate erodes and is then typically filled with sand. These marks are recognizable as indentations on the bottom of beds. The quality of the paleocurrent indicators depends not only on the movement of the sedimentary material but also on the depositional dynamics; in fact, during the deposition of the turbidity currents, the more energetic ones dig furrows at the bottom. However, in several cases, such as those shown in Figure 6-6 e, f and Figure 6-7 e, f, the presence of ichnofossils (*Thalassinoides* and *Helminthorhapse*), Peakall et al. (2020) has perturbed the layers, partially or fully eliminating the footprints of paleocurrents; as a result, several outcrops were not considered reliable and therefore prudently discarded or cautiously analyzed. In some cases, flame structures (Figure 6-7) and ripple marks (Figure 6-7 d) were used to evaluate the paleocurrent direction as the line orthogonal to the lineation that these structures form when crosscutting the bed surface. The best-preserved turbidity paleocurrents were observed in the northern sector of the Cilento due to the occurrence of fine-grained sediments compared to the southern part, where Pollica and San Mauro Fms show a dominance of conglomerates (Torrente Bruca facies).

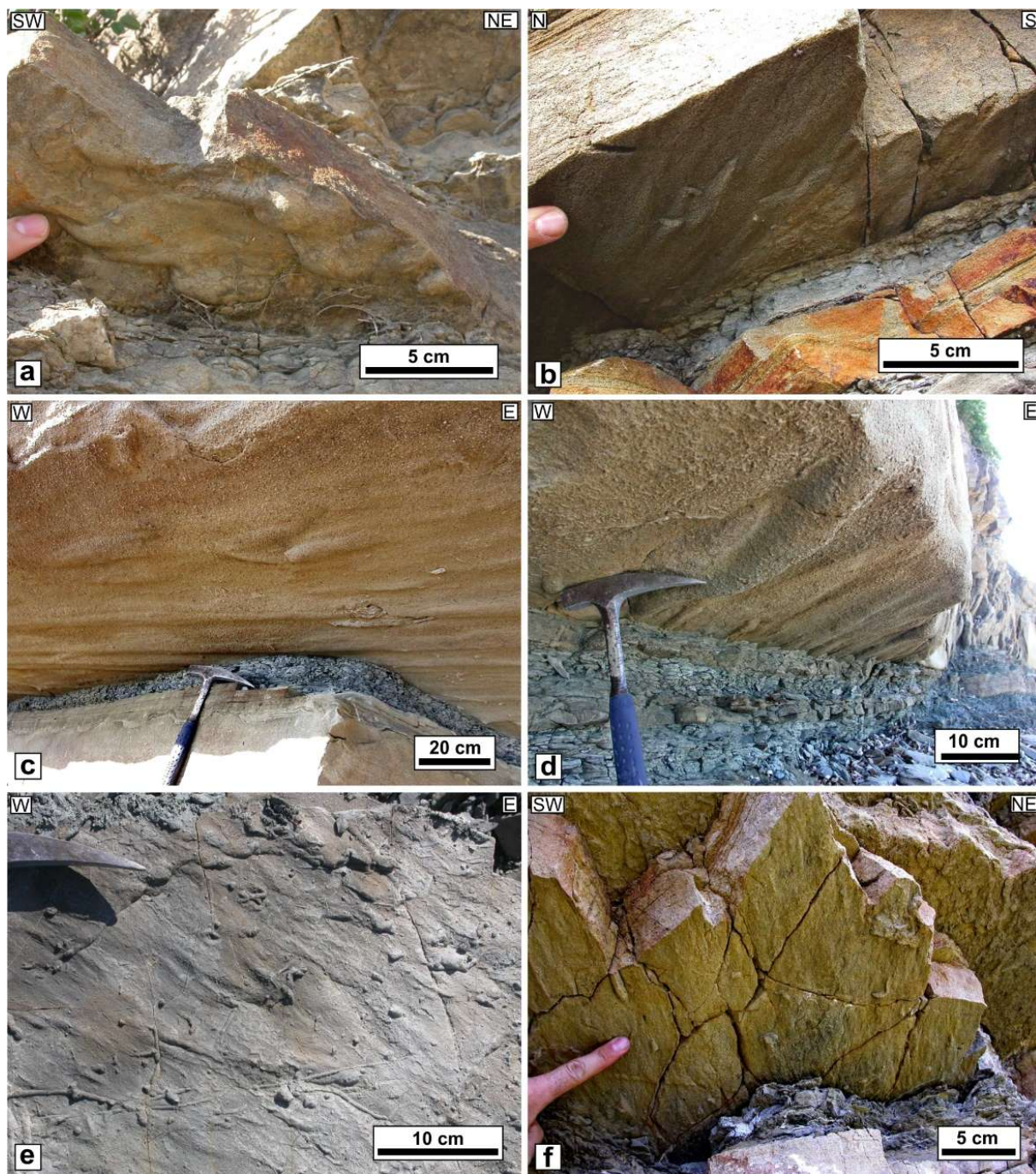


Figure 6-6: Examples of paleocurrent indicators. Pollica Fm (Punta Licosa): (a) bulbous and (b) spindle flute casts. Flute casts in San Mauro Fm: (c) Punta Tresino and (d) Baia di Trentova. (e-f) Flute casts and ichnofossils (*Thalassinoides*) in San Mauro Fm (Baia di Trentova).

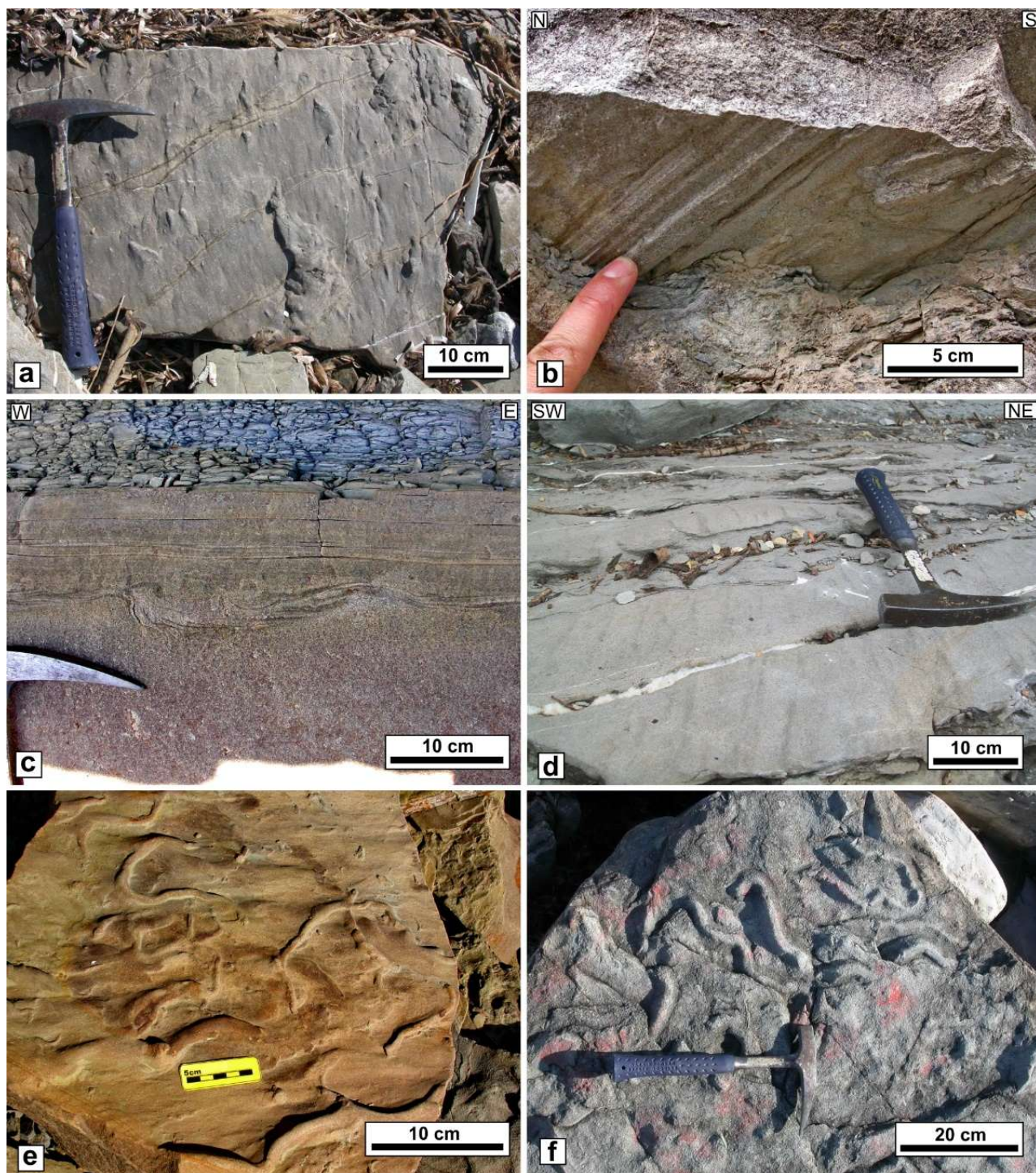


Figure 6-7: Examples of paleocurrent indicators. (a) Flute casts in the San Mauro Fm (Baia di Trentova). (b) Groove cast indicators in the San Mauro Fm (Galdo). (c) Flame structure in the San Mauro Formation (Punta Tresino). (d) Ripple marks in the San Mauro Fm (Agropoli). (e) Ichnofossils (*Helminthorhaphe*) superposed on flute clasts in the San Mauro Fm (Punta Tresino). (f) Ichnofossils (*Helminthorhaphe*) in the San Mauro Fm (Agropoli).

6.7.3 Restored paleocurrent lineations

For each site, the mean trends of the restored paleocurrent directions have been calculated and plotted as arrows in the geological map (Figure 6-8 a). It is worth nothing that the paleocurrent directions form a generally consistent pattern where parallel lineations are dominant in some areas, and orthogonal sets occur in others. Furthermore, in several places, opposite trends are present.

We grouped all restored paleocurrent lineations according to the hosting rock nomenclature (Formations). The rose diagram of Pollica Fm (Figure 6-8, b) indicates the main preferred trends toward the SSW and NE and, secondarily, the SW, ESE and NW.

The rose diagram of San Mauro Fm (Figure 6-8, c) shows the main directions toward the E, S, SW and NE. To restore the flow directions with the correct orientation in the uppermost Burdigalian-lower Tortonian period (age of the CG), we must consider that the Cilento area rotated counterclockwise at an average angle of 60° from the Miocene, according to the studies on magnetic lineations in the carbonates of Alburno and Bulgheria Mts (Uchman, 1995) located nearby the study area (Figure 6-2 a). Therefore, we rotated all the data clockwise around a vertical axis of 60°. The resulting rose diagrams (Figure 6-8 d, e) indicate that the restored turbiditic flows were directed mainly toward the WSW, NW, SE, SSE and NNE for the Pollica Fm and the SSE, ESE, WSW and WNW for the San Mauro Fm.

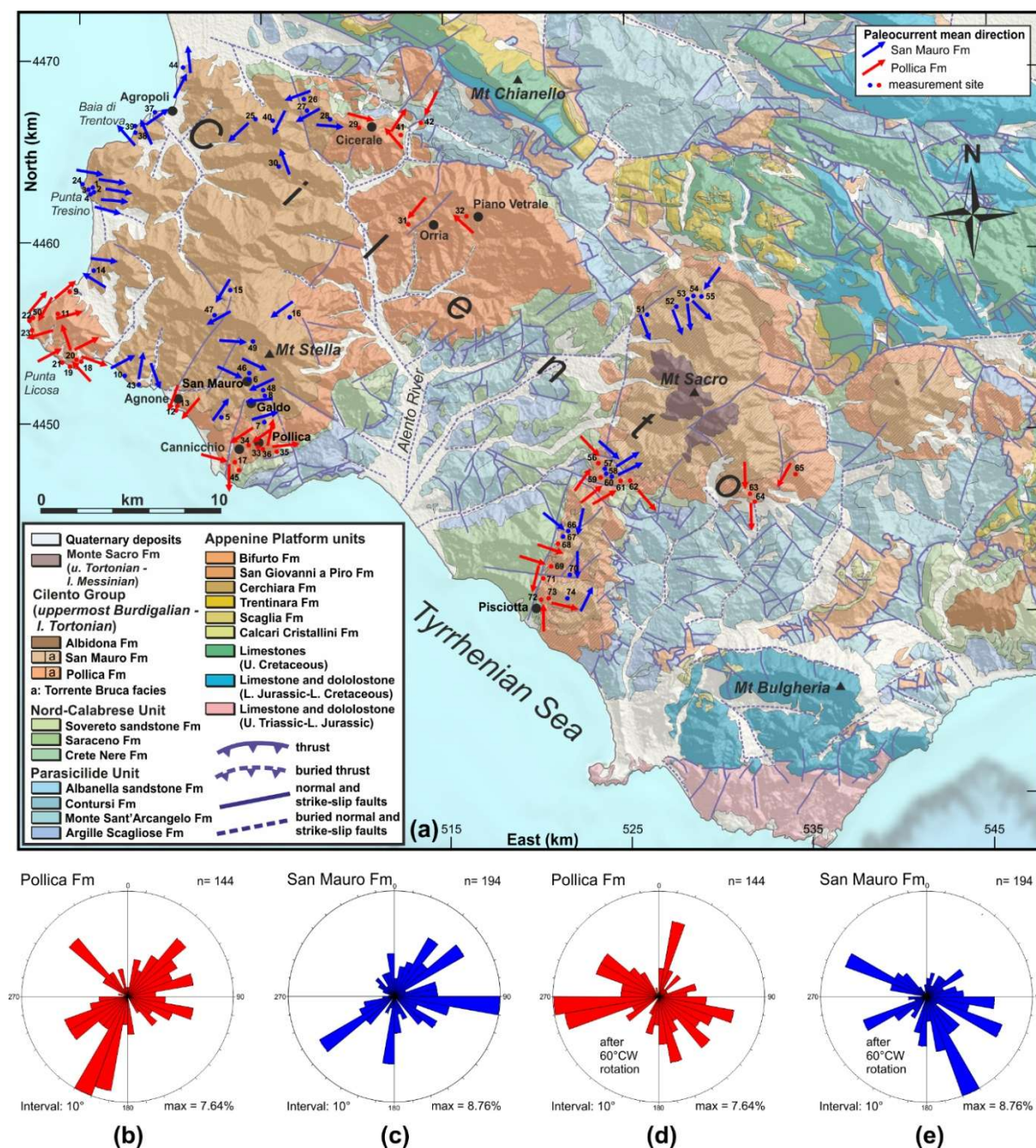


Figure 6-8: (a) Geological map of Cilento with oriented indicators of the mean paleocurrents. Rose diagrams of the directions of the restored paleocurrents: (b) Pollica Fm; (c) San Mauro Fm. After 60° clockwise (CW) rotation: (d) Pollica Fm; (e) San Mauro Fm.

6.8 Discussion

The paleocurrent analysis performed on the CG deposits provides some valuable information that is helpful in reconstructing the geometry of the wedge-top basin formed in the early stages of the southern Apennine orogeny. As previously stated, the paleocurrent pattern exhibits large heterogeneity. In neighbouring areas, the turbidity paleocurrents are typically nearly parallel with consistent flow directions. However, in other areas, they form orthogonal sets or show

opposite flow directions, a common feature in confined turbidite settings (Gattacceca & Speranza, 2002; Sinclair & Tomasso, 2002), such as in structurally controlled elongated basins like wedgetop basins (Tinterri & Tagliaferri, 2015; Marini et al., 2016). Based on the paleocurrent analyses for the two formations, preferred orientations have been observed. Generally, both formations show different preferred flow directions, suggesting a non-uniform basin architecture made of the superposition of several turbidite lobes (Critelli, 1987; Guida et al., 1988; Critelli & Le Pera, 1990; Milli et al., 2007). We suggest that the well-represented restored flow directions toward the S, SSE and NNE (Figure 6-8 d, e) are related to the turbidites that flow parallel to the depocenter axis, with a mean N-S direction, accordingly with other paleogeographic reconstructions, e.g., Milli et al. (2007). It is possible to interpret other preferred directions as being at a high angle, converging toward the depocenter axis.

The detrital evolution of the CG deposits indicates a mixed provenance from the CPT, LAC and Apennine Platform, as well as the volcanoclastics from the Miocene synorogenic volcanic arc located westward of the Cilento Basin (Critelli & Le Pera, 1994). It is worth noting as the Numidian sandstones, extensively occurring in the Apennine Platform foredeep deposits and Lagonegro–Molise Basin succession (Vitale et al., 2013b; Vitale & Ciarcia, 2018) and coeval with the Pollica Fm, show a detrital evolution (Figure 6-8 a), indicating a continental block provenance, being very different from CG sandstones, which are characterized by a recycling provenance. Based on this feature, it appears that during the uppermost Burdigalian–Langhian, the Cilento Basin was separated from the Apennine Platform foredeep basin and that any sediment material came from the recycling of the LAC and CPT. Considering the paleocurrent data and clastic provenance, we suggest a simplified model of the CG basin (Figure 6-9) defined by a linear depocenter with a roughly N-S trending axis and clastic supply mainly from the LAC and Apennine Platform carbonates (to the E), LAC and CPT (to the W) and volcanoclastics from a volcanic arc located to the west, likely belonging to the Sardinia block (Savelli, 2002).

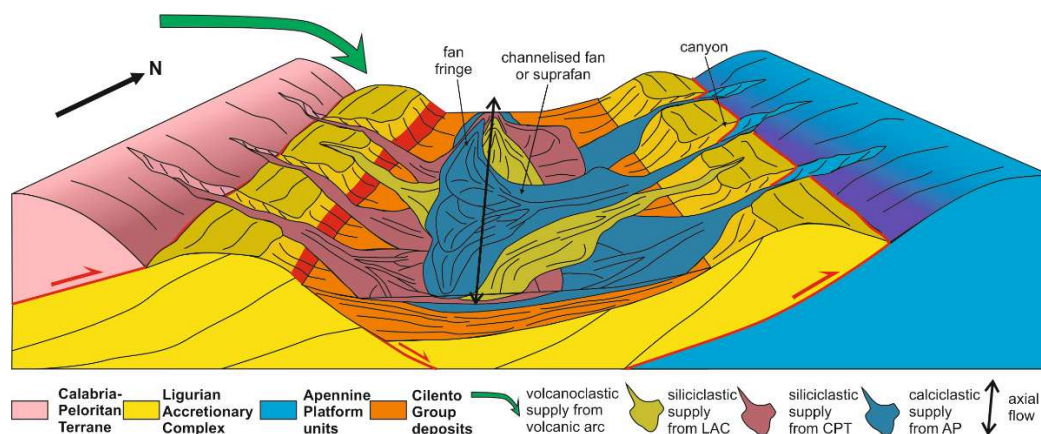


Figure 6-9: 3D model of the Cilento Group wedge-top basin.

The petrography of the clastic content of the CG indicates a detrital evolution from metamorphic-bearing sandstones, including abundant volcanic and sedimentary detritus in the lower part, to plutonic-bearing sandstones in the upper part of the succession (Critelli & Le Pera, 1994). This abrupt change is consistent with the first clastic recycling from the LAC itself and an unroofing of the CPT, with erosion shifting from the uppermost rocks, such as the sedimentary cover, to the deeper plutonic rocks over time.

According to Vitale et al. (2011), the formation of the CG wedge-top basin can be attributed to the extensional collapse of the LAC. This event likely occurred due to the buttressing of the over-thickened orogenic prism against the Apennine Platform in the uppermost Burdigalian. Low-angle extensional detachments associated with synorogenic extension favored the development of accommodation space in wedge-top basin depocentres. This hypothesis has been supported by (Cieszkowski et al., 1995), the Authors of which analyzed the thermal maturity of organic and inorganic fractions of fine sediments of the CG. Their results indicate two rapid sediment accumulation phases during the CG deposition. The first has an age ranging between 16 and 13 Ma (uppermost Burdigalian–Langhian), and the second has an age ranging between 6 and 3 Ma (upper Messinian–Pliocene). A lower accumulation rate separated these two phases. The high accumulation rate of 2.7 mm/yr for the first stage of CG sedimentation is consistent with rapid subsidence, such as that occurring in an extensional rather than a compressional setting (Cieszkowski et al., 1995). On the other hand, the lower rate (1.1 mm/yr) of the second phase could be associated with the out-of-sequence deformation that affected the CG deposits in the lower Pliocene, which featured a renewed orogenic thickening due to the diffuse thrust faults and folds. This possibly includes the regional SW-verging recumbent fold in the Orria–Piano Vetrare area (Figure 6-2 a; Zuppetta & Mazzoli, 1997), probably associated with the back thrusting of the Apennine Platform carbonates onto the CG deposits.

To better explain the sedimentary evolution of the Cilento Basin within the framework of the orogenic evolution of the southern Apennines, we illustrate in the cartoons of (Figure 6-10) the first stages of the Apennine construction from Oligocene to Tortonian. In the Oligocene–Aquitanian (Figure 6-10 a), the ophiolitic successions of the Ligurian Ocean were subducted and added to the subduction channel (Critelli, 2018). In this period, the easternmost sector of the Ligurian Ocean (Nord-Calabrese, Parasicilide and Sicilide units) experienced a foredeep stage. In the early Burdigalian (Figure 6-10 b), the Ligurian successions were piled up by frontal accretion forming the LAC, subsequently (Figure 6-10, c) overthrust onto the Apennine Platform domain, the latter forming a buttress that caused the over-thickening of the LAC followed by a collapse (Figure 6-10), d). On the extended orogenic prism, the Cilento Basin formed and was first filled by the Pollica Fm sediments (Figure 6-10, d) and subsequently by San Mauro Fm sediments (Figure 6-10, e). The latter included a calciclastic supply (“Fogliarina” megabeds) originating from the Apennine Platform units that, in the Langhian–Serravallian, were included in the orogenic prism. Finally, in the upper Tortonian–lower Messinian (Figure 6-10, f), the whole chain was covered by the wedge-top basin Castelvete Group clastic deposits, including the Monte Sacro Fm in Cilento and Gorgoglione and Oriolo Fms in the easternmost sectors overlying the Lagonegro–Molise Basin thrust sheets. In the Calabria sector, to the south, the coeval clastic deposits of the Amantea Group were deposited during the opening of the Tyrrhenian back-arc basin (Vitale et al., 2019a).

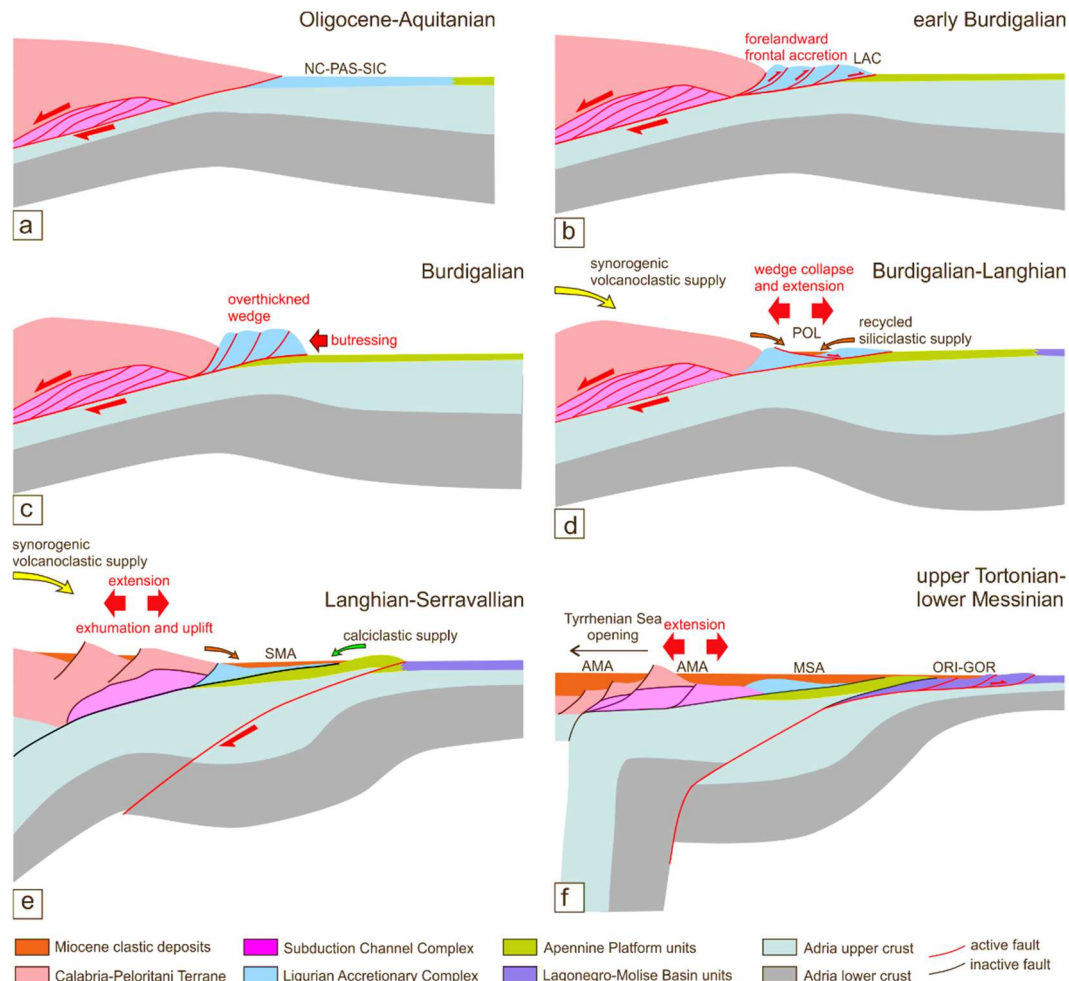


Figure 6-10: Tectonic-sedimentary evolution of the Ligurian Accretionary Complex from Oligocene to upper Messinian. NCA: Nord-Calabrese Domain; PAS: Parasicilide Domain; SIC: Sicilide Domain; LAC: Ligurian Accretionary Complex; POL: Pollica Fm; SMA: San Mauro Fm; AMA: Amantea Group Basin; MSA: Monte Sacro Fm; ORI: Oriolo Fm; GOR: Gorgoglione Fm.

Finally, Figure 6-11, shows the Langhian–Serravallian paleogeographic maps when the Cilento Basin formed on top of the early Apennine orogenic chain. In the Langhian (Figure 6-11 a), the Cilento Basin was fed by different clastic sources, including the CPT and synorogenic volcanoes located in the W and SW, and the LAC both from the W and E, accordingly to the paleocurrent preferred directions of the Pollica Fm turbidites. Synchronously, the whole western-central Mediterranean area was the location of the Numidian sandstone deposition, beginning from the Rif and Betic chains and passing to the Kabyrides, Sicilian Maghrebides and finally to the Lagonegro–Molise Basin, and the foredeep basin of the Apennine Platform (Mattei et al., 1999). In the Serravallian (Figure 6-11 b), the Cilento Basin was fed again by the CPT, LAC and synorogenic volcanoes with the addition of sediments derived from a source located to the E of calciclastics, originating from the erosion of the Apennine Platform thrust sheets that were forming in that period (Vitale & Ciarcia, 2013b). On the map, the synchronous

basin of the Amantea Group is shown. It was forming along the western side of the CPT, being related to the stretching of the chain and anticipating the opening of the Tyrrhenian back-arc basin (Vitale et al., 2019).

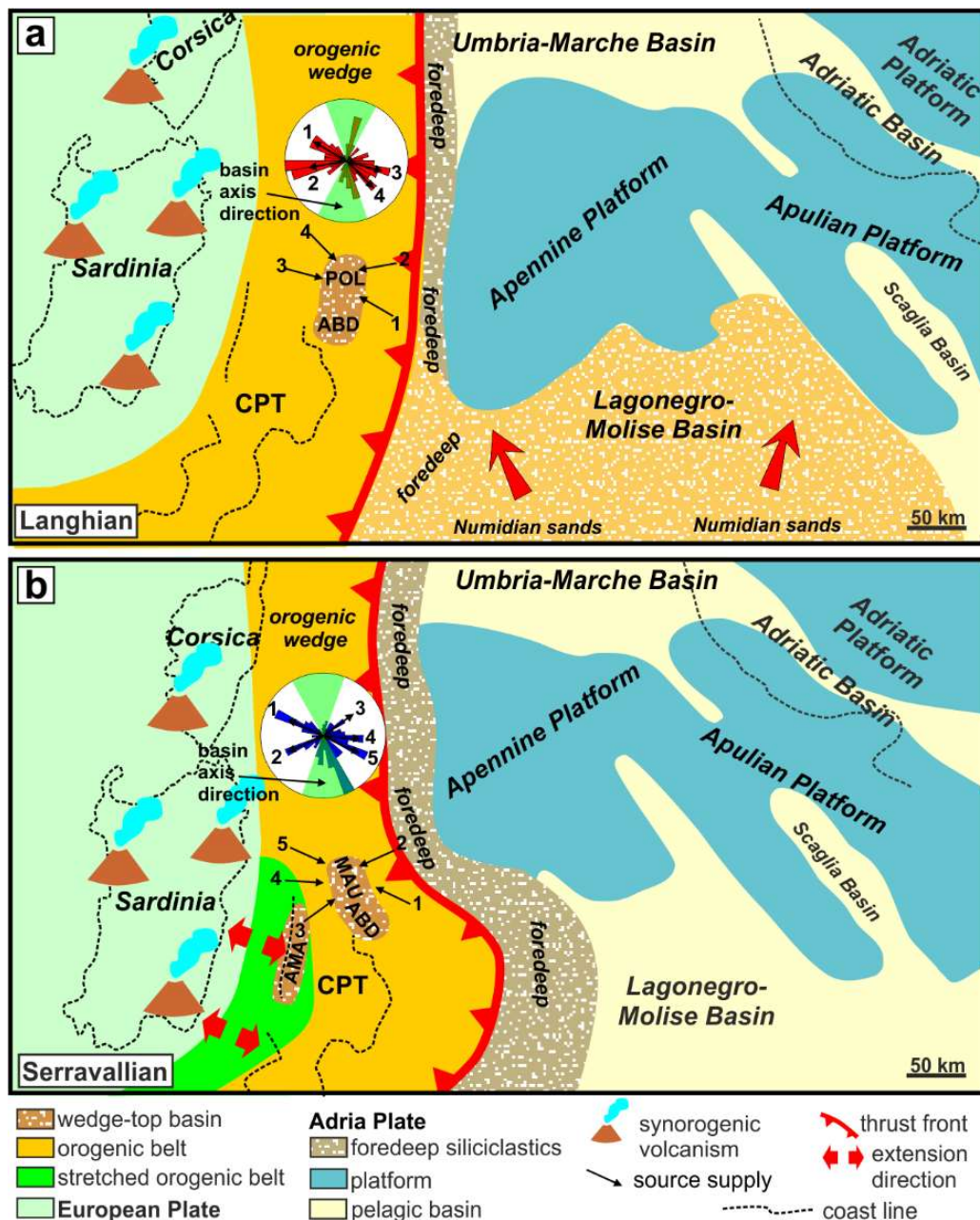


Figure 6-11: Paleogeographic maps of the central sector of the Mediterranean area in (a) Langhian and (b) Serravallian.

6.9 Conclusions

We analyzed the turbidity paleocurrents of the uppermost Burdigalian-lowermost Tortonian Cilento Group in the southern Apennines (Italy). This turbiditic succession consists of

sandstones, conglomerates, marls and pelites grouped in the uppermost Burdigalian-Langhian Pollica Fm and the Serravallian-lowermost Tortonian San Mauro Fm. The CG succession deposited in the first wedge-top basin of the southern Apennine Orogeny formed in an extensional environment resulting from the wedge collapse during the overthrusting onto the western margin of the Apennine Platform.

We analyzed flute and groove casts, flame structures and ripple marks. Due to the intense deformation of the succession, we reconstructed the original attitude restoring the horizontal beds hosting the paleocurrent features. We collected 338 measurements from 74 sites across the whole of Cilento. Results indicate that turbidity paleocurrents form a complex spatial pattern, with some areas with parallel and concordant directions and others with opposite or orthogonal flow directions. This feature is common in tectonic-controlled depressions such as wedge-top basins. When analyzed as a whole, the reconstructed paleocurrent flows show some preferred directions. Paleocurrents in the Pollica Fm, restored to the Langhian paleogeography, are characterized by preferred directions toward W-NW and E to SSE. Paleocurrents in the San Mauro indicate the same previous directions with a dominance of the SSE azimuth about parallel to the basin axis.

According to the detrital evolutions, paleocurrent directions and paleogeographic reconstructions, the sources of the recycled siliciclastic sediments were the CPT, located to W, and the LAC that hosted the Cilento Basin. The volcanoclastic source was the synchronous volcanic arc located in the Sardinia block. Finally, the carbonate supply that formed the “Fogliarina” megabeds in the San Mauro Fm, was originated from the Apennine Platform units located to E, witnessing the development of the first tectonic piling up of the Apennine carbonates in the Serravallia

Chapter 7 Concluding Remarks

The study of the Tempa Roccarossa *mélange*, Ligurian Accretionary Complex, and Cilento Group deposits provides a detailed understanding of the tectonic, deformation, structural, sedimentary, and thermobaric evolution of some key sectors of the southern Apennines. The study offered complementary insights into the complex dynamics of shear zones, fluid migration, and sedimentation, contributing to the region's orogenic development. By synthesizing data on deformation, thermobarometry, and fluid interaction, this discussion highlights the interplay of these processes during the Miocene–Pliocene orogeny.

5.1 Tectonic and Structural Evolution

The Tempa Roccarossa *mélange* and the Ligurian Accretionary Complex both exemplify the tectonic complexity within the southern Apennines. Shear zone-induced deformation played a pivotal role in the evolution of the Tempa Roccarossa *mélange*, with this study documenting multiple deformation stages (D1-D3). The S1, S2, and S3 foliations, together with the formation of porphyroclasts and shear structures, represent the progression from brittle to ductile shear. In parallel, the structural analysis of Ligurian successions revealed folding, boudinage, and thrusting processes, further indicating the development of a complex shear zone.

The Pioppi Shear Zone and the Tempa Roccarossa *mélange* both exhibit a dominant simple shear component, with secondary pure shear and a flattening strain, indicating a similar thinning-widening deformation that accommodates the crustal shortening. The presence of an oblate strain geometry aligns the findings with those of shear zones in other orogenic settings (Xypolias, 2010). This similarity points to the importance of shear zone development in driving tectonic deformation and forming key geological structures in the Apennine orogen.

5.2 Thermal and Metamorphic Evolution

Thermal estimates based on geothermometric data from different methods (Illite crystallinity, Raman geothermometry, and multi-equilibrium technique) provide insights into the deformation and thermobaric history of the region. The Tempa Roccarossa *mélange* exhibited temperature estimates ranging from 200°C to 250°C, corresponding to burial depths of 10–13 km. These values align with the anchizonal metamorphism.

Similarly, the Ligurian Accretionary Complex further confirms anchizonal metamorphism, with temperature estimates ranging from 250°C to 280°C and pressures of 0.3–0.4 GPa, corresponding to burial depths of 11–15 km. These estimates support the interpretation that both the Tempa Roccarossa and LAC units were subjected to similar thermal conditions during the orogenic processes, influenced by tectonic loading.

5.3 Fluid interaction and Veining

Fluid migration in both the Tempa Roccarossa *mélange* and the Ligurian accretionary wedge is an essential component of their deformation and anchimetamorphism. The transformation of smectite to illite, chlorite, and muscovite through fluid-driven reactions is a key feature of the dehydration processes during burial (Emerman & Turcotte, 1983; Moore & Vrolijk, 1992). This process is evident in both the Tempa Roccarossa *mélange* and the Ligurian Accretionary Complex, where vein development is associated with fluid expulsion during tectonic burial. These fluids, migrating through fractures and shear zones, likely influenced the mineral assemblages, contributing to the deformation of the sedimentary rocks in these orogenic regions.

In the LAC units, the presence of fluid-rich veins further emphasizes the role of fluid migration during the tectonic burial phase. The calcite and quartz veins mark significant fluid movement, which likely facilitated the formation and deformation of the basin fill. This fluid movement was crucial in driving the tectonic processes and anchimetamorphism observed in both the Tempa Roccarossa and Ligurian units.

5.4 Burial and Exhumation History

The exhumation history of the Tempa Roccarossa *mélange* and LAC units is an integral part of the overall tectonic evolution. Both orogenic regions experienced significant tectonic burial and subsequent exhumation during the Miocene–Pliocene orogenic events. For the LMB units, the Tortonian–Pliocene phase of tectonic activity is linked to thrusting, delamination, and exhumation, leading to the formation of thrust sheets and the stacking of LMB units on the Apennine Platform. The tectonic exhumation of the LMB units hosting the Tempa Roccarossa *mélange* was achieved both by thrusting and surficial delamination by means of the activity of low-angle normal faults. Deep-seated thrust faults were generated following the thrusting of

the allochthonous wedge over the Apulian Platform. In the case of the Ligurian units, the tectonic exhumation occurred in different stages, including the collapse of the oceanic accretionary wedge during the docking with the Apennine Platform through low-angle normal faults. Hence, there is a similarity between the LAC and the LMB wedge that formed both by thin-skinned tectonics (frontal accretion for the LAC and off scraping for the LMB units) and both were exhumed following a buttressing effect during the docking with a thicker crust.

5.5 Sedimentary Evolution and Paleocurrent Patterns

The paleocurrent analysis of the Cilento Group deposits provides critical insights into the geometry of the wedge-top basin and the sediment supply from multiple sources. The study highlights the heterogeneous paleocurrent patterns observed in the CG deposits, with flow directions ranging from parallel to orthogonal, reflecting the structural complexity of the basin (Gattacceca & Speranza, 2002; Sinclair & Tomasso, 2002). These variations are consistent with confined turbidite settings typical of wedge top basins (Tinterri & Tagliaferri, 2015; Marini et al., 2016). The restored flow directions towards the S, SSE, and NNE are consistent with a linear depocenter aligned along an N-S axis, with sediment input from the Ligurian Accretionary Complex, Apennine Platform, and synorogenic volcanic sources to the west (Milli et al., 2007).

The detrital evolution of the CG deposits suggests a mixed provenance, with contributions from the CPT, LAC, and volcanoclastics. This indicates that the Cilento Basin evolved through sediment recycling, with a shift in provenance occurring during the uppermost Burdigalian–Langhian period. The transition from continental block provenance to a recycling provenance points to a shift in the tectonic regime, marking the separation of the Cilento Basin from the Apennine Platform foredeep (Cieszkowski et al., 1995; Vitale et al., 2011). This sedimentary evolution reflects the tectonic collapse of the Ligurian Accretionary Complex (LAC) and the subsequent development of the Cilento wedge-top basin.

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