Michard *et al. Swiss Journal of Geosciences* (2022) 115:19
https://doi.org/10.1186/s00015-022-00419-8

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Swiss Journal of Geosciences

ORIGINAL PAPER

Open Access

The Maira-Sampeyre and Val Grana Allochthons (south Western Alps): review and new data on the tectonometamorphic evolution of the Briançonnais distal margin

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Abstract

Here we describe the structure, the high-pressure, low-temperature (HP-LT) metamorphism and tectonic evolution of the Briançonnais distal margin units from the south Western Alps. The studied area extends southwest of the Dora-Maira (U)HP basement units and east-southeast of the classical Briançonnais nappes. A new structural map accompanied by geological profiles shows the thrusting of the oceanic nappes (Monviso and Queyras units) onto the distal Briançonnais units (D1 and D2 late Eocene deformation phases) under blueschist-facies conditions. Subsequent deformation during the Early Oligocene (D3 deformation phase) took place under greenschist-facies conditions and was associated with back-folding and -thrusting in the units overlying the Dora-Maira massif and with exhumation related to normal reactivation of former thrusts within the latter massif. Two large cover units, detached from their former distal Briançonnais basement, are redefined as the Maira-Sampeyre and Val Grana Allochthons (shortly: Maira-Grana Allochthons = MGA) including, (i) the Val Maira-Sampeyre unit involving Lower and Middle Triassic formations, seemingly detached from the Dora-Maira units during the subduction process, and (ii) the Val Grana unit with Middle-Upper Triassic and Early-Middle Jurassic formations, which was probably detached from the Maira-Sampeyre unit and correlates with the "Prepiemonte units" known from the Ligurian Alps to the Swiss Prealps. Three major shear zones involving tectonic mélanges of oceanic and continental rocks at the base of the Val Grana, Maira-Sampeyre and Dronero units testify to an early phase of exhumation within the subduction channel in front of the Adria plate. We present a new metamorphic map based on published and new petrological data, including new thermometric data obtained by Raman spectroscopy of carbonaceous material (RSCM). The T_{RSCM} values range from ~400 °C to >500 °C, going from the most external Val Grana unit and overlying Queyras schists to the uppermost Dora-Maira unit. During the Late Triassic, the width of the Briançonnais *s.l.* domain can be restored at ~100 km, whereas it reached ~150 km after the Jurassic rifting. A significant, second rifting event affected the Briançonnais domain during the Late Cretaceous-Paleocene, forming the Longet-Alpet chaotic breccias, which deserve further investigations.

Keywords: Briançonnais distal margin, Jurassic rifting, Cretaceous-Paleocene extension, Metamorphic wedge, Raman spectroscopy of carbonaceous material (RSCM), Exhumation, High-pressure metamorphism

Editorial handling: Vladica Cvetković.

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«Je regarde [...] les observations de détails comme l'unique base d'une connoissance solide mais je voudrois qu'en observant ces détails, on ne perdit jamais de vue les grandes masses et les ensembles et les ensembles et que la connoissance des grands objets et de leurs rapports fût toujours le but que l'on se proposât en étudiant leurs petites parties.»—H.B. de Saussure, 1779, p. 3

1 Introduction

Being in the center of Europe and remarkable for their wonderful summits (Mont Blanc, 4807 m a.s.l.), the Western Alps captivated European geologists and mineralogists from the late eighteenth century (De Saussure, 1779) onward. At the advent of plate tectonics in the latest 1960's, the geological study of the Western Alps took a new start. The close association of ophiolites and continent-derived crustal units and the processes of thrusting and metamorphism in the internal zones of the belt became more intelligible than in the time of Argand's pioneering work (Argand, 1911, 1924). The description of the Western Alps belt (Fig. 1) in terms of rifting, drifting, subduction and collision began in the 70's (see Frisch, 1979 and references therein), then benefited from deep seismic profiles, followed by tomographic studies in the following decades (e.g., Roure et al., 1990, 1996; Schmid and Kissling, 2000; Lippitsch et al., 2003; Handy et al., 2021). Knowledge about present-day oceanic ridges and passive margins was applied to understand Alpine ophiolites and inverted Mesozoic margins (Lemoine et al., 1985, 1986; Lemoine and Trümpy, 1987; Lagabrielle and Lemoine, 1997). The general occurrence of high-pressure (HP), low-temperature (LT) metamorphism was recognized in the Internal (Penninic) zones (Bearth, 1959, 1966). This was followed by the discovery of ultra-high-pressure (UHP) metamorphism in a continental crustal slice of the southern Dora-Maira massif (Chopin 1984, 1987; Goffé and Chopin, 1986), and in the Zermatt metaophiolites further to the north (Reinecke, 1991). Geochronological studies established that UHP-HP metamorphism developed progressively from the Late Cretaceous to the Eocene (Duchêne et al., 1997; Hunziker et al., 1992; Monié, 1990; Rubatto et al., 1999; Tilton et al., 1991), coeval with the subduction of Alpine Tethys and European margin lithosphere beneath the Africa-related Adria microplate and followed by rapid cooling and decompression (Rubatto and Hermann, 2001; Avigad et al., 2003).

During the last two decades, several papers deciphered the most critical areas of the Western Alps (e.g., Agard et al., 2001; Dal Piaz et al., 2001; Le Bayon and Ballèvre, 2006; Gasco et al., 2011; Compagnoni et al., 2012; Angiboust et al., 2012; Manzotti et al., 2014, 2018; Weber et al., 2015; Ballèvre et al., 2018, 2020; Groppo et al., 2019). Other papers proposed structural, tectonic and/or metamorphic syntheses of Western Alps in the frame of other parts of the Tethyan belt (e.g., Michard et al., 1996; Oberhänsli et al., 2004; Schmid et al., 2004, 2017; Rosenbaum and Lister, 2005; Diehl et al., 2009; Handy et al., 2010; Bousquet et al., 2008, 2012; Dumont et al., 2011; Kissling and Schlunegger, 2018; Le Breton et al., 2020; Agard, 2021; Gnos et al., 2021). The structure of the belt is roughly deciphered down to 100 km depth (Fig. 1C) and its paleogeography restored in its major aspects (Fig. 1B). Indeed, the Western Alps have been and still are the "mother" of subduction-collision orogens, and its HP-LT Internal Zones are characteristic of "Ampferer-type", amagmatic subduction of a narrow oceanic domain and adjacent thinned continental crust, opposite to the "Benioff-type", arc-related subduction of a large ocean (Agard, 2021; McCarthy et al., 2020). Regarding the dynamics of plate convergence, the Alps fall into the category of slab-pull orogens (Faccenna et al., 2013; Jolivet et al., 2021), consistent with the narrowness of the subducted oceanic lithosphere.

Despite the permanent research activity in the belt, a small but critical area (framed in Fig. 1A) of the Internal Zones remained unexplored since the late 1960's. This area of the Cottian Alps (the southern part of the French-Italian Alps between the Torino and Cuneo transects) is crossed by the valleys of the Maira and Grana rivers, and mainly occupied by Triassic-Jurassic series that tectonically overlie the southern Dora-Maira units. The area has been made famous by the work of Franchi (1898) describing the occurrence of determinable Liassic ammonites in the lowest "calcescisti lucenti" (in French "Schistes lustrés", which are glittery calcareous-clastic metasediments) above Upper Triassic dolostones (similar to the "Hauptdolomit facies" of Eastern Alps) in Val Grana. One of us (A.M.) mapped the area at scale 1:50,000 in the latest 50's and published a geological monograph (Michard,

(See figure on next page.)

Fig. 1 **a** Sketch tectonic map of the Western Alps with location of the studied Maira-Sampeyre and Grana Allochthons (MGA), after Schmid et al. (2017), modified, with traces of the seismic profiles ECORS-CROP, NFP20-W and CIFALPS. Outline of the geophysical Ivrea body after Ceriani et al. (2001). Eclogite domain boundary after Malusà et al. (2011). Ac: Acceglio village; Alp Mar: Alpes maritimes; Arg: Argentera-Mercantour; A-C: Arnasco-Castelbianco; Amb: Ambin; Ar: Arvieux; Bel: Belledonne; Can: Canavese Line Ch: Chaberton-Grande Hoche; DB: Dent Blanche; DM: Dora-Maira; Em-Ub: Embrunais-Ubaye; G: Gondran; GM: Grande Motte; GP: Gran Paradiso; IL: Insubric (Peri-Adriatic) Line; MH: Monferrato Hills; MR: Monte Rosa; Mrg: Marguareis; Mt Bl: Mont Blanc; NB: Nappe de la Brèche; Pel: Pelvoux; PF: Penninic Front; R: Rochebrune; RB: Roc del Bouchet; RC: Roche des Clots and Péouvou units; SVL: Sestri-Voltaggio Line; TH: Torino Hills; Vi, Viso; Z: Zermatt. **b** Restoration of the Alpine domain in cross-section before plate convergence, after Manzotti et al. (2014) and Ballèvre et al. (2020). **c** Sketch lithospheric-scale cross-section along the CIFALPS transect, after Malusà et al. (2021), modified after Michard et al. (2004), Schmid et al. (2017), Ballèvre et al. (2020). Acronyms as above with Br: Briançonnais; SL: Schistes lustrés

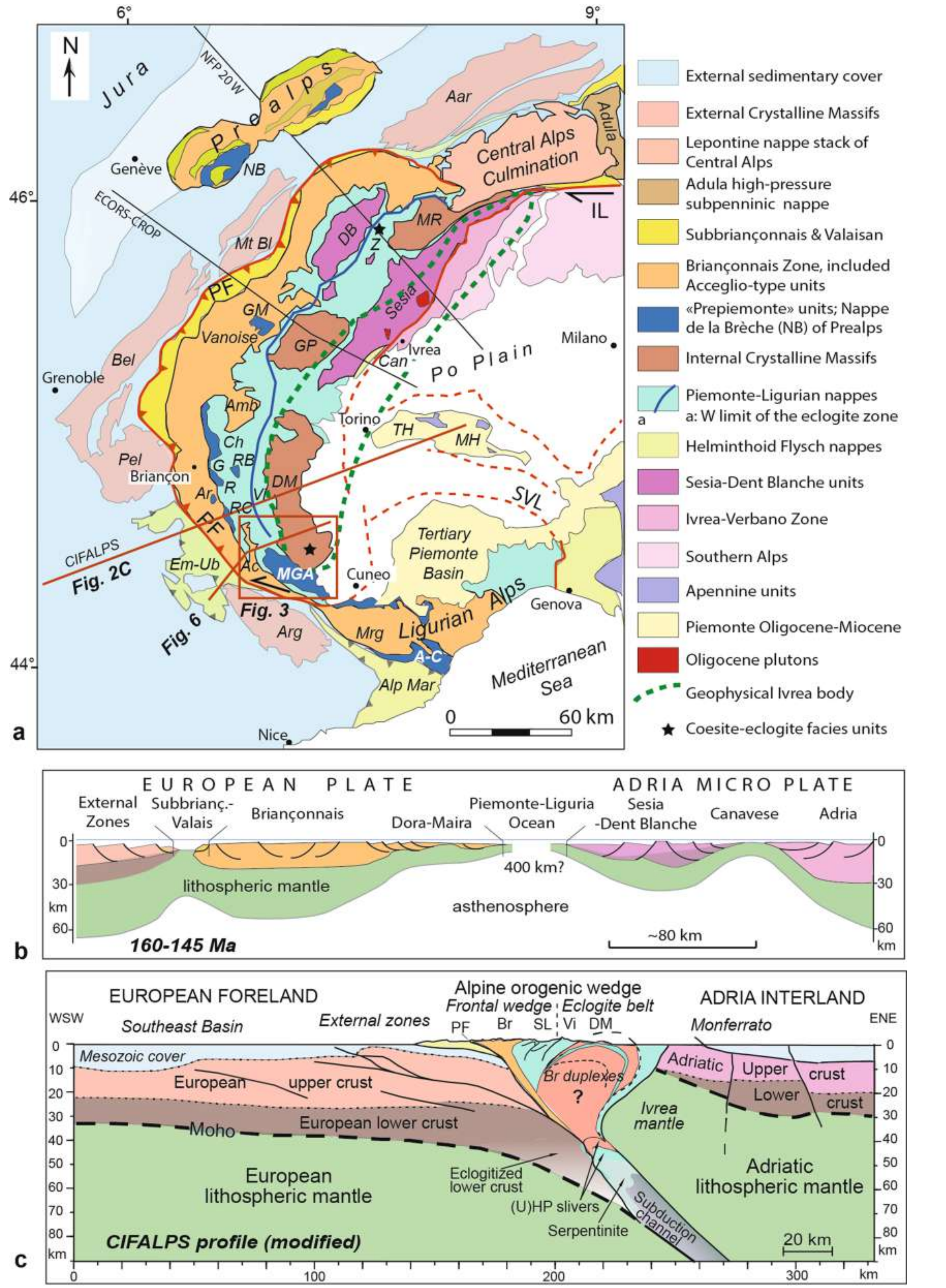

Fig. 1 (See legend on previous page.)

1967) immediately before seminal publications founded the paradigm of plate tectonics (Isacks et al., 1968; Le Pichon, 1968). Michard (1967) proposed a stratigraphic continuity between Liassic "Schistes lustrés" (linked to the shallow-water, epicontinental Triassic deposits) and overlying "Schistes lustrés", which host lenses of serpentinites and metabasites and are rich in metashales and metacherts. Soon afterwards this stratigraphic interpretation was questioned (Lemoine, 1971; Schumacher, 1972; Michard and Schumacher, 1973) as it could not be reconciled with the growing wealth of new observations in support of an oceanic origin for the ophiolite-bearing "Schistes lustrés" in terms of the sedimentary cover of an oceanic basement (Piedmont-Liguria, or Piemonte-Liguria Ocean), formed during the Late Jurassic-Cretaceous (Lemoine et al., 1970; Elter, 1971; Decandia and Elter, 1972; Lemoine et al., 1978; Lagabrielle, 1981; Lemoine et al., 1986). Unfortunately, the Grana and Maira valleys have not been re-studied by modern geologists after the 1970's, except for a survey of the metamorphism by Bousquet et al. (2008) and a map of the northern border of Val Maira by Mondino (2005).

The present paper first of all presents a critical review of the geology of this key region in the light of modern geological knowledge. Additionally, we also present new field and laboratory results, particularly concerning metamorphism, including a set of Raman Spectroscopy of Carbonaceous Material (RSCM) data for the first time. We pay special attention to the dominantly Triassic and Liassic units cropping out along the Maira and Grana valleys, hereafter labeled as the *Maira-Sampeyre and Val Grana (in short, Maira-Grana) Allochthons* (MGA) and to their relationships with the surrounding tectonic units. These are the southern Dora-Maira continental units to the northeast (interpreted as the most distal parts of the Briançonnais terrane; Ballèvre et al., 2020); the Monviso ophiolites and Queyras ophiolite-bearing Schistes lustrés (Piemonte-Liguria Ocean) to the northwest and west, and the Briançonnais units further to the west and southwest. In the studied area the latter are mainly represented by units derived from the "Acceglio Zone", here labeled Acceglio sensu lato (*s.l.)* paleogeographic domain, deprived of Triassic carbonates (Debelmas and Lemoine, 1957; Lemoine, 1960; Lefèvre and Michard, 1976; Michard et al., 2004).

The combination of published and new structural and petrological-thermometric data presented below allow us to discuss the relationships of the Maira and Grana allochthonous units (MGA) with other similar, so-called "Prépiémontais" (Prepiedmont, Prepiemonte), Triassic-Liassic units (Deville, 1986; Ellenberger, 1958; Lemoine, 1967; Pantet et al., 2020) as well as with the Schistes lustrés and various units derived from the Briançonnais sensu lato domain (Acceglio *s.l.* and Dora-Maira units). In the discussion, we will pay special attention to the extensional events that took place from the Triassic up to Late Cretaceous-Eocene in the Briançonnais domain *s.l.*. These events that framed the Briançonnais passive margin remain poorly studied so far with respect to those that framed the conjugated Adria margin (e.g., Bertotti et al., 1999; Manatschal and Bernoulli, 1999; Mohn et al., 2010, 2012; Sutra and Manatschal, 2012; Beltrando et al., 2015; Chenin et al., 2017).

2 Geological setting

It is recognized that the complex structure of the Western Alps (Fig. 1A) formed at the expense of the following three major domains of the Mesozoic paleogeography (Fig. 1B) inherited from the Pangea breakup (Handy et al., 2010; Schmid et al., 2017; Kissling and Schlunegger, 2018; Le Breton et al., 2020).

The External Zones (Helvetic-Dauphinois-Provençal zones) are derived from the proximal part of the European passive margin, which comprises a crustal basement exposed in the uplifted External Crystalline Massifs, and a detached, folded Triassic-Neogene cover (Fig. 2A; Bellahsen et al., 2014; Nibourel et al., 2021). The External Zones are bounded to the east by the major out-of-sequence thrust of the Internal or Penninic Zones ("Front Pennique", FP; Ceriani et al., 2001) of Oligocene age (equivalent of the "Briançonnais Frontal Thrust" first described by Tricart (1986), which connects to the NE with the Rhône-Simplon detachment (Campani et al., 2014) bounding the Central Alps culmination or Lepontine Dome, and to the SE with the sinistral Argentera-Cuneo Line (Schmid et al., 2017). At an earlier stage (late Eocene) some internal Penninic units were already thrust over the External Zones west and in front of the FP, namely the Helminthoid Flysch nappes of the Embrunais-Ubaye and Alpes Maritimes, and the Prealps nappes north of Mont Blanc.

The *Briançonnais paleogeographic domain* (sensu lato) represents the distal part of the thinned European margin that was separated from the proximal part by the partly oceanic, SW-ward narrowing *Valais-Subbriançonnais* rift during the Cretaceous (Beltrando et al., 2012; De Broucker et al., 2021; Loprieno et al., 2011). On the scale of lithospheric plates, the Briançonnais *s.l.* domain is regarded as a terrane rifted from Europe during the Early Cretaceous, loosely connected to Corsica, and displaced NNW-ward by ~ 200 km along the proximal European margin during the Eocene (Frisch, 1979; Stampfli et al., 1998; Schmid and Kissling, 2000; Handy et al., 2010). Units derived from the Briançonnais *s.l.* domain form the classical Briançonnais nappes of the Briançon area (Briançonnais sensu stricto) and its northern equivalents,

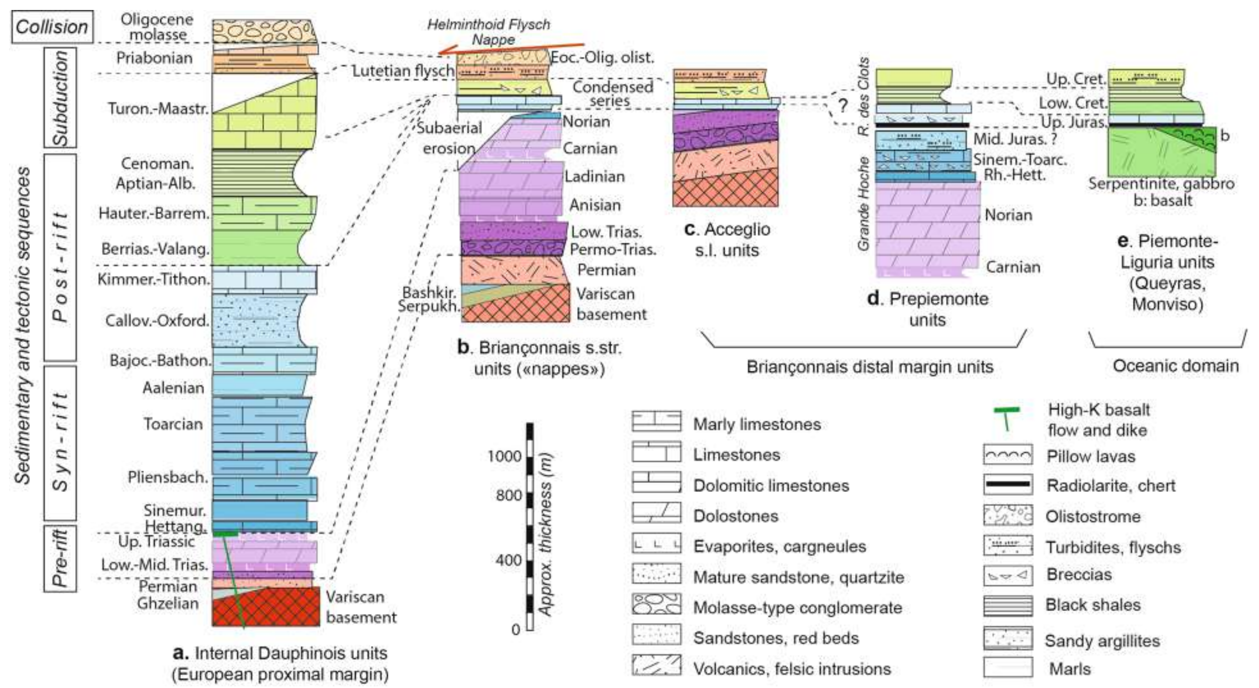

Fig. 2 Stratigraphic columns of the main paleogeographic domains exposed in the southern Western Alps. **a** European proximal margin (Dauphinois) after Lemoine et al. (1986). **b** Briançonnais *s.str.* (Classical Briançonnais of Briançon area). **c** Acceglio *s.l.* units after Michard et al. (2004). **d** Prepiemonte units after Dumont et al. (1984). **e** Piemonte-Liguria oceanic domain after Lemoine et al. (1986) and Deville et al. (1992)

the Siviez-Mischabel Nappe (Genier et al., 2008; Pantet et al., 2020; Scheiber et al., 2013). The Mesozoic-Eocene series of the Briançonnais *s.str.* are characterized by thick Triassic carbonates (Fig. 2B; Debelmas and Lemoine, 1957; Lemoine, 1960, 1961; Deville, 1986), which have been partly eroded during the late Early-Middle Jurassic emersion that affected most of the Briançonnais domain (Debelmas, 1987; Lemoine et al., 1986). The Acceglio *s.l.* tectonic units (also labeled in the literature "Ecailles intermédiaires" after Lemoine (1961), and "Ultrabriançonnais" after Lefèvre and Michard, 1976) whose Triassic carbonates have been totally or almost totally eroded (Fig. 2C) are generally located in a more internal position with respect to the units derived from the Briançonnais *s. str.* They are in close neighborhood with other units whose facies differs from that of the Briançonnais *s.str.* in being characterized by thick Upper Triassic dolostones followed by Liassic and Dogger (?) syn-rift series rich in bedded carbonate breccias then sands (Fig. 2D). Since the latter series have been regarded as transitional to the "Piémontais" ophiolitic domain they were labeled "Prépiémontais" units (Ellenberger, 1958; Lemoine, 1967)—a classical terminology with varied subsequent adaptations (see for details Additional file 1). Hereafter we use the term "Prepiemonte" for these units, which might include in some cases Upper Jurassic to Upper Cretaceous deep-sea sedimentary facies (Dumont et al., 1984) but clearly belong to the Briançonnais *s.l.* domain. Although separated from the Classical Briançonnais units (Briançonnais *s.str.*) by the ophiolitic Schistes lustrés nappes in map view (Fig. 1), the Internal Crystalline Massifs are widely regarded as derived from the pre-Triassic basement of the Briançonnais domain s.l. (e.g., Ballèvre et al., 2018). The lowermost, monometamorphic units of the Gran Paradiso and Dora-Maira massifs (Money and Pinerolo units, respectively) compare with the "Zone houillère" of the classical Briançonnais *s.str.* (e.g., Manzotti et al., 2015, 2016; Ballèvre et al., 2018, 2020). They are overlain by units involving both monometamorphic orthogneisses and polymetamorphic schists, which derive from a Variscan basement. The Dora-Maira polymetamorphic rocks record a pre-Alpine amphibolite-facies Barrovian metamorphism (Vialon, 1966; Chopin et al., 1991; Sandrone et al., 1993). In the eclogitic unit of northern Dora-Maira (Muret unit), Nosenzo et al. (2021) constrain the pre-Alpine *P*–*T* evolution from about 4–5 kbar, 500 °C to 6–7 kbar, 650 °C, consistent with Barrovian metamorphism, which they date ~ 325 Ma. The Acceglio *s.l.* derived units, most Prepiemonte type units and the Internal Crystalline Massifs were subducted and affected by HP to UHP metamorphism. In terms of their paleogeographic setting, they are grouped hereafter under the name "Briançonnais distal margin".

The Briançonnais *s.str.*, Acceglio *s.l.* and Prepiemonte units continue southeast of the Argentera-Cuneo Line in the Ligurian Alps, although in a different structural setting (Seno et al., 2005). Units derived from the Briançonnais and Prepiemonte domains are also recognized in the French-Swiss Prealps nappes, where they correspond to the "Médianes Rigides" and "Nappe de la Brèche", respectively (e.g., Lemoine, 1967; Mosar, 1989). Notice that these rootless, non or hardly metamorphic Prealps nappes also include Subbriançonnais-Valaisan and Piemonte-Liguria elements ("Médianes plastiques", ophiolites and flyschs, respectively).

The *Piemonte-Liguria domain* whose stratigraphy is summarized in Fig. 2E includes three complexes of units that all derive from the Piemonte-Liguria Ocean (northeastern Alpine Tethys; Stampfli et al., 1998). These complexes of units differ by their metamorphic grade (Agard, 2021). The most internal, i.e., the Monviso and Zermatt complexes, are eclogitic, the latter being locally affected by UHP metamorphism (Reinecke, 1991). The Queyras and Tsaté complexes are more external and their metamorphism ranges within the blueschist-facies (Deville et al., 1992; Agard et al., 2001; Bousquet et al., 2008, 2012; Manzotti et al., 2021). Lastly, the Gets nappe of the French-Swiss Prealps and Chenaillet ophiolitic klippe east of Briançon are hardly affected by greenschist-facies metamorphism. According to Manatschal et al. (2011) and Li et al. (2013), the Chenaillet ophiolite is derived from the western part of the Piemonte-Liguria Ocean margin. However, since it lacks significant Alpine deformation and because it tectonically overlies the Queyras complex, Schmid et al. (2017) inferred an upper plate position of this klippe in an intra-oceanic subduction scenario, and hence placed the Chenaillet ophiolites at the eastern margin of the Piemonte-Liguria Ocean. The ophiolites of the Gets nappe, only metamorphosed under sub-greenschist facies conditions (Bill et al., 2001a) originate from an oceanic branch of Alpine Tethys (Elter et al., 1966) that formerly extended between the Sesia continental extensional allochthon and the distalmost Adria margin (Canavese; e.g., Beltrando et al., 2015; Festa et al., 2020).

The Piemonte-Liguria Ocean opened slowly during the late Bathonian to late Kimmeridgian at least (De Wever and Caby, 1981; Bill et al., 2001b; O'Dogherty et al., 2006; Cordey and Bailly, 2007) and reached a width of ~ 400 km during the Early Cretaceous (Stampfli et al., 1998; Handy et al., 2010, their Fig. 8b; Li et al., 2013; Le Breton et al., 2020), although higher estimates have also been proposed (Agard, 2021; Vissers et al., 2013). The protoliths of the metasedimentary series associated with the ophiolite bodies typically derive, from bottom to top, from ophiolitic breccias, radiolarites, micritic limestones, black shales, and calcareous-pelitic turbidites (flysch), ranging in age from Bathonian to Late Cretaceous, respectively (Fig. 2E; Lemoine and Tricart, 1986; Deville et al., 1992; Lemoine, 2003; Li et al., 2013). These metasediments are the genuine, most extended "Schistes Lustrés", contrasting with the restricted Schistes lustrés-like Prepiemonte sequences, as quoted above (Sect. 1). The highest unit in the Vanoise area, namely the Pointe du Grand Vallon unit, is a blueschist-facies meta-flysch devoid of any metabasite, which yielded planktonic foraminifera of upper Maastrichtian age; it could be an equivalent of the Helminthoid Flyschs of Piemonte-Liguria origin (Deville et al., 1992).

West of the southern Cottian Alps, the Upper Cretaceous *Helminthoid Flysch* constitutes the main part of the *Embrunais (-Ubaye) nappes* (Fig. 2). Subbriançonnais and Briançonnais slivers form the lowest units of these nappes that emplaced upon the External Zones by the end of the sedimentation of the Oligocene foredeep deposits (Kerkhove et al., 1984; Dumont et al., 2011). The Embrunais Helminthoid Flyschs would originate from the innermost Piemonte-Liguria domain and belong to the upper plate in the subduction setting like the Chenaillet klippe; their thick, turbiditic sedimentation is thought to record the subduction onset at ~ 90 Ma (Stampfli et al., 1998).

The Africa-derived *Adria microplate* (Argand's "African promontory") occupies an upper-plate position in the Alpine orogen east and north of our area of interest. Along the Central and Eastern Alps transect, upper crust and Mesozoic cover of the proximal margin of Adria are exposed in the Southern Alps, and the Ivrea Zone exposes its verticalized lower crustal and uppermost mantle underpinnings that originate from the Adria distalmost margin (Handy et al., 1999). The occurrence of a thick slice of mantle, i.e., the Ivrea mantle wedge (or "geophysical body") belonging to the thinned Adria margin beneath the outcropping parts of the Ivrea Zone, and further to the south in the subsurface underneath the Dora-Maira units, is well constrained by geophysical data (Fig. 1A; Berckhemer et al., 1968; Masson et al., 1999; Schmid and Kissling, 2000; Zhao et al., 2015, 2016; Schmid et al., 2017). The distal part of the Adria margin is also exposed in the Lower and Middle Austroalpine nappes of the Eastern Alps close to the Central Alps culmination (Manatschal and Bernoulli, 1999; Mohn et al., 2010), and in the strongly dismembered sediments of Canavese Zone belonging to the Southern Alps (Beltrando et al., 2015; Elter et al., 1966; Ferrando et al., 2004; Festa et al., 2020). The Sesia-Dent Blanche nappes, which constitute the highest tectonic elements of the Western Alps northwest of the Canavese Zone (or Line), probably derive from an extensional allochthon detached from the Adria distal margin since they underwent high-pressure metamorphism (Manzotti et al., 2014).

The most internal parts of the Western Alps arcuate belt do not crop out south of the area of the Sesia Zone, being covered by the deposits of the Tertiary Piemonte basin (Carrapa and Garcia-Castellanos, 2005). The Early Oligocene, lowermost beds of this basin unconformably overlie Briançonnais-derived crystalline basement and the roots of the ophiolites of the Western Alps, including the ophiolitic units of the Ligurian Alps, as well as much of the NE-verging Apenninic units (Fig. 1A; Lorenz, 1986; Mosca et al., 2010; Molli et al., 2010; Maino et al., 2013; Marroni et al., 2017; Piana et al., 2021). Ligurian flysch units of the frontal Apennines (Elter et al., 1966) crop out in the Monferrato-Torino Hills as a result of north-directed Mio-Pliocene thrusting and folding that also affected the Ligurian Alps (Schmid et al., 2017).

3 Structural data

3.1 Methodology for drawing a new structural map

The outlines of the major lithological units presented in a new structural map of the southern Cottian Alps (Fig. 3) were taken from the Geological/Environmental map of Piemonte, scale 1:250,000, compiled by Piana et al. (2017). Our input consisted in delineating the various tectonic units based on their distinct structural and petrological characters (see Sect. 3.2). For this purpose, we used the following maps at the scale 1:50,000: (i) the map of the Maira and Grana valleys published by Michard (1967) and complemented by the unpublished thesis of Schumacher (1972) in the upper Val Grana, and (ii) the Larche and Aiguille de Chambeyron sheets of the Geological map of France (Gidon, et al., 1977, 1994) for covering the western fringes of the studied area, which include the Acceglio-Longet antiformal band and the overlying Schistes lustrés (also mapped by Lefèvre and Michard, 1976, and Lefèvre, 1982). However, in contrast to the map of Michard (1967), we defined tectonic contacts by separating continental sedimentary sequences that include Triassic carbonates at their base from juxtaposed oceanic sequences that include serpentinites, despite of the apparent structural and metamorphic continuity of the units. Such tectonic contacts are frequently well-marked by lenses of serpentinites or cargneules (also spelled cornieules = evaporitic tectonic breccias; Masson, 1972; Fudral et al., 2010). Admittedly, where such markers are lacking, tracking a tectonic limit between ocean-derived Schistes lustrés (Cretaceous) and Schistes lustrés (Jurassic) that are part of the Prepiemonte sedimentary sequence remains uncertain in places.

The mapped area comprising the MGA was extended to the southern Dora-Maira units (Val Varaita and Val Po) in the NE, to the southern Monviso complex in the NW and to the Acceglio *s.l.* units in the W and SW. The tectonic contacts within the Dora-Maira units and westerly adjacent Monviso ophiolite and Queyras Schistes lustrés were taken from the literature (see below). The Geological map of Italia, scale 1:100,000, sheet Dronero-Argentera, 2d edition (Malaroda et al., 1971) was also consulted; it does not differ substantially from Michard (1967) in the MGA area, and its background is consistent with that of the 1:250,000 Piemonte map (Piana et al., 2017).

3.2 Tectonic units and their mutual relationships

Figure 3 illustrates the major tectonic units in map view and cross-sections are shown in Figs. 5 and 6. Metamorphic data are presented in map view and cross-section in Figs. 10 and 13, respectively. Table 1 will help the reader with a summary of the nomenclature used hereafter (for more details, please see Additional file 1).

The *Sanfront-Pinerolo (SP)* is the lowermost of the Dora-Maira units (Argand, 1911; Vialon, 1966), and then of the whole studied area. The SP unit includes graphite-rich conglomeratic schists considered for long as Carboniferous, which indeed yielded a most recent zircon population with an age of ~ 330 Ma (Manzotti et al., 2016). These metasediments are intruded by fine-grained orthogneiss derived from Permian granodiorite, locally dated at about 290 Ma (Bussy and Cadoppi, 1996). Additionally, conglomeratic quartzite and meta-dolostones and marbles occur in the Sanfront-Monte Bracco area, referred to Permo-Triassic clastics and Triassic carbonates, respectively; these are considered as relics of the Mesozoic cover of the SP unit. The Alpine metamorphic grade reached garnet-blueschist to eclogite facies conditions (Avigad et al., 2003; Groppo et al., 2019).

The *Brossasco-Isasca (BI) unit* is made up of a polymetamorphic complex of mica-schists, marbles and metabasites that are associated with mono-metamorphic orthogneisses derived from Permian granites, whose metasomatized parts yielded the famous coesite-bearing lenses of pyrope whiteschists (Chopin, 1984; Chopin et al., 1991; Kienast et al., 1991; Schertl et al., 1991; Henry, 1990; Biino and Compagnoni, 1992; Henry et al., 1993; Chopin and Schertl, 1999; Compagnoni & Rolfo, 2003; Castelli et al., 2007, 2014; Campomenosi et al., 2021). This UHP unit overlies the HP Sanfront-Pinerolo unit along a former major thrust contact that was overprinted by ductile late extensional structures indicating top-SW normal faulting (Fig. 6) under more moderate P–T conditions (Avigad et al., 2003). The small "San Chiafreddo unit", defined by Compagnoni et al. (2012), is mapped here as part of the BI unit.

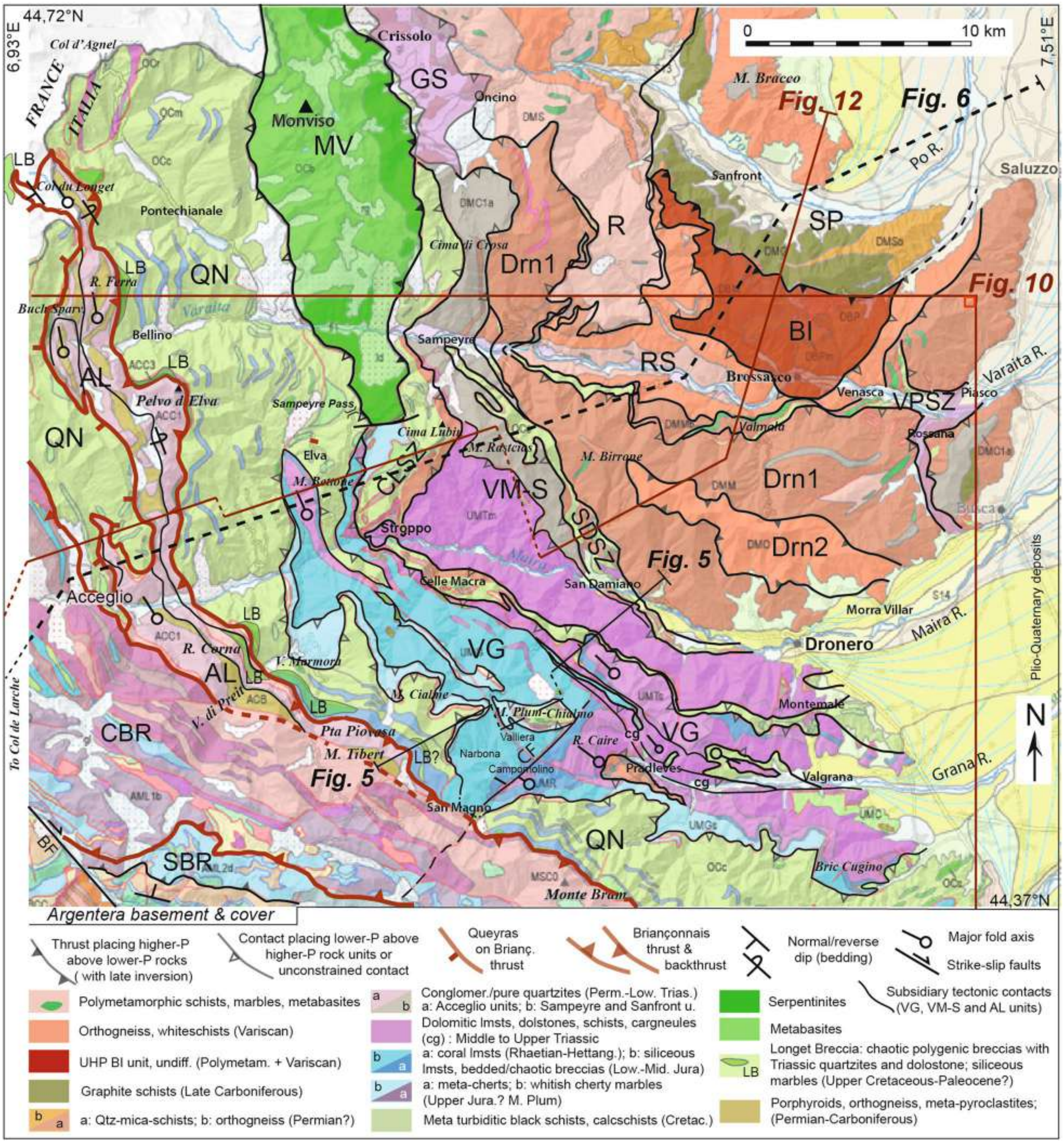

Fig. 3 Structural map of the southern Cottian Alps with traces of cross-sections Figs. 5, 6, 12 and location of map Fig. 10. Background map: Piemonte geological map: 250,000 (Geoportale Arpa; Piana et al., 2017), modified. Cargneule (cg) is only distinguished in the Pradleves fault zone. See Table 1 for details on tectonic units and shear zones. Acronyms: AL, Acceglio *s.l.* units (mainly Acceglio-Longet band); BF: Bersezio fault; BI, Brossasco-Isasca UHP unit; CBR: Classical Briançonnais units (Briançonnais *s. str*. domain); CF: Colletto fault; CLSZ, Cima Lubin shear zone; GS, Giulian-Sea Bianca unit; Drn, Dronero unit (1, lower; 2, upper); LB, Longet-type breccias; MV, Monviso complex (eclogite-facies metaophiolites and metasediments); QN, Queyras nappe (blueschist-facies ophiolitic Schistes lustrés); RS, Rocca Solei unit; R, Ricordone unit; SBR, Subbriançonnais units; SDSZ, San Damiano shear zone; SP, Sanfront-Pinerolo unit (Carboniferous); VG, Val Grana unit (Middle Triassic-Jurassic); VM-S, Val Maira-Sampeyre unit (Permo-Triassic to Middle Triassic); VPSZ, Valmala-Piasco shear zone

Table 1 Structural units of the South Cottian Alps (see map Fig. 3)

Acro-nyms	Name of tectonic units	Paleogeographic domain	Main lithologies	Metamorphic facies
QN	Queyras nappe	Piemonte-Liguria Ocean	Radiolarites, marbles and calcschists with ophiolite bodies	Blueschist
MV	Monviso ophiolite complex		Mainly ophiolites with minor metasediments	Eclogite
VG	Val Grana unit	"Prepiemonte" – type facies of the Briançonnais *s.l.* distal margin	Mesozoic metasediments (Ladinian-Pliensbachian)	Blueschist
VM-S (cf. GS further N)	Val Maira – Sampeyre unit (cf. Giulian-Sea Bianca unit)		Permian-Mesozoic metasediments (Upper Permian-Ladinian)	Blueschist
CLSZ	Cima Lubin shear zone	Shear zones representing tectonic mélanges of oceanic and margin units	Mesozoic metasediments (dolostones, calcschists) and minor ophiolite bodies	Blueschist
SDSZ	San Damiano shear zone			Blueschist
VPSZ	Valmala-Piasco shear zone			HT- blueschist
Drn2	Upper Dronero unit	Dora-Maira crystalline Massif representing part of the former basement of the Briançonnais *s.l.* distal margin	Polycyclic schists and monocyclic metased. and granites (Permian)	Quartz eclogite
Drn1	Lower Dronero unit		Polycyclic basement	Blueschist
R	Ricordone unit		Polycyclic basement	Quartz eclogite
RS	Rocca Solei unit		Polycyclic basement	Quartz eclogite
BI	Brossasco-Isasca unit		Polycyclic basement	Coesite eclogite
SP	Sanfront-Pinerolo unit		Monocyclic metasediments (Carboniferous- Triassic) + Permian felsic intrusives	Garnet blueschist to eclogite
AL	Units of the Acceglio- Longet band (Acceglio anticline, Rocca Corna and Pelvo d'Elva units) and their southern equivalents	Acceglio-type facies domains of the Briançonnais *s.l.* distal margin (Acceglio, "Ecailles intermédiaires" or Ultrabriançonnais *auct.*)	Monocyclic metasediments (Up. Carbon.-Up. Cretac.-Paleoc., with Middle Triassic—Liassic gap) + Permian felsites Polycyclic schists (scarce)	HT- blueschist
CBR	Classical Briançonnais nappes or Briançonnais *s. str.* (various nappes, see Fig. 6)	Briançonnais proximal margin (= Briançonnais *s.str.* domain)	Cf. Acceglio units but with thick Middle ± Upper Triassic dolostones ± Lower Liassic (rare) + Eocene "Flysch noir"	Greenschist to LT- blueschist
SBR	Subbriançonnais units	Valaisan and Subbriançonnais rift domains	Up. Triassic (evaporites) – Eocene sediments	Non- metamorphic

The following two units overlie the coesite-eclogite facies Brossasco-Isasca unit and reached quartz-eclogite facies conditions: the lower one is labeled *Rocca Solei (RS) unit* and the next higher *Ricordone (R) unit* (equivalent to "Units II and III, respectively, in Henry et al., 1993). Both display lithological associations similar to those of the BI unit (Compagnoni & Rolfo, 2003). Thin slivers of quartzites, dolostones and calcschists are pinched along the tectonic contact between Rocca Solei and Ricordone units (Henry et al., 1993). These Mesozoic slivers connect with similar Mesozoic slivers that are part of the Valmala-Piasco shear zone.

The *Valmala-Piasco shear zone (VPSZ)* is a major mélange-type shear zone that truncates the top parts of both Ricordone and Rocca Solei units, and locally even the Brossasco-Isasca unit (Figs. 3, 6). In addition to lenses of Triassic carbonates (Piasco, Rossana) the VPSZ also displays serpentinites (Venasca-Serravalle quarry), metabasites and calcschists (Balestro et al., 2020; Michard, 1967). During the latest stages of its tectonic evolution this shear zone represented an extensional detachment that exhumed the eclogitic units in the footwall with respect to the overlying, blueschist-facies Lower Dronero unit (Drn1). The overall trace of the VPSZ in map view underlines the dome-shaped structure of the eclogitic units of the southern Dora-Maira units.

The *Dronero unit*, originally defined by Michard et al. (1993) and later tentatively divided into *lower and upper Dronero units* (Drn1 and Drn2, Fig. 3) by Balestro et al. (1995), includes polymetamorphic garnet-chloritoid schists that originate from pre-Permian basement, metagranitoids (orthogneiss, sometimes with whiteschists) and metabasites similar to those observed in the Ricordone and Rocca Solei units (Balestro et al., 1995; Henry et al., 1993), as well as ankerite-chloritoid silvery schists of presumably Permo-Carboniferous age or Lower Permian age (Balestro et al., 2020; Henry, 1990; Michard, 1967). According to Balestro et al. (1995), eclogite assemblages are only preserved in the metabasites of the upper Dronero unit whereas the lower unit only shows garnet-blueschist-facies assemblages.

A second major mélange-type shear zone, referred to as *San Damiano shear zone (SDSZ*), also involves tectonic lenses of calcschists, metabasites and continent-derived rocks (gneiss, Triassic carbonates). It is found in the hangingwall of the southwestern margin of the lower Dronero unit. This shear zone omitted the eclogitic upper Dronero unit in its footwall, suggesting that it also accommodated late-stage normal faulting. It closely resembles the Valmala-Piasco shear zone in terms of its mixed lithological content.

The *Val Maira-Sampeyre (VM-S) unit* is the lower tectonic unit amongst the Maira-Sampeyre and Val Grana Allochthons. It is found in the hangingwall of the San Damiano shear zone. The base of the VM-S unit, namely the "Sampeyre unit" as defined by Michard (1967) can be followed all the way from Val Maira to Val Varaita. Note that we interpret the Sampeyre unit sensu Michard (1967) no longer as a part of the Dora-Maira units. Rather it represents a lower subunit (Sampeyre subunit) of the larger Val Maira-Sampeyre unit. This lower subunit is made up of conglomeratic quartzites (Permo-Triassic) and fine-grained quartzites (Lower Triassic). It is separated from the overlying Anisian-Ladinian carbonates of the Val Maira subunit (corresponding to "Unité I" in Michard, 1967) by a folded fault along which the lithologies of the upper Val Maira subunit are wedged out in map and profile view (see Figs. 3, 6). In Fig. 4 we propose that the folded fault between the two subunits possibly represents a reactivated paleo-fault. The stratigraphic column of the Val Maira-Sampeyre unit is presented for convenience as Additional file 1: Fig. S1A.

The *Cima Lubin shear zone (CLSZ)*, originally labeled "Unité II" in Michard (1967), tectonically separates the Val Maira-Sampeyre unit from the overlying Val Grana unit described below. It represents a third mélange-type ophiolite-bearing shear zone. This lithologically heterogeneous shear zone involves tectonic lenses of silvery mica-schists (Permian?), quartzites (Lower Triassic), cargneules and dolostones (Middle-Upper Triassic), as well as calcschists and metabasites. To the north, the CLSZ continues into the so-called "Roccenie formation" (Mondino, 2005), a metasedimentary sequence that overlies the Giulian-Sea Bianca Unit (GS, Fig. 3) and directly underlies the basal serpentinites of the Monviso ophiolitic unit (Agard et al., 2001; Angiboust et al., 2012; Castelli et al., 2014; Lombardo et al., 1978; Schwartz et al., 2000b). The overall steep westward dip of the CLSZ in the north, from the Po-Varaita crest to the Varaita-Maira crest, gives way further to the south and across the Maira valley to a shallow, southwestward to southward dip locally deformed by kilometer-size late-stage folds. These are: (i) an antiform north of the Stroppo village in Val Maira (Fig. 3); (ii) the Pradleves antiform in Val Grana (Fig. 5), and (iii) the Valgrana antiform (next to the locality Valgrana in lower Val Grana) exposing a tectonic window of (Val Maira) dolostones below the CLSZ mélange and Val Grana dolostones (Fig. 3). These late-stage folds formed during the back-folding and—thrusting that took place during phase D3 discussed later.

The *Val Grana (VG) unit* exposes Triassic carbonates that predominate in the east, i.e., in the lower Grana valley, whereas the Jurassic strata dominate northwest of the Colletto transverse fault, which operated as a normal wrench (tear) fault during late-stage folding of the unit (Fig. 3). The carbonate sequence of the Val Grana

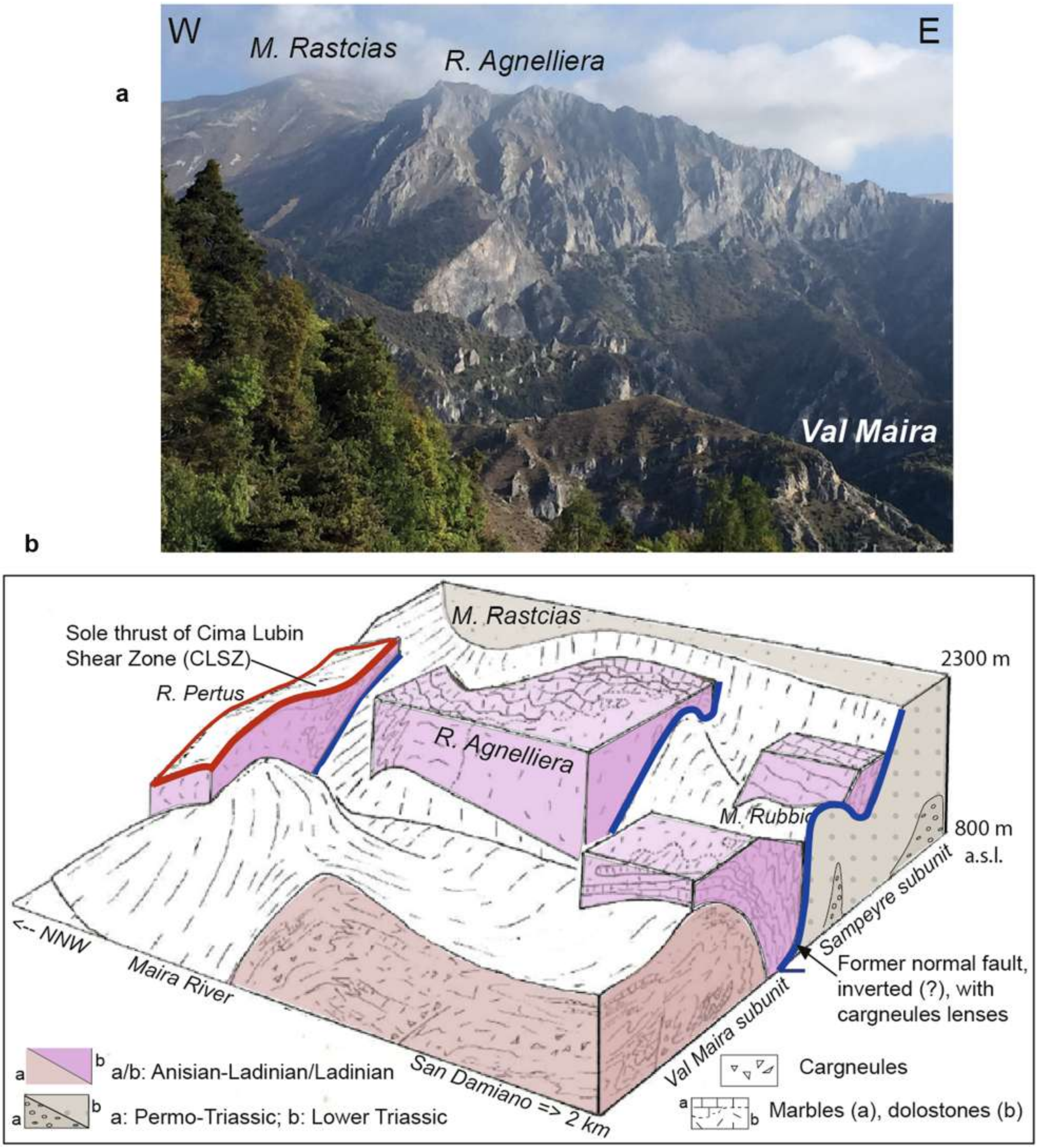

Fig. 4 Panorama (**a**) and block-diagram (**b**) showing the relationships between the meta-carbonates of the Val Maira subunit and the underlying meta-clastic, Permo-Triassic and Lower Triassic Sampeyre subunit. The folded contact between the two subunits may be interpreted as an inverted former normal fault within the Briançonnais margin. Note the occurrence of dolomitization fronts in the Ladinian formations and the abundance of cargneules in the Anisian-Ladinian

unit includes Ladinian dolostones and subordinate sandy limestones, black cherts and meta-cinerites, followed upward by Carnian-Norian dolostones. The transition to the Jurassic beds is marked by Rhaetian-Hettangian coral and cherty limestones. These are overlain by the meta-calcarenites with bedded to chaotic breccias whose

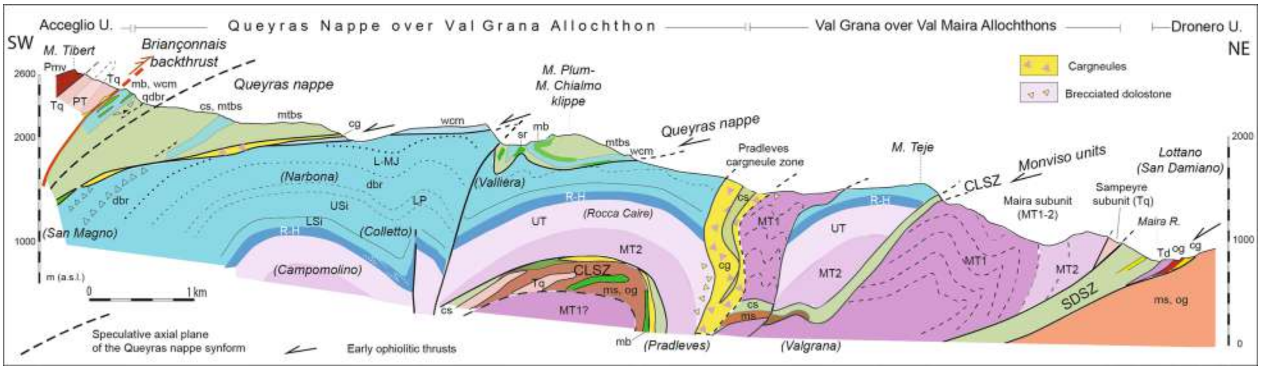

Fig. 5 Semi-schematic geological cross-section of the Maira-Sampeyre and Grana Allochthons and adjoining units. See Fig. 3 for location of the topographic line; the deepest parts of the section are projected from the SE over 1–5 km, complying with the thicknesses of the geological formations. The entire section has been affected by blueschist-facies metamorphism before its exhumation under greenschist-facies conditions. *Stratigraphic abbreviations*: L-MJ: Lower-Middle Jurassic coarse/chaotic carbonate breccias; LP: Lower Pliensbachian cherty marbles; LSi: Lower Sinemurian siliceous marbles; MT: Middle Triassic, with 1: Anisian-Ladinian evaporitic carbonates, and 2: Ladinian dolostones; Pmv: Permian meta-volcanics; PT: Permian–Triassic conglomerates (Verrucano Piemontese, "anagenites"); R-H: Rhaetian-Hettangian coral marbles; Tq: Lower Triassic (Werfenian) quartzites; USi: Upper Sinemurian turbiditic carbonate breccias and marbles; UT: Upper Triassic dolostones;. *Lithostratigraphic abbreviations*: cg: cargneules (with dolostone/marbles blocks, particularly in the Pradleves cargneule zone); cs: calcschists; dbr: dolostone breccias; mb: metabasite; ms: mica-schist; mtbs: meta turbiditic black shales; og: orthogneiss; qdbr: quartzite and dolostone breccias (Longet breccias, LB in Fig. 3); sr: serpentinite; wcm: whitish cherty marbles

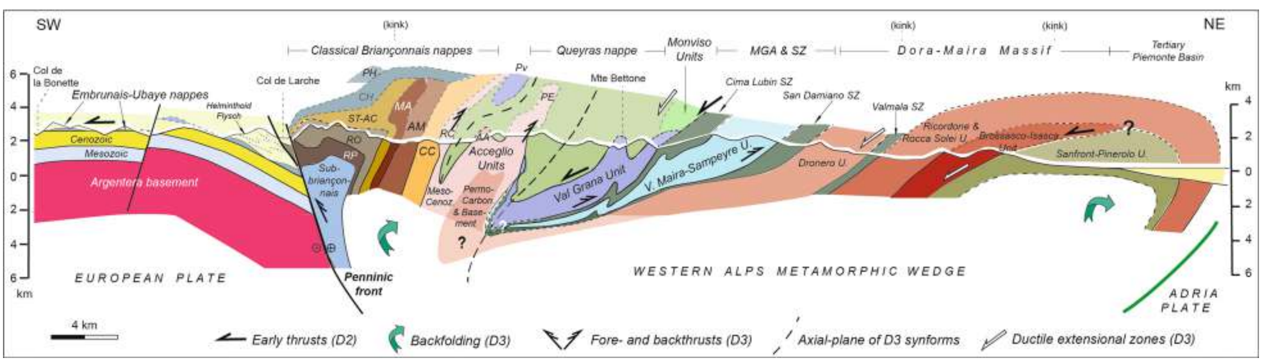

Fig. 6 Orogen-scale profile across the Cottian Alps. See Fig. 3 for the trace of the profile, and text in this section and in the Discussion (Sect. 5). AA: Acceglio antiform; AC: Aiguille de Chambeyron unit; AM: Aiguilles de Mary unit; CC: Ceillac-Chiappera unit; CH: Châtelet unit; MA: Marinet unit; MGA: Maira-Sampeyre and Grana Allochthons; PE: Pelvo d'Elva antiform; PH: Peyre Haute unit; Pv: Péouvou « Prepiemonte» unit; RC: Roure-Combrémond units; RO: Rouchouze unit; RP: Rocca Peroni unit; ST: Sautron unit; SZ: shear zone with tectonic mélange

base is Sinemurian-Lower Pliensbachian (Narbona valley, Fig. 8c; Sturani, 1961; Ellenberger et al., 1964; Michard, 1967; see also Additional file 1: Fig. S1B). The Norian dolostones reappear in Val Maira where they form the core of the kilometer-scale, east-verging Monte Bettone anticline (Figs. 3, 8e, f) formed during D3 back-folding and -thrusting as well. There, an angular unconformity occurs between the tilted Norian beds and overlying Rhaetian-Hettangian formation (Additional file 1: Fig. S3).

The Maira-Sampeyre and Val Grana Allochthons are obliquely overlain by two lithologically different tectonic units derived from the Piemonte-Liguria Ocean: (i) the *Monviso (MV) ophiolitic complex*, consisting of eclogitic units that dominantly expose metagabbros and metabasites with subordinate oceanic metasediments (breccias, marbles, calcschists), underlain by a thick basal sole of serpentinites (Lombardo et al., 1978; Deville et al., 1992; Schwartz et al., 2000b, 2013; Agard et al., 2001; Angiboust et al., 2012; Castelli et al., 2014; Angiboust and Glodny, 2020), and, (ii) the ophiolite-bearing *Queyras nappe (QN),* consisting of three units (not separated in Fig. 3) of blueschist-facies oceanic metasediments hosting boudinaged kilometric meta-ophiolite bodies (Lagabrielle and

Polino, 1988; Deville et al., 1992; Schwartz et al., 2000a, b; Tricart and Schwartz, 2006; Herviou et al., 2021). Between the Upper Po and Varaita valleys (Figs. 3, 6), the Queyras nappe tectonically overlies the Monviso complex along a west-dipping, late-stage extensional fault (Ballèvre et al., 1990; Schwartz et al., 2007).

In turn, the Monviso complex overlies the Cima Lubin shear zone and the Val Maira-Sampeyre unit or its northernmost equivalent, the *Giulian-Sea Bianca (GS) unit* (Balestro et al., 2011; Michard, 1967) along another west-dipping late extensional fault that can be traced southward into the hangingwall of the Cima Lubin shear zone (Fig. 3). However, immediately south of the Varaita-Maira watershed (Sampeyre Pass), the Monviso serpentinites disappear in map view, mainly due to the mega-lens geometry of the whole Monviso complex, and additionally to a minor north-verging normal fault. Only rare potential remnants of the thick Monviso complex are found further south as lenses in the Cima Lubin shear zone. Thus, south of the Sampeyre Pass and down to the alluvial plain of Cuneo, it is the Queyras nappe that directly overlies the Val Grana unit (see profile of Fig. 6 around and west of Monte Bettone, and discussion, Sect. 5). The metasedimentary series of the Queyras nappe include Upper Jurassic meta-radiolarites and Tithonian-Berriasian marbles, Lower Cretaceous metamorphosed turbiditic black shales and monotonous calcschists dated as Upper Cretaceous (Tricart and Schwartz, 2006; Herviou et al., 2021 and references therein).

The Val Grana unit and overlying Piemonte-Liguria units are bounded to the west and SW by the backthrusted *Acceglio-Longet (AL) antiformal band* derived from the Acceglio s.l. paleogeographic domain (Lefèvre and Michard, 1976; Schwartz et al., 2000a). This band of dominantly siliceous rocks forms the hangingwall of the major D3 synform depicted in Fig. 6. Another major backthrust located at the margin of the studied area, west of Acceglio locality, brings the Roure and Combrémond units, with their sedimentary cover, similar to the Acceglio-Longet units (Le Guernic, 1967; Michard et al., 2004), over a second synform occupied by the Queyras rocks (Fig. 6). The tight anticlines (Acceglio and Pelvo d'Elva) and associated slivers that form the Acceglio-Longet band, and the two Roure and Combrémond units represent previously stacked slivers derived from one and the same Acceglio *s.l.* paleogeographic domain (see Discussion, Sect. 5.2.4).

Upper Cretaceous-Paleocene polygenic chaotic breccias (Longet-type breccias; LB in Fig. 3) associated with light gray and occasionally siliceous marbles appear in the tectonic contact with the Acceglio *s.l.* units (Pelvo d'Elva inverted limb and Longet area). These puzzling, chaotic breccias occur as discontinuous tectonic lenses at the very bottom of the Queyras Schistes Lustrés, in terms of the original nappe stack that formed during D2, i.e., before D3 (Leblanc, 1962; Lemoine, 1967; Lefèvre and Michard, 1976; Gout, 1987). Some of these breccia lenses could have been detached from their Acceglio *s.l.* substrate during thrusting of the Queyras nappe (see Discussion, Sect. 5.2.4).

3.3 Setting of the Maira-Grana Allochthons at the scale of the orogen

In order to put the tectonic units described above into an orogen-scale context we constructed the profile presented in Fig. 6. Its trace indicated in Fig. 3 extends further to the SW in direction of azimuth 210° towards Col de Larche and ends at Col de la Bonette.

The westernmost part of the profile shows the Penninic Front, i.e. a major fore-thrust formed during D3 that also thrusts the previously emplaced Embrunais-Ubayae nappe stack. This part of the profile is a slightly modified version of the profile published on the 1:50,000 map, sheet Larche (Gidon et al., 1977), complemented with along-strike projections of the Subbriançonnais units exposed in Valle Stura di Demonte, based on the 1:100,000 geological map sheet Argentera (Malaroda et al., 1971). Lateral projections are based on the pronounced NW oriented axial plunge of the edge of the Argentera external massif that also affects the neighboring Briançonnais units. The slope of the Penninic Front follows the constraints provided by crustal P-wave tomography provided by Diehl et al. (2009) and Schmid et al., (2017, their Fig. 2a).

The part of the section between Col de Larche and Acceglio was constructed by using the French 1:50,000 map, sheets Larche and Aiguille de Chambeyron (Gidon et al., 1977 and 1994, respectively). The along-strike SE-ward projection of a section exposed along the Upper Ubaye valley into the section of Fig. 6 is based on an axial plunge of 11° towards azimuth 313° across the Briançonnais s.str. units, the Roure-Combrémond Acceglio *s.l.* unit and the Péouvou Prepiemonte unit. The projection of the Acceglio-Longet anticlines, however, used an axial plunge of 9° towards azimuth 340°. Both these values correspond to the average plunge of the D3 retrofolds extracted from pole figure data provided by Caron et al. (1973), Lefèvre and Michard (1976) and Michard et al. (2004) that apply to these two sectors of the profile. In this part of the profile the Briançonnais nappe pile became tilted into a steeply foreland-dipping orientation due to D3 backfolding.

The Roure-Combrémond and Acceglio-Longet units are interpreted as being connected at depth around a

F3 synform with the Queyras Piemonte-Liguria units in the core, following Michard et al., (2004; their Fig. 4). We refer to this synform as the *external Queyras synform*. However, the exact geometry of the structures around this synform at depth are unknown and largely conceptual.

The link of the Acceglio *s.l.* units with the area exposing the MGA units buried at depth is provided by the *internal Queyras synform*. According to our interpretation this synform separates the steeply SW-dipping units of the various Briançonnais s.l. units described above from the more flat-lying part of the profile located NE of the synform, which exposes a normal way up nappe stack. This interpretation is firmly supported by the vergence of large-scale F3 folds affecting the MGA units discussed earlier (see detailed profile of Fig. 5) and the vergence of the spectacular Mte Bettone backfold crossed by the profile of Fig. 6 (see also Fig. 8e) and located NE of the axial plane of this mega-fold.

For constructing the part of the section NE of the internal Queyras synform, located outside map sheets Larche and Aiguille de Chambeyron, we used the detailed map of Michard (1967) covering the Maira-Sampeyre and Val Grana Allochthons and adjacent areas, and the map compilation presented in Fig. 3 modified after Piana et al. (2017) that also covers the Dora-Maira units. The westernmost part of the section shows that the Maira-Sampeyre and Val Grana Allochthons, together with the underlying units constituting the Dora-Maira units, form a normal way-up original nappe stack located in the lower limb of the internal Queyras synform. The units of the Dora-Maira massif define an asymmetric antiform related to a hypothetical backfold, partly concealed by the Plio-Pleistocene deposits of the Tertiary Piemonte basin (Cassano et al., 1986; Pieri & Groppi, 1981). The existence of such a backfold is a corollary of the necessity to root the Piemonte-Liguria oceanic units along the W-dipping interface between the top of Ivrea Zone of the Adria plate and the base of the Dora-Maira units (Debelmas et al., 1983; Schmid et al., 2017).

3.4 Meso- and microstructures

The MGA and neighboring units have all been affected by (U)HP-LT metamorphism, ranging from the coesite-eclogite-facies in the Brossasco-Isasca unit of the southern Dora-Maira "massif" to blueschist-facies in the Queyras and Acceglio *s.l.* units (Goffé et al., 2004; Bousquet et al., 2008; see Sect. 4 for details). Therefore, their lithologies exhibit the usual, polyphase structure of rocks that were subducted, then exhumed, and eventually affected by the latest collisional events. Figure 7 provides a brief overview of the orientation of mesostructures that have been analyzed in terms of superimposed phases of deformation by previous workers in the areas of the southern Dora-Maira units (Henry et al., 1993), the Val Grana and Val Maira transects (Horrenberger & Michard, 1978; Schumacher, 1972), and in the Acceglio-Longet antiformal band (Lefèvre and Michard, 1976; Michard et al., 2004).

The stereoplots from the Dora-Maira basement rocks (Fig. 7a, b) show a weakly dispersed flat-lying foliation that overprints earlier high-P parageneses and formed during top-to-SW shearing occurring under retrograde greenschist-facies conditions, associated with the exhumation of the Dora-Maira units along low-angle extensional faults (Avigad et al., 2003; Henry et al., 1993). Cataclasites and pseudotachylites are locally hosted in the top-to-the SW/W mylonitic foliation (Dana, 2020; Ferré et al., 2015; Henry et al., 1993; Zechmeister et al., 2007). According to Rubatto and Hermann (2001), greenschist-facies conditions in the Dora-Maira units prevailed during the 33–30 Ma time interval. This time interval approximately coincides with the timing of D3 back-folding and—thrusting in the Val d'Aosta sections (35–31 Ma according to Bucher et al., 2004) as well as with the age of fore-thrusting at the Penninic Front behind the Pelvoux Massif (34–31 Ma according to radiometric dating by Simon-Labric et al., 2009). Although the structural relation of such normal faulting within the Dora-Maira nappe stack with D3 backthrusting affecting the adjacent Val Grana and Val Maira units remains unclear we assume this extensional exhumation to be part of the D3 deformation phase.

The authors of Fig. 7c, d presented stereoplots covering Mesozoic units. They all described a first schistosity mostly parallel to bedding (S0-1), overprinted by an S2 foliation that is axial-planar to isoclinal D2 folds. In general, S1 and S2 are hard to be distinguished in the absence of F2 folds. Hence, F1 and F2 can be confused in many cases. Likewise, fold axes related to D2 and D3 are often hard to separate. However, S3 foliations related to backfolding can often easily be discriminated from earlier structures. A penetrative S3 foliation is observed in incompetent lithologies that underwent intense shearing during D3 leading to the progressive rotation of fold axes into parallelism with the direction of shearing within S3. This is manifested by the distribution of F3 fold axes within a great circle (Fig. 7e, f, h). The directions of transport were derived according to a method described by Caron et al. (1973) that utilizes the great circle distribution of pre-D3 lineations (not shown in Fig. 7e, f, h). This clearly documents the intensity of shearing in incompetent lithologies, i.e., within the micaceous Permian–Triassic beds of the Acceglio-Longet Band (Fig. 7h) and in the underlying Schistes lustrés of the Queyras nappe (Fig. 7e, f). The large-wavelength, NW-trending D3

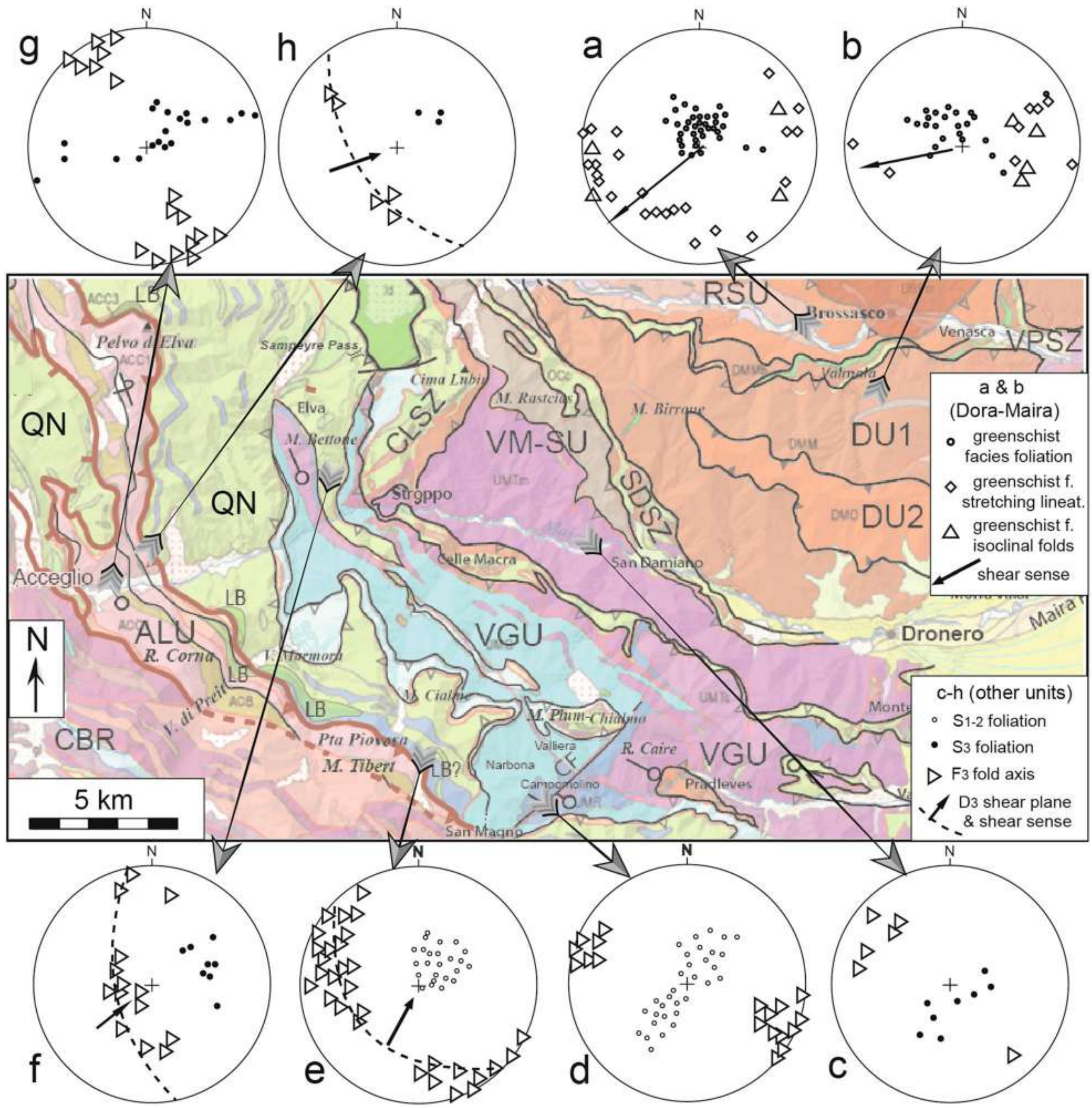

Fig. 7 Orientation of the structural elements in the studied area. Stereoplots (Wulff net, lower hemisphere) simplified after Henry et al., 1993 (**a**, **b**), Horrenberger & Michard, 1978 (**c**, **f**), Schumacher, 1972 (**d**, **e**) and Lefèvre and Michard, 1976 (**g**, **h**). In **g**, the black dots mainly correspond to S_3 fanned cleavage, but possibly also to the folded S_2 foliation

anticlines formed by stiff slabs of quartzite or dolostone are illustrated in the Acceglio (Fig. 7g) and Narbona-Campomolino (Fig. 7d) diagrams, respectively. The effect of this D3 folding is less visible in the diagram of Fig. 7c from the Val Maira Anisian-Ladinian weak formations derived from an evaporitic sequence.

Figure 8 illustrates typical structures observed in the Maira-Sampeyre and Val Grana Allochthons at the outcrop and landscape scale. The occurrence of superimposed structures is particularly clear in the case of the Lower-Middle Jurassic beds near Colletto, 1 km NW of Campolino (Fig. 8a, c, d). At the larger scale, competent horizons appear to be only affected by late stage D3 open folds such as in case of the slab of Triassic dolostones (Rocca Caire and Campomolino anticlines; Fig. 8c).

However, at the meter scale (Fig. 8a, d) as well as under the microscope (Fig. 9a, b) the structures exhibit superimposed foliation and crenulation planes that reveal multi-stage deformation. The dominant (regional) foliation overprints a previous foliation, which is sheared obliquely or microfolded within microlithons bounded by the younger (S3) foliation (Fig. 8a, b). In line with Schumacher (1972) and Caron et al. (1973), we label these foliations S2 and S3, respectively, assuming that a previous S1 foliation formed sub-parallel or parallel to bedding S0. In contrast, the dolostone slabs of the same unit generally show scattered, strait phengite lamellae devoid of preferred orientation at the microscopic scale. The development of a tectonic cleavage in the dolostone beds has been only observed in the Mte Bettone tight anticline (Fig. 8e, f). Detachment of the incompetent Liassic formations above their competent Triassic base during the deformation process is documented by decametric recumbent folds located in the Rhaetian beds downstream Campomolino (Schumacher, 1972).

In the Anisian-Ladinian formations of the Val Maira sub-unit, the early metamorphic foliation S1-2, which nearly parallels bedding, is severely deformed by recumbent folds and minor thrust faults (Fig. 8h). The protoliths of this formation are dolomitic limestones interbedded with evaporitic clays (Michard, 1967), which acted as a detachment horizon between the Val Maira Middle Triassic formations and their Lower Triassic base (Sampeyre subunit) notwithstanding the role of an early normal fault (Fig. 4). These lithologies are likely at the origin of the 400 m-thick cargneule zone of Pradleves (Figs. 3, 5).

In the Queyras Nappe, whose metamorphic grade compares with that of the underlying Val Grana unit (see chapter 4), the meso- and microstructures observed in the Upper Jurassic-Cretaceous Schistes Lustrés also compare with those in the underlying Early-Middle Jurassic sequences. Tight F2 folds folding S0/S1 are well illustrated in the metacherts of the Mte Plum klippe (Fig. 8g); their axial plane is deformed by open F3 folds at a large scale (see cross-section, Fig. 5). The planar structure of banded competent glaucophanite (see microscopic features in Fig. 9g) is probably defining a less deformed S2 foliation. In contrast, the S3 foliation is particularly penetrative in the Ussolo calcschists sampled at short distance of the Acceglio-Longet backthrust (Figs. 8b, 9b). In the Ussolo outcrops, ankeritic calcschists display pseudomorphs of lawsonite prisms ("lawsonite A"-type of Lefeuvre et al., 2020), which contain relics of the pre-S3 foliation (Fig. 9e).

On the western and southwestern border of the Maira-Sampeyre and Val Grana Allochthons and west of the main (internal) D3 synform, the Acceglio-Longet antiformal band (Fig. 3) also offers superimposed microstructures involving a poorly preserved S1 foliation, a well-marked S2 foliation locally associated with isoclinal folds, and a dominant S3 crenulation-type foliation (Schumacher, 1972; Lefèvre and Michard, 1976; Michard et al., 2004). The jadeite orthogneiss at the bottom of the overturned Pelvo d'Elva unit (see Fig. 10 for location) formed at the expense of a Permian (?) alkaline granite (Lefèvre and Michard, 1965). It now displays a typical mylonitic S/C' structure (Fig. 9h). This D3 microstructure developed during the retrograde alteration of jadeite into muscovite and albite and the concomitant quartz recrystallization in the pressure shadows (see also the macroscopic view of the same sample in Additional file 2: Fig. S1).

The microstructures in the southern Dora-Maira eclogite- and blueschist-facies schists are comprehensively described in Henry et al. (1993) and Avigad et al. (2003). There, the Alpine nappe contacts are overprinted by mylonitic microstructures, which developed under retrograde greenschist-facies conditions. Foliation and C' shear bands all indicate ductile top-SW extensional overprint (Fig. 7). As mentioned earlier, the main greenschist facies foliation "Sm" described by Henry et al., (1993; their Fig. 7) indicating extensional unroofing developed roughly at the same time as the

(See figure on next page.)

Fig. 8 Outcrop and landscape scale structures. **a** Penetrative S2 foliation with flattened quartz lenses crenulated by late foliation S3; locally, minor shear planes C oblique on S3 indicates top-NE shear sense; Battuira (44.4274 × 7.2128), Val Grana Upper Liassic-Dogger(?) turbiditic limestones. **b** Tight crenulation schistosity (S3) entirely transposing bedding and earlier foliation (S2) in the Queyras calcschists beneath the Pelvo d'Elva backthrust; Ussolo (44.4910 × 7.0260). **c** Campomolino open anticline (F3) seen from the crest north of Punta Castellar (44.4067 × 7.1837); Val Grana unit, Sinemurian bedded limestones and breccias topped by the Pliensbachian Bercia limestones. **d** Close-up of carbonate breccias to the SW of (**c**); F3 folds (deforming a thin breccia bed and minor calcite veins) in foliated marbles showing lenticular texture and tilted, flattened dolostone pebbles; Val Grana Middle-Upper Liassic member west of Punta Castellar above San Magno (44.4041 × 7.1771). **e** Western (normal) limb of the Monte Bettone F3 anticline from the Vallone d'Elva (hazardous!) road; a minor parasitic anticlinal fold is seen below the major hinge; Norian bedded dolostones of Val Grana unit. **f** Detail of an oblique cleavage (S3) that exceptionally affects the dolostone beds, slightly oblique (N130E) with respect to the fold axis in map view (N160E); lowest Vallone d'Elva (44.4980 × 7.1010). **g** Tight similar folds (F2) deforming S0-1 with tilted axial plane due to later major folding (F3, not shown) in Upper Jurassic meta-cherts of the Queyras klippe, eastern cliffs of Monte Plum (44.4297 × 7.2303). **h** Folding of an early metamorphic foliation by superimposed recumbent F3 (?) folds; Anisian-Ladinian dolomitic limestones and argillites associated with cargneules west of Lottulo, Val Maira subunit (44.4940 × 7.2113)

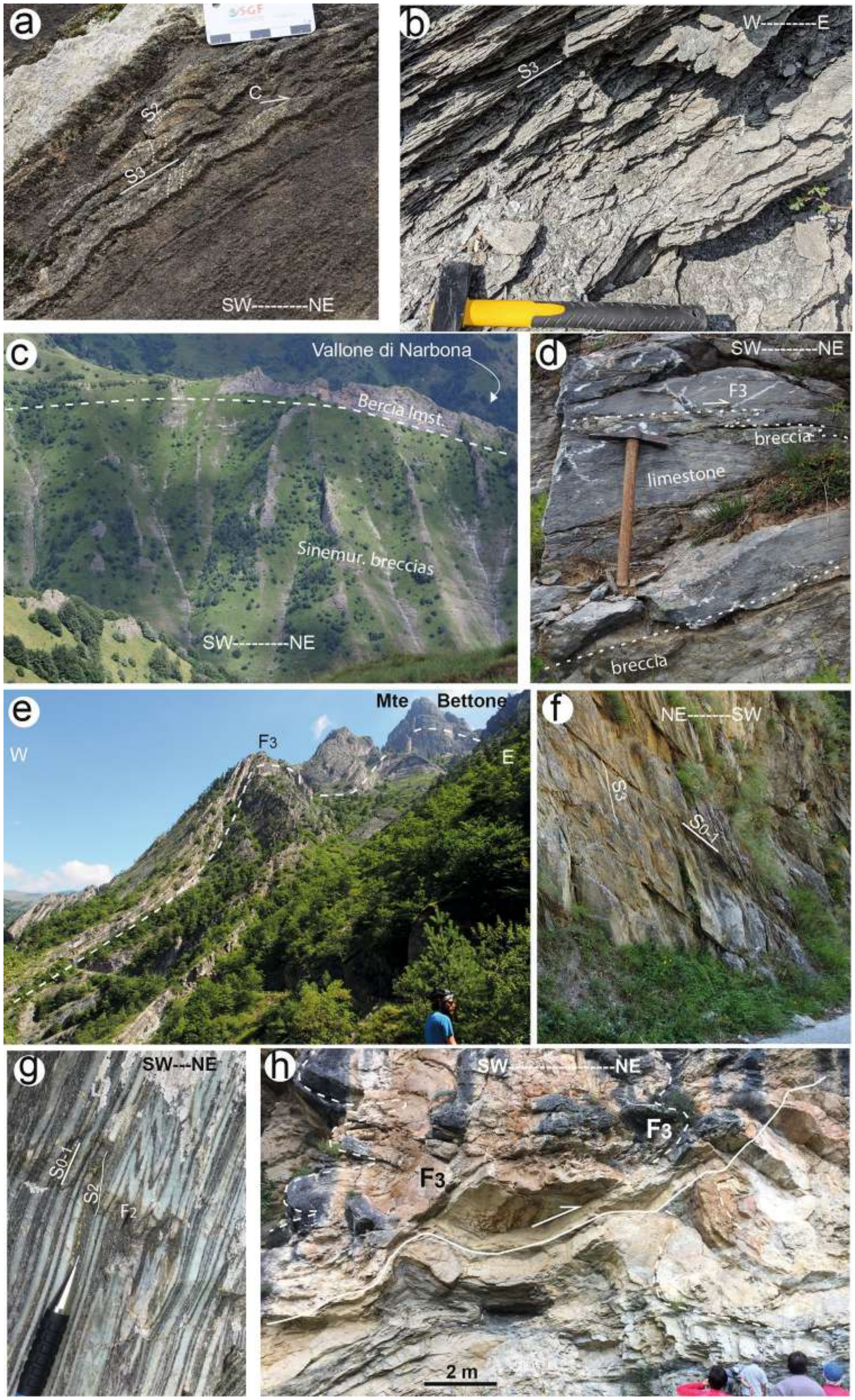

Fig. 8 (See legend on previous page.)

S3 foliation related to the D3 back-folding event in the overlying Maira-Sampeyre and Val Grana Allochthons, although the kinematics are different (top-SW normal faulting in the Dora-Maira units vs. top-NE thrusting in the other units).

4 Metamorphic mineralogy and Raman geothermometry

4.1 Mineral composition of metamorphic rocks

4.1.1 Summary of published results

Blue amphibole ("gastaldite"), lawsonite and jadeite have been described by Franch (1898, 1900) in the Acceglio-Longet antiformal band and the bordering ophiolitic Schistes lustrés in Val Grana (Rio Grande and Mte Ruera; Fig. 10), as well as chloritoid ("sismondine") in the Schistes lustrés between Elva and Stroppo (Val Maira). Lefèvre and Michard (1965) described the partly retrogressed minerals jadeite, glaucophane and lawsonite in the Acceglio-Longet band (see Additional file 2: Figs. S1, S2). Michard (1967) found the mineral assemblages ("parageneses") quartz-phengite-lawsonite ± chlorite ± chloritoid in the MGA, and glaucophane-lawsonite-chlorite in the juxtaposed Schistes lustrés metabasites. He pointed out that these HP-LT minerals become progressively altered to higher-T, lower-P minerals when approaching the Dora-Maira units. For example, lawsonite is well preserved in the metabasites of the Acceglio-Longet band, next to jadeite-bearing orthogneiss (Lefèvre and Michard, 1976). However, it is altered into epidote, albite, white mica ± chlorite in the metatuffites interlayered in the Norian dolostones of the Pradleves anticline (Val Grana; see Additional file 2: Fig. S3) and in the Ladinian schists of the Val Maira sub-unit (Fig. 9c).

In the recent literature, the MGA and overlying Schistes Lustrés of the Queyras Nappe were altogether considered to belong to the blueschist-facies zone of the Western Alps, with peak metamorphic conditions of ~ 300–400 °C, 8–15 kbar (Goffé et al., 2004; Oberhänsli et al., 2004; Bousquet et al., 2008, 2012). The metamorphic facies of the Acceglio-Longet antiformal band was classified as upper blueschist-facies (transitional to eclogite-facies), with peak metamorphism at about 450 °C, 12–14 kbar (Goffé et al., 2004; Michard et al., 2004; Schwartz et al., 2000a). Goffé (2002) and Goffé et al. (2004) gave a precise description of the retrograde evolution towards the greenschist-facies based on Fe–Mg carpholite (Fe, Mg)$^{2+}$$Al_2Si_2O_6(OH)_4$ in the metapelitic rocks. Fresh carpholite fibers occur in the innermost units of the classical Briançonnais (Ceillac-Chiappera unit; see also Michard et al., 2004), relic fibers in most of the Queyras Schistes lustrés, and entirely transformed fibers in the easternmost Queyras units, next to the Monviso units (Fig. 10).

In the Colle di Sampeyre area, Mondino (2005) described scattered occurrences of omphacite-garnet and glaucophane-lawsonite associations in metabasites found within both the lowest Queyras Schistes lustrés and the Cima Lubin shear zone. In other words, metamorphic grade reaches the blueschist-eclogite transitional facies around the southern tip of the Monviso serpentinites. Further to the north, and outside of our study area, the P–T conditions in the Queyras Schistes lustrés have been shown to evolve from *ca.* 12 kbar, 330 °C (low-T blueschist) in the western (uppermost) unit towards *ca.* 18–20 kbar, 450–470 °C (blueschist-eclogite facies transition) found in the most internal (lowest) unit, which overlies the Monviso eclogitic complex (Agard et al., 2001; Herviou et al., 2021; Lefeuvre et al., 2020; Schwartz et al., 2013). Quartz-eclogite to coesite-eclogite facies conditions of metamorphism are well-known in the Dora-Maira units beneath the MGA, as reported above (chapter 3.2). Greenschist-facies retrogression also affected these HP(UHP) eclogitic units (e.g.,Avigad et al., 2003; Castelli et al., 2014; Henry et al., 1993).

4.1.2 New results

We collected 49 samples in the Varaita, Grana and Maira valleys, among which 42 have been studied (Table 2) from the point of view of their mineralogy/petrology, and 29 for Raman geothermometry. Petrological studies were carried out in two laboratories: (i) Ecole Normale Supérieure, Paris, for samples AC1-40 (optical microscopy) and 41 (electron microscopy); and (ii) Geosciences Rennes for samples GR20 85-91 collected in the Mte Plum klippe. RSCM measurements were performed at the French Geological Survey, Orléans.

(See figure on next page.)

Fig. 9 Microphotographs of key structures. Abbreviations for mineral according to Whitney and Evans (2010). **a** Superimposed foliations S2 and S3 in the calcschists/black schists of Battuira, Val Grana unit (sample AC01; same outcrop as Fig. 8a). **b** Pressure solution foliation (S3) bounding microlithons with crenulated previous S2 foliation in Liassic calcschists at Ponte Marmora, Val Grana unit (sample 38-M2111). **c** White mica (wm)-epidote-albite pseudomorph after lawsonite (?) in Middle Triassic carbonaceous schists (sample 35-M2108, Val Maira subunit). **d** Chloritoid-phengite association in quartz-mica-schist from the Cima Lubin tectonic mélange in the core of Pradleves anticline (sample AC09M). **e** Pseudomorph after lawsonite (?) including traces of a previous foliation (S2 ?), now transposed into S3 (sample 40-M2113, Queyras Schistes lustrés above Ussolo; cf. Fig. 8b). **f** Metachert with blue amphibole (glaucophane) overgrown by green amphibole (actinolite), Mte Plum klippe (sample GR20-89; cf. Fig. 8g). **g** Metabasite from the same klippe as (**f**), sample GR20-87. **h**: Jadeite-bearing mylonitic orthogneiss from the Acceglio-Longet band west of Pelvo d'Elva (sample 41 = 73Bel13; location Fig. 10); see Additional file 2, Figs. S1, S2 for a macroscopic view of the S/C′ mylonitic structure and more petrological details

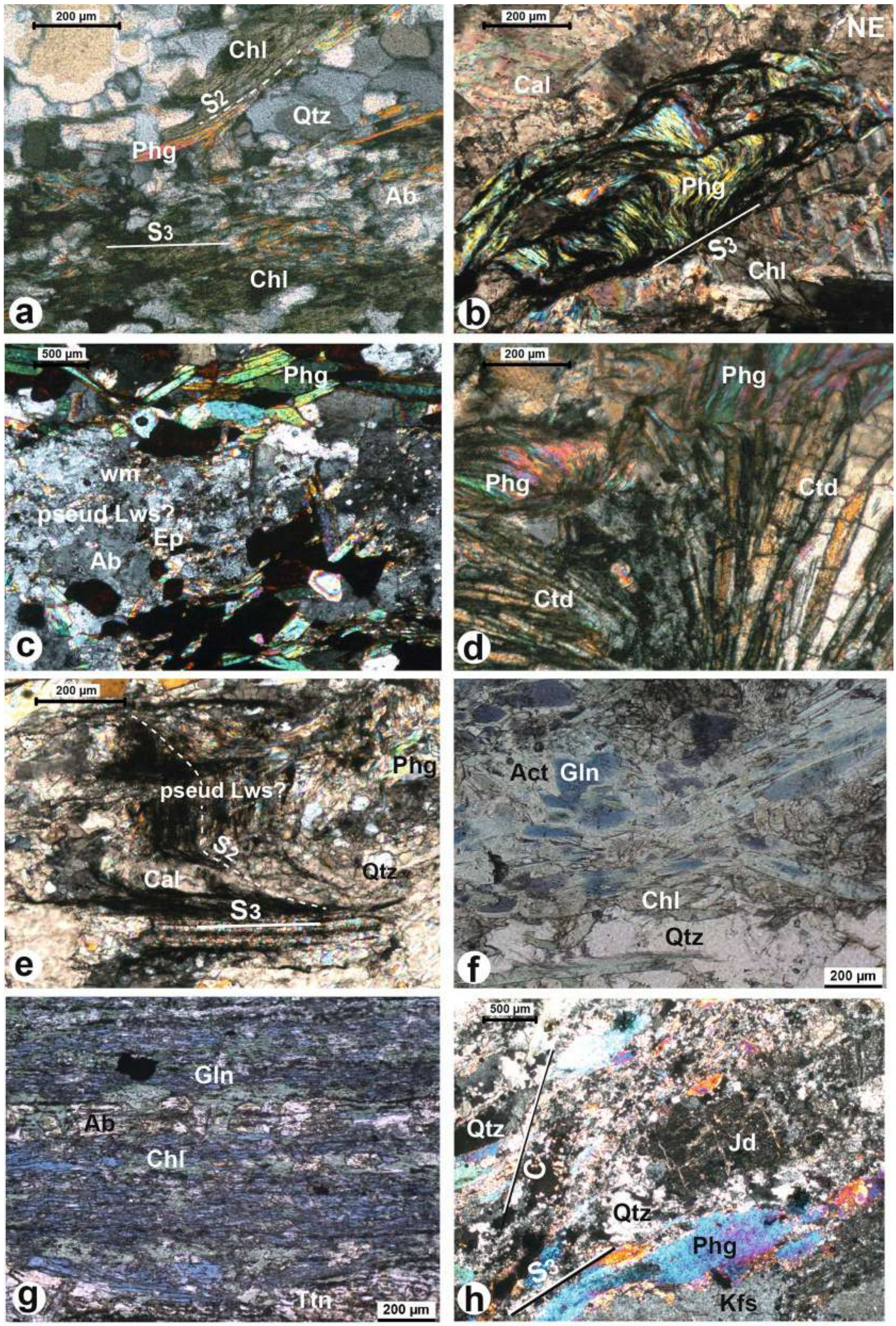

Fig. 9 (See legend on previous page.)

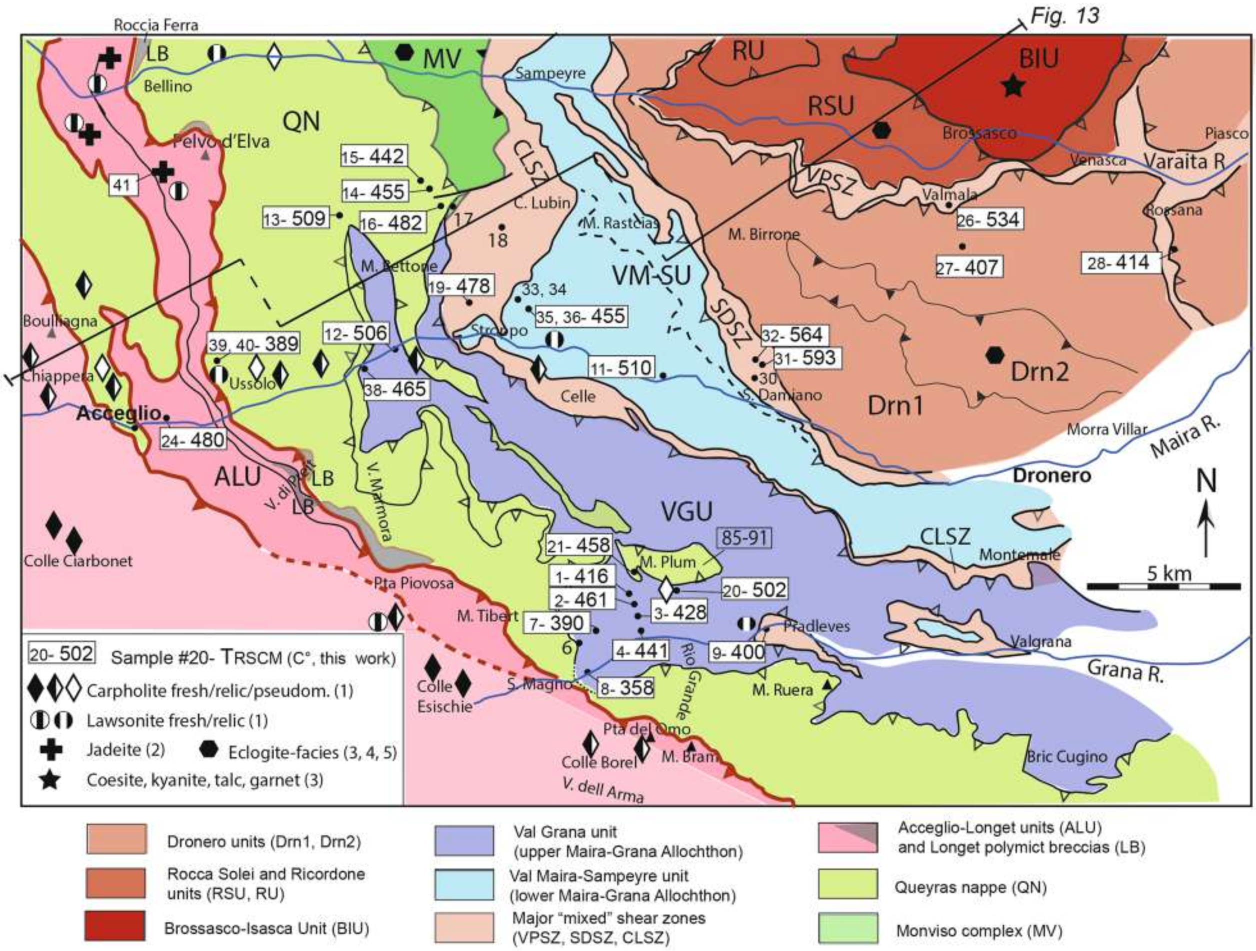

Fig. 10 Metamorphic mineral distribution from the literature and RSCM geothermometry of the samples studied in this work. See Fig. 3 for location, and Table 2 for details. Samples AC01 to AC27 are shown as 1 to 27 for clarity. The T_{RSCM} value plotted at Ussolo corresponds to the mean results obtained from samples 39, 40a and 40b.- 1: Goffé (2002); 2: Lefèvre and Michard (1976), Schwartz (2000); 3: Henry et al. (1993), Castelli et al. (2014); 4: Balestro et al. (1995); 5: Angiboust et al. (2012), Schwartz et al. (2013)

While the main mineralogical/textural characters of typical samples from the MGA were reported in chapter 3.4 (Fig. 9a–d, h), an additional study of the jadeite orthogneiss from the basement of the Pelvo d'Elva unit is presented in Additional file 2. This rock was compared (Lefèvre and Michard, 1976) to the Sapey Gneiss of Vanoise whose age was first regarded as possibly Permian but subsequently shown to be Upper Ordovician (452 ± 5 Ma; Bertrand et al., 2000).

Here we report details regarding the mineralogy and texture of the metamorphic rocks that form the Mte Plum klippe of the Queyras Nappe. We paid particular attention to this tectonic element as its significance has been debated in the early 70's (Lemoine, 1971; Michard and Schumacher, 1972). All samples described below, except serpentinite and marble, were taken in the Mte Plum scree, which offers a vast array of material, but can be correlated to the outcrops.

Serpentinites (sample GR 20 85) are massive rocks displaying a brecciated appearance (Fig. 11). No distinct layering, nor alternating layers with different sizes, can be seen. A mineralogical distinction between clasts and matrix is not apparent. We conclude that the brecciation is of tectonic rather than sedimentary origin as proposed by Michard (1967). The thin section displays relics of either pyroxene or olivine, in a matrix of fine-grained serpentine, chlorite and/or talc.

Metabasites (samples GR 20 86-88) are finely laminated-foliated, displaying varied combinations of sodic amphibole ("glaucophane"; Additional file 3: Fig. S1), epidote, Fe-rich chlorite, titanite and sulfides. The main foliation is defined by the shape fabric of blue

Table 2 Sample locations and RSCM results (TRSCM)

Sample	Locality	Coordinates (degrees)	Nature	Formation, unit, Fig. #	R2		T_RSCM(°C)			
					Mean	SD	Mean	SD	SE	N
AC01 L	Battuira	44.4274/7.2128	Calcschists (cs)	Valliera meta-lmsts, (VGU). Figures 6a, 7a	0.5	0.02	416	11	3	10
AC02 L	Valliera soprana	44.4255/7.2130	Quartz-schist	Valliera beds (VG unit)	0.4	0.03	461	16	5	10
AC03 L	Valliera-Colletto	44.4153/7.2173	Liassic lmst-cs	Narbona Member (VG u.)	0.46	0.04	428	6	1	11
AC04 L	P Colletto	44.4127/7.2175	Meta-siliceous lmsts	Hettangian/ Lower Sinemurian (VG unit)	0.45	0.06	441	29	8	11
AC06 M	~300 m NW Pta Castellar	44.4069/7.1848	Metabasite	Contact QN/VGU	–	–	–	–	–	–
AC07 L	~700 m NNE Pta Castellar	44.4100/7.1903	Meta-siliceous lmsts with dolostone grains	Top of Mid-Upper Liassic Lmbr (VG unit)	0.56	− 0.6	390 °C	28	9	10
AC08 L	Above Chiappi	44.4041/7.1771	Meta-calcaro-dolomitic banded, coarse breccia	Main part of "Lmb" (VG u.) Fig. 6d	0.63	0.05	358	25	6	13
AC09 M	1st bridge upstream Pradleves	44.4128/7.2652	Silvery quartz-schist	Permian (?), part of CLSZ. Figure 7d	0.54	0.07	400	35	–	2
AC11 L AC11 M	Lottulo	44.4910/7.2301	Yellowish calcschist	Anisian-Lower Ladinian (VM-S unit). Figure 6h	0.28	0.07	507 512	29 34	9 –	10 10
AC12 L	Lower Vallone Elva	44.4980/7.1010	Dolostone with oblique cleavage	Norian M. Bettone (VG u. in Val Maira). Figure 6f	0.30	0.03	506	14	4	11
AC13 M	Upper Vallone Elva	44.5400/7.0801	Schistes lustrés	QN above M. Bettone	0.29	–	509	26	–	5
AC14 LM	100 m W Sampeyre Pass		Ctd-bearing black schists	Queyras nappe	0.41	0.03	455	16	5	11
AC15 LM	300 m W Sampeyre Pass	44.5521/7.1173	Idem	Queyras nappe	0.44	0.03	442	17	5	11
AC16 L	Costa Cavallina	44.5464/7.1234	Lmst in calcschists (chocolate table)	CLSZ	0.35	0.05	482	25	6	15
AC17 LM	Costa Cavallina	44.5439/7.1294	Serpentinite/meta-rodingite?		–	–	–	–	–	
AC18 LM	East of C. Cavallina	44.5378/7.1453	Bedded metabasite		–	–	–	–	–	
AC19 L	Below S Martino	44,5160/7.1140	Schistes lustrés	CLSZ	0.36	0.10	478	48	–	10
AC20 L	Below Plum klippe	44.4260/7.2220	Schistes lustrés with lmst beds	QN ? Valliera cs (?)	0.33	0.06	502	25	–	13
AC21 L	M. Plum south slope	44.3313/7.2129	M. Plum lmst stuck over serpentinite	QN klippe (?)	0.41	0.04	458	18	5	5
AC24 L	Acceglio anticline core	44.4878/7.1775	Metasedimentary intercalation; Permian (?)	Acceglio-Longet band (AL unit)	0.36	0.05	480	25	7	11
AC26 L	Valmala	44.4912/7.2304	Calcschists	VPSZ	0.23	0.12	534	56	15	13
AC27 L	Above Valmala	–	Mica-schist	Lower (?) Dronero unit	0.52	0.03	407	15	5	10
28 (M21-01)	Battuira	44.4274/7.2128	Triassic carbonate	VPSZ	0.5	0.04	414	21	5	14
30 (M21-03)	Valliera soprana	44.4255/7.2130	Triassic dolostone	SDSZ	–	–	–			
31 (M21-04)	San Damiano Podio	44.4938/7.2644	Quartz-mica-schist	SDSZ			593	24	7	10
32 (M21-05)	San Damiano Podio	44.5112/7.1513	Quartz-mica-schist	SDSZ	0.23	0.1	564	39	10	
33 (M21-06)	Caudano-Centenero road	44.5108/7.1503	Middle Triassic dolostone	Val Maira subunit	–	–	–			
34 (M21-07)	Caudano-Centenero road	44,5108/7,1503	Middle Triassic dolostone	Val Maira subunit	–	–	–			
35 (M21-08)	Centenero village (from a trench)	44.5095/7.1546	Middle Triassic black shales	Val Maira subunit. Figure 7c	0.43	0.03	455	9	2	13

Table 2 (continued)

Sample	Locality	Coordinates (degrees)	Nature	Formation, unit, Fig. #	R2		T_RSCM(°C)			
					Mean	SD	Mean	SD	SE	N
36 (M21-09)	Centenero village (from a trench)	44.5095/7.1546	Middle Triassic black shales	Val Maira subunit	0.41	0.04	455	19	6	10
37 (M21-10)	Ponte Marmora	44.4906/7.0924	Liassic calcschist	VG unit in Val Maira	0.39	0.05	463	24	7	10
38 (M21-11)	Ponte Marmora power station	44.4906/7.0924	Liassic calcschists	VG unit in Val Maira	0.39	0.06	465	29	8	10
39 (M21-12)	Above Ussolo	44.4910/7.0260	Cretaceous (?) Schistes lustrés	Queyras nappe	0.58	0.01	381	8	2	13
40 (M21-13) (a &b samples)	Above Ussolo	44.4910/7.0260	Cretaceous (?) Schistes lustrés	Queyras nappe Figs. 6b, 7e	0.54	0.01	387 399	9 24	3 7	5 10
GR20-85	Below Mte Plum	44.4316/7.2139	Serpentinite	Sole of M. Plum klippe	–	–	–	–	–	
GR20-87 & 88	Monte Plum-Bars la Chiau	44.4320/7.2197	Metabasite	Klippe of the Queyras nappe	–	–	–	–	–	
GR20-89	Monte Plum-Bars la Chiau	44.4300/7.2246	Banded metachert	Klippe of the Queyras nappe Figs. 6g, 7f	–	–	–	–	–	
GR20-90	Monte Plum-Bars la Chiau	44.4300/7.2246	Marble	Klippe of the Queyras nappe	–	–	–	–	–	
GR20-91	Monte Plum-Bars la Chiau		Banded metabasite	Klippe of the Queyras nappe. Figure 7g	–	–	–	–	–	
41 (Bel73-13)	Grangie Sagneres, summit 2985 west of Pelvo d'Elva		Jd-orthogneiss (Permian?)	Acceglio-Longet band (AL). Figure 7h	–	–	–	–	–	

Unit acronyms (AL, etc.) as in Fig. 3 and Table 1. Other abbreviations: cs, calcschists; Lmbr: Middle Liassic breccias; lmst: limestone (protolith); u.: unit. Figures showing outcrop or thin section relative to any sample are indicated for convenience. R2: Raman parameter proposed by Beyssac et al. (2002) to quantify graphitization processes; N: number of Raman spectra; SD: standard deviation; SE: standard error (SE = SD/√N)

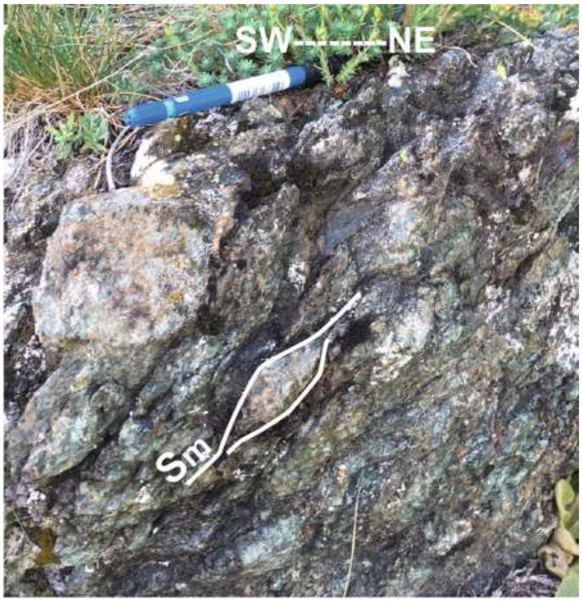

Fig. 11 Serpentinite from the sole thrust of the Mte Plum klippe: lens-shaped, unfoliated serpentinite bodies of various size surrounded by a fine-grained serpentinite matrix showing a mylonitic foliation (Sm), likely polyphase (S2–S3). Location: ~ 50 m to the NE of sample AC21 in map Fig. 10 and point "sr" in cross-section Fig. 5

amphiboles, and the alignment of tiny crystals of titanite. Epidote seems to partly overgrow this foliation. Tight folds defined by the shape fabric of amphiboles show that the main foliation is a composite one. Another sample of metabasite (sample GR 20 91) displays a quartz vein, parallel to the foliation, and shows compositional layers a few mm thick.

Metacherts (sample GR 20 89) are finely foliated and layered rocks, the main constituents being quartz, blue amphibole (elongated, defining the foliation, strongly pleochroic, displaying occasionally boudinaged crystals), chlorite (pale green, with low interference color, so a magnesian variety), epidote and opaques (mostly sulfides). The glaucophane (Additional file 3: Fig. S1) is commonly overgrown by actinolite (Fig. 9f). Some layers are richer in quartz, some almost exclusively made of quartz. White mica is occasional, and present in larger quantities in a few quartz-rich layers, defining very thin plates parallel to the main foliation. Microprobe analyses show a typical phengitic composition (Si = 3.19 pfu).

The *marble* (sample GR 20 90) displays a foliation with mm-sized darker spots in relief on the weathered surface (fossils? albite? lawsonite?). In thin section, the spots are recognized as calcite aggregates that do not present diagnostic features allowing the identification of their precursor.

To conclude, the Mte Plum outcrops offer the same mineral assemblages as the neighboring Schistes lustrés, consistent with their interpretation as a klippe of the Queyras nappe.

4.2 Raman spectroscopy of carbonaceous material (RSCM)

The grade of graphitization of carbonaceous material (CM) is used to evaluate maximum temperatures reached in the sampled material using the empirically calibrated RSCM method. This method, based on the characterization of the CM structures by Raman spectroscopy, allows one to calculate the peak temperature in the range 200–650 °C with a precision generally better than 50 °C (Beyssac et al., 2002; Lahfid et al., 2010). The peak temperature (T_{RSCM}) recorded by CM during the thermal transformation process is insensitive to the retrograde path of rocks or the overprint events. During the last decades, RSCM geothermometry has been frequently used to decipher the thermal evolution of the internal Western Alps, particularly that of the Piemonte-Liguria metasedimentary "Schistes Lustrés" and the Briançonnais *s.l.* (Beyssac et al., 2002; Gerber, 2008; Gabalda et al., 2009; Plunder et al., 2012; Negro et al., 2013; Angiboust et al., 2012, 2014; Schwartz et al., 2013; Lanari et al., 2012; Lefeuvre et al., 2020; Manzotti et al., 2021). We applied this empirical geothermometer to 29 samples selected among the metasedimentary rocks (according to their apparent richness in carbonaceous material) of the MGA and juxtaposed units (Fig. 10). Details on the analytical method can be found in Delchini et al. (2016).

The results are first presented in Fig. 10 and Table 2. In the *Val Grana unit*, 9 samples yielded results between 358 ± 25 and 506 ± 13 °C, with the lowest T_{RSCM} values observed in the southernmost part of the unit. Four samples from Middle Triassic beds of the *Val Maira-Sampeyre unit* yielded higher results, ranging between 455 ± 19 °C and 512 ± 33 °C. Significantly, the highest T_{RSCM} values correspond to the deepest, Anisian-Ladinian part of the Val Maira subunit whereas the lowest values correspond to the uppermost, Ladinian part.

In the *Queyras "Schistes lustrés"*, the two samples from the most external part of the nappe yielded low T_{RSCM} values (381 ± 8 and 387 ± 6 °C) with respect to those from the inner part of the nappe and the Mte Plum klippe (five results from 442 ± 16 °C to 509 ± 25 °C).

Six samples were collected in the *mélange-bearing shear zones*. In the Cima Lubin shear zone, T_{RSCM} values of 478 ± 48 °C and 400 ± 35 °C were obtained in Val Maira and Val Grana, respectively. Higher values at 593 ± 24 and 564 ± 39 characterize the quartz-schists samples from the San Damiano shear zone. The Valmala-Piasco shear zone only yielded discordant values at 534 ± 56 °C and 414 ± 22 °C.

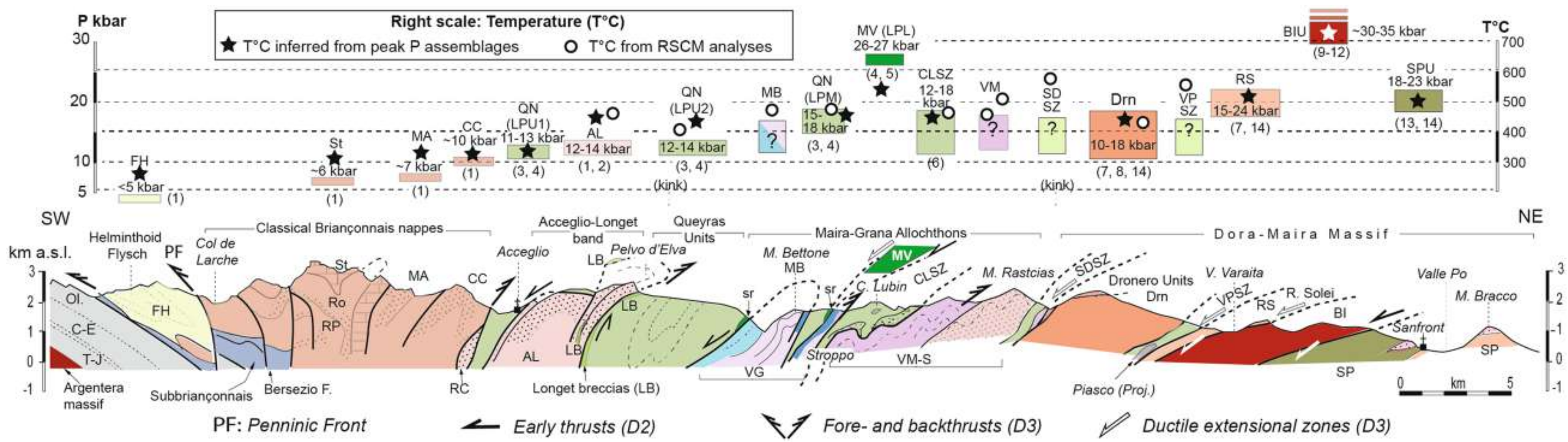

Fig. 12 Peak P–T metamorphic conditions from the literature (1–14) and T_{RSCM} values (this work) along a cross-section of the southern Cottian Alps. See trace of cross-section in Figs. 3 and 10. *Caution*: the two scales (P on the left, T on the right) are independent. Western part of cross-section (Briançonnais nappes west of Acceglio) after Gidon et al. (1977); central part after Michard (1967); eastern part (Dora-Maira units) after Avigad et al. (2003). Acronyms as Fig. 3, with CC: Ceillac-Chiappera unit; C-E: Cretaceous-Eocene; MA: Marinet-Aiguilles de Mary; MB: Mte Bettone; Ol: Oligocene ("Grès d'Annot"); QN: Queyras nappe divided into LPU1/LPU2/LPM = Liguria-Piemonte Upper1 & 2/Medium units in Herviou et al. (2021); Ro: Rouchouze; RP: Rocca Peroni; sr,: serpentinite; St: Sautron; T-J: Triassic-Jurassic. 1: Michard et al. (2004); 2: Houfflain and Caby, 1987; Schwartz et al. (2000a); 3:Agard et al. (2001), Goffé et al. (2004); 4: Schwartz et al. (2013), Lefeuvre et al. (2020), Herviou et al. (2021); 5: Angiboust et al. (2012); 6: Mondino (2005); 7: Henry et al. (1993); 8: Castelli et al. (2014); 9: Chopin and Schertl (1999); 10: Castelli et al. (2007); 11: Ferrando et al. (2009); 12: Campomenosi et al. (2021); 13: Avigad et al. (2003); 14: Groppo et al. (2019)

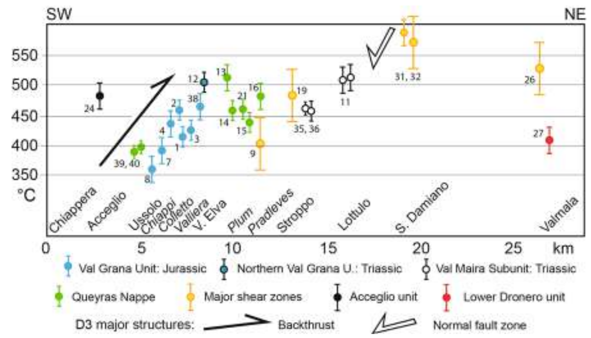

Fig. 13 Diagram T_{RSCM} /distance from Chiappera (west of Acceglio village) along the SW–NE transect of Fig. 12 (see trace in Fig. 10). Localities from the Grana valley in italics. Small numbers in the graph indicate the analyzed samples (see Table 2)

Additionally, a Permian schist sample from the *Acceglio anticline* yielded a T_{RSCM} value at 480 ± 25 °C.

Leaving aside an isolated result from the lower Dronero unit, the T_{RSCM} values obtained in the present work are all consistent with the blueschist-facies mineral assemblages reported in the study area (Sect. 4.1).

5 Discussion

5.1 The South Cottian metamorphic wedge

The MGA and neighboring units are part of the tectonic-metamorphic wedge of the Western Alps. In the deep regional cross-section (Fig. 6), they appear as two distinct units separated by a major shear zone (SZ) made up of a tectonic mélange of continental and oceanic rocks (Cima Lubin Shear Zone). In the following, our discussion will concern the entire wedge except for its most external part, i.e., the Briançonnais *s.str.* nappes described elsewhere (Gidon, et al., 1977, 1994; Lefèvre, 1982; Michard et al., 2004). Hence, the discussion will also concern the deepest units of the wedge, which belong to the southern Dora-Maira massif units and include the coesite-bearing Brossasco-Isasca unit.

5.1.1 Metamorphism

Figure 12 shows the peak pressure and temperature values that characterize the tectonic units exposed in the study area. Part of these values refer to the peak pressure conditions of metamorphism and are compiled from the literature, while the peak temperature values result from the RSCM analyses carried out in this work (chapter 4.2). The latter values are shown with more detail in a complementary diagram T_{RSCM}/distance along the same transect (Fig. 13). Figure 12 illustrates the increase of metamorphic grade in the Permian-Mesozoic continental units derived from the Briançonnais *s.l.* paleogeographic domain, when going from the external classical Briançonnais units (Sautron and Marinet-Aiguilles de Mary: ~ 6–7 kbar, 300–340 °C) to the more internal Ceillac-Chiappera unit (~ 10 kbar, 330 °C), to the Acceglio-Longet band (~ 12–14 kbar, 430–480 °C). Within this antiformal band, we note a reasonable fit between our T_{RSCM} result (480 ± 25 °C) and those inferred from the mineral assemblages (430 ± 20 °C, Michard et al., 2004; 450 ± 25 °C, Schwartz et al., 2000a). Petrological estimates of peak P–T conditions are not available for the Triassic-Liassic series of the Maira-Grana Allochthons. However, T_{RSCM} values from the Val Grana unit are in the same range as those from the Acceglio-Longet band, mainly grouped between 420 and 470 °C, with slightly higher T in the northernmost and possibly deeper rocks (samples #38 and 12, Fig. 13). Likewise, the Val Maira-Sampeyre unit shows T_{RSCM} values in the range ~ 460–510 °C.

The Dora-Maira units underlying the MGA across the San Damiano Shear Zone (Fig. 6) exhibit contrasting P–T values that do not correlate with tectonic position (Fig. 12). The lower Dronero unit records blueschist-facies P–T conditions comparable to those of the Val Maira-Sampeyre unit (Henry et al., 1993), while the overlying upper Dronero unit exhibits eclogite-facies assemblages (Balestro et al., 1995). Beneath the Valmala-Piasco Shear Zone, the Rocca Solei unit and underlying Brossasco-Isasca unit exhibit Alpine quartz-eclogite and coesite-eclogite-facies conditions, respectively (Chopin and Schertl, 1999; Castelli et al., 2007; Ferrando et al., 2009; Groppo et al., 2019; Campomenosi et al., 2021). The lowermost Sanfront-Pinerolo unit shows upper-blueschist to eclogite-facies conditions of equilibration (Avigad et al., 2003; Groppo et al., 2019). Our T_{RSCM} value for sample #26 from the Valmala-Piasco Shear Zone (534 ± 56 °C) compares favorably with the peak temperature at 500–520 °C inferred by Groppo et al. (2019) for the underlying Rocca Solei eclogites. In contrast, the available T_{RSCM} values for the San Damiano Shear Zone (564 ± 39 °C and 593 ± 24 °C) appear significantly higher than the temperature close to 450–470 °C in the lower Dronero unit during blueschist-facies metamorphism (Groppo et al., 2019). This apparent discrepancy cannot be explained yet and requires complementary analyses.

The P–T conditions of metamorphism of the Piemonte-Liguria nappes have been thoroughly analyzed north of our study area (Agard et al., 2001, 2009; Schwartz et al., 2013; Lefeuvre et al., 2020; Herviou et al., 2021). Our T_{RSCM} results for the Queyras Schistes lustrés samples #39–40 and #13–16 (Fig. 13) compare with those obtained further to the north by Schwartz et al. (2013) in their Medium Temperature Blueschist (MT-BS) unit that overlies the Acceglio-Longet Band, and in their High Temperature Blueschist (HT-BS) unit overlying the Monviso Complex, respectively. These temperature values are equivalent within uncertainty to the peak metamorphic temperatures inferred by Lefeuvre et al. (2020) and Herviou et al. (2021) for their LPU2 and LPM units, which correspond to the MT-BS and HT-BS units of Schwartz et al. (2013), respectively. The Monviso Complex contrasts with the Queyras units by its higher, eclogite-facies grade of metamorphism, bordering the coesite stability field (550 °C, 26–27 kbar) in the Lago Superiore subunit but reaching lower peak P–T conditions in the Monviso subunit (500 °C, 22–24 kbar; Angiboust et al., 2012; Schwartz et al., 2013). The T_{RSCM} value of 478 ± 48 °C we obtained in the calcschists of the Cima Lubin Shear Zone in the southern prolongation of the Monviso Complex is consistent within uncertainty with the values obtained by Schwartz et al. (2013) in the Monviso eclogites.

5.1.2 Deformation phases: from early thrusting to backfolding

At the outcrop and thin section scales (see chapter 3.4), and in line with Schumacher (1972) and Caron et al. (1973), we recognized three major "phases" of deformation D1 to D3 in the metamorphic units that constitute the South Cottian wedge.

D1 produced S1 that is subparallel or parallel to lithological contrasts (S0). F1 folds are rare and mostly missing. This primary foliation S0-1 is folded by F2 folds. The D1 phase occurred under peak-pressure conditions in the blueschist-facies MGA and overlying Queyras Schistes lustrés (Caron et al., 1973; Schumacher, 1972). In the underlying eclogite-facies units of southern Dora-Maira, Henry et al. (1993) recognized an early (U)HP foliation deformed by syn-greenschist facies folds, whereby the correlation of these folds with the structures in our area of investigation remains unclear.

D2 is occasionally well defined by isoclinal to tight folds (Fig. 8g), which transpose the earlier foliation S1 into the main foliation S2. The early HP minerals are deformed in the hinges of such F2 folds and partly transformed into greenschist-facies minerals (Fig. 9f). Therefore, the new foliation S2 developed during exhumation of the blueschist-facies units. On a larger scale, exhumation of high-pressure rocks is commonly seen to go hand in hand with nappe stacking associated with buoyant uplift and/or extrusion of nappe bodies within the subduction channel together with erosion and collapse of the uppermost parts of the wedge (e.g., Chemenda et al., 1996; Malavieille et al., 1998; Bucher et al., 2004; Brun and Faccenna, 2008). Notice that distinguishing F2 from subsequent F3 folds is often not easy (Fig. 9a). In the case of the Dora-Maira continental units, the equivalent deformation phase related to exhumation is represented by syn-greenschist facies mylonitic structures (Henry et al., 1993; Avigad et al., 2003).

D3 in the units SW of and overlying the Dora-Maira massif corresponds to the back-folds and -thrusts clearly visible at the large scale (Fig. 6) and generally associated with an S3 foliation. F3 back-folds are observed at all scales, up to kilometric (Figs. 4, 8e, 8h). S3 commonly appears as a crenulation and/or pressure solution cleavage associated with scarce greenschist-facies recrystallizations (Fig. 9a, b, e); in incompetent lithologies it is pervasive and associated with very large strains related to top-ENE to –NE shearing. The S3 schistosity swings around from a NNW-SSE strike to a WNW-ESE strike in the northwest and southeast parts of the study area (Figs. 3, 7), respectively. This is the result of late-stage oroclinal bending of the southernmost Western Alps in connection with the rotation of Corsica-Sardinia after about 20 Ma ago (e.g., Schmid et al., 2017). The D3 backthrusting structures are absent within the Dora-Maira massif that underwent extensional unroofing during the greenschist-facies D3 event. The Pradleves cargneule zone (Fig. 5) could result from the hydraulic extrusion (Fudral et al., 2010) of the Triassic evaporitic rocks (brecciated in the sole of the Val Grana unit) between the Pradleves-Rocca Caire and Valgrana D_3 anticlines.

5.1.3 Timing of the tectono-metamorphic evolution

The available data from the literature point to a Middle-Upper Eocene age for the peak metamorphic conditions in both the continental and oceanic units of the working area. The eclogites from the Monviso complex equilibrated at ~45 Ma (Rubatto and Hermann, 2003; Rubatto and Angiboust, 2015) or ~51 Ma (Garber et al., 2020), while the overlying metasedimentary unit yielded $^{40}Ar/^{39}Ar$ ages from ~60 to ~50–45 Ma (Agard et al., 2002). The age of UHP metamorphism in the Brossasco-Isasca unit is now established at ~35 Ma (Gebauer et al., 1997; Rubatto and Hermann, 2001; Gauthier-Putallaz et al., 2016; Xiong et al., 2021). The other, surrounding Dora-Maira units, for which geochronological data have long been lacking, show a trend of peak metamorphic ages younging downwards in the nappe stack, from ~40 Ma to ~33 Ma, according to rutile U-Pb dating by Bonnet et al. (2022). All the Dora-Maira units were therefore part of the subducting Briançonnais margin beneath the Piemonte-Liguria oceanic material and the overlying leading edge of the Adria plate at about 40–33 Ma. Hence, it is clear that the Briançonnais *s.l.* distal margin was deeply subducted in the Late Eocene. The oldest structural imprints attributed to a poorly defined D1 phase could have occurred under peak burial conditions affecting the Briançonnais margin. In the southern Dora-Maira units,

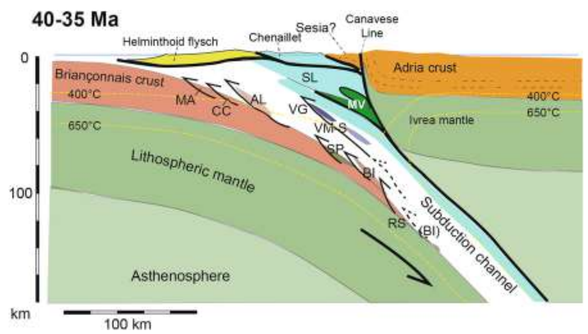

Fig. 14 Interpretation of the origin of the southern Cottian Alps metamorphic units at the expense of the subducted Briançonnais distal margin. The half-arrows feature the locus of detachment where each unit was sampled. Acronyms as follows: A: Acceglio-Longet; BI: Brossasco-Isasca; CC: Ceillac-Chiappera; MA: Marinet; MV: Monviso complex.RS: Rocca Solei; SP: Sanfront-Pinerolo; SL: Schistes lustrés (Queyras nappe); VG: Val Grana; VM-S: Val Maira-Sampeyre; The subduction channel is left "open" (white) for clarity; it is supposed to contain a mélange of Schistes lustrés-type metasediments and meta-ophiolites (mostly serpentinites) from the Piemonte-Liguria Ocean. The ophiolitic Schistes lustrés are divided into a lower-plate accretionary prism and an upper-plate system (e.g., Chenaillet; Schmid et al., 2017; Ballèvre et al., 2020; Agard, 2021). The occurrence of a southern prolongation of the Sesia-Dent Blanche unit in the Cottian transect is dubious but shown here to better suggest the correlation with this northern transect. Late Eocene–Oligocene emplacement of Helminthoid flysch nappes over External Briançonnais after Kerckhove (1963) and Kerckhove et al. (1984). Length and thickness of the Briançonnais crust inspired by the natural examples (e.g., Chenin et al., 2017). Approximate position of isotherms 400 °C and 600 °C after Zhao et al. (2020)

UHP and HP planar-linear and shear fold structures have been preserved in eclogite boudins and escaped the later lower-grade, greenschist-facies overprint (Henry, 1990; Henry et al., 1993). Unfortunately this strongly partitioned deformation did not yield any kinematic information about the earliest movements leading to exhumation.

Nappe emplacement together with folding structures attributed to the D2 phase at a smaller scale, are already associated with retrogression of the HP-LT mineral assemblages to greenschist-facies minerals, retrogression that will continue during the subsequent D3 phase associated with backfolding and extensional unroofing. In other words, building of the tectono-metamorphic wedge mostly occurred at crustal depths although it initiated within the subduction channel (see next subsection) during the D1 phase. Radiometric dating of the D2 and D3 phases are not available for our working area. However, radiometric data for D2 (43–39 Ma) are available in an area further to the north along the ECORS-CROP transect (Villa et al., 2014) suggesting that D2 followed soon after peak metamorphism, which is also likely to be the case for our study area.

In the Gran Paradiso massif, the northern equivalent of the Dora-Maira units (Fig. 1A), 42–41 Ma may be so far the best age bracket available regarding peak metamorphism (Manzotti et al., 2018). In the corresponding transect, Bucher et al. (2004) estimated D3 backfolding to have occurred between 35 and 31 Ma, an estimate that most likely also applies to the Dora-Maira transect, since an along-strike change in the age of this event is hard to imagine. Exhumation and cooling below ~250 °C took place at around 30 Ma based on few on-site zircon fission-track ages available in the Dora-Maira UHP unit (Gebauer et al., 1997) and those from the Gran Paradiso massif (Hurford and Hunziker, 1989; Malusà et al., 2005; Rosenberg et al., 2021). Hence, 30 Ma would approximately date the crossing of the ductile-brittle transition, which would mark the end of the D3 phase. On-site apatite fission-track ages along a N-S transect across the Dora-Maira units yielded ages between 27 and 13 Ma (Beucher et al., 2012). An additional constraint regarding the timing of exhumation around 30 Ma is that (i) during the Early Oligocene, the lowermost beds of the Piemonte Basin started to unconformably overlie the blueschist-facies Briançonnais units and the oceanic units of the Ligurian Alps and Northern Apennine junction (Lorenz, 1986; Mosca et al., 2010; Molli et al., 2010; Maino et al., 2013; Marroni et al., 2017; Piana et al., 2021); (ii) Jourdan et al. (2013) report that short-lived fast erosional exhumation occurred at about 30–28 Ma ago according to an analysis of sedimentary deposits in the fore and retro-deeps of the Alps. In summary, exhumation of the wedge units close to the surface was completed at around 30 Ma ago.

5.1.4 Exhumation scenario

The mechanisms that permit (U)HP-LT continental and/or oceanic units previously subducted down to mantle depth to be subsequently exhumed up to the surface without major T increase has been repeatedly explored since the late 70's (see Guillot et al., 2009 for review). By the earliest 90's, the discussion around the exhumation of the Dora-Maira UHP and HP eclogitic units launched by Chopin (1987) concentrated on two end-member mechanisms, (i) extrusion in the "subduction zone" followed by "corner flow" in the wedge (Henry, 1990; Michard et al., 1993), and (ii) "metamorphic core complex" mechanism of extensional thinning of a previously thickened crust (Avigad, 1992). According to the latter author, the contact of the Brossasco-Isasca UHP unit onto the Sanfront-Pinerolo blueschist-facies unit represents a (former) thrust, unlike the various contacts of the quartz-eclogite unit (Rocca Solei or Dronero; see Fig. 12) over the Brossasco-Isasca unit, considered as low-angle normal faults since they omit parts of the intermediate metamorphic facies.

Unfortunately, no kinematic marker can be directly linked to thrusting under UHP conditions. All the above-mentioned contacts have been similarly overprinted by more steeply inclined top-SW normal fault zones exhibiting greenschist-facies shear bands (Henry, 1990; Henry et al., 1993).

In this discussion, we take advantage of the more recent papers dedicated to exhumation processes in the Alpine belt (e.g., Agard et al., 2002; Levi et al., 2007; Guillot et al., 2009; de Sigoyer et al., 2004; Huet et al., 2009; Burov et al., 2014; Gross et al., 2020; Agard, 2021; Candioti et al., 2021). Thereby we adopt the concept of exhumation in three steps, largely in line with Michard et al. (2004). A first step takes place within the "subduction channel" (well-described in Guillot et al., 2009; this name replaces that of "subduction zone" used in Michard et al., 1993). This is followed by a second step involving nappe stacking and nappe refolding, which forms the collisional wedge. The third step is the extensional collapse of the wedge, which starts while the wedge is still growing and is accompanied by erosion. Figure 14 illustrates a potential scenario for the Maira-Grana Allochthons and adjoining units of the Briançonnais *s.l.* distal margin. They were deeply encroached in the subduction beneath Adria during the first step, and this was followed by the transition towards a second step leading to nappe stacking. In this qualitative model we assume that the Maira-Sampeyre and Val Grana Allochthons (VM-S + VG units) detached from their Variscan basement and were temporarily accreted to the hangingwall of the subduction channel while the Briançonnais crust continued subducting down to greater depth, giving birth to the high-P metamorphism of the Dora-Maira basement units. This hypothesis offers an explanation for the widespread lack of Triassic metasediments above the Dora-Maira units, except for scarce, minor quartzite and dolomite slivers along contacts between basement units (e.g., Valmala-Piasco shear zone) or linked to the Sanfront-Pinerolo unit (Sanfront area, Fig. 3).

The tectonic sketch featured in Fig. 14 represents but one possible interpretation of the structure of the Briançonnais *s.l.* distal margin during the transient stage of an early step in the exhumation of its detached units. Admittedly it is not the only possible interpretation. For example, Butler (2013) and Ballèvre et al. (2020) proposed that the distal part of the Briançonnais margin involved crustal boudinage resulting in at least one extensional allochthon (future Dora-Maira units) separated from the main part of the marginal crust by exhumed serpentinized mantle rocks. A similar pre-orogenic setting was also suggested by Méresse et al. (2012) for the Corsica transect, based on the occurrence of "mixed" metasedimentary breccias displaying both oceanic and continental clasts. In the Western Alps, "mixed" breccias have been described by Dumont et al. (1984) in the Prafauchier sequence (Upper Jurassic?) above the "Prepiemonte" Rochebrune unit or next to it (Barféty et al., 1995). Similar breccias could occur in the uppermost beds of the Val Grana unit (Valliera beds) west of Mte Plum (Michard, 1967, p. 241). However, we consider the extensional disruption of continental allochthons like those observed at the toe of the Iberian margin (e.g., Sutra and Manatschal, 2012, their Fig. 3B) as plausible, but unproven in the case of the south Cottian transect.

5.2 Restoring the Briançonnais *s.l.* distal margin

In the previous section, we presented a schematic scenario of the main units that have been sampled and exhumed on top of the still subducting Briançonnais distal margin (Fig. 14). Now we try to restore the original layout of the basement and cover continental units from the Triassic to the Late Cretaceous, i.e., during the extensional evolution of the Briançonnais passive margin. However, it is first necessary to precisely define the various tectonic units representative of this margin.

5.2.1 Relationships between the Maira-Sampeyre and Val Grana units

We defined in the MGA the Val Grana unit on top and the Val Maira-Sampeyre unit below (Fig. 5). They are separated from each other by the thick and continuous Cima Lubin Shear Zone (CLSZ), which includes a mélange of ophiolite lenses and continental schists analogous to those of the Dronero unit. Moreover, the T_{RSCM} cluster around 420-470°C in the Val Grana unit, and around 460-510 °C in the Val Maira unit (Fig. 13). This difference of ~50 °C is likely beyond error and suggests a difference of burial ranging between 4 and 6 km (assuming a geotherm between 10 °C and 8 °C/km, currently accepted in the area; see, e.g., Michard et al., 2004; Groppo et al., 2019; Angiboust and Glodny, 2020; Agard, 2021) between these units during their blueschist-facies metamorphism. The Val Maira-Sampeyre unit itself overlies the Dronero unit across the intervening San Damiano Shear Zone (SDSZ) that also includes ophiolites similar to the CLZS. We may infer that the Val Grana and Val Maira-Sampeyre units have been located a few kilometers away from each other along the subducting Briançonnais distal margin before being detached and thrusted with ophiolites and basement slivers of the mélanges above and below (Fig. 14). The sedimentary sequence of the Val Maira-Sampeyre unit encompasses the Lower and Middle Triassic whereas that of the Val Grana unit includes Middle and Upper Triassic to Lower and possibly Middle-Upper Jurassic sequences (Michard, 1967; Additional file 1: Fig. S1). Therefore, it is tempting to propose that the Val Grana sequence could correspond to the upper part of the incomplete Val Maira-Sampeyre sedimentary sequence. In this perspective, we may propose that the Val Grana unit detached from the Val Maira-Sampeyre unit and ceased subducting by underplating at a depth ~50 km while the truncated lower unit kept subducting for some kilometers more. Detachment of the Val Grana unit would have occurred on some Ladinian tuffite beds.

Alternatively, we could hypothesize that the Val Grana unit detached from the Val Maira-Sampeyre unit across a low-angle normal fault during the late extensional thinning of the Briançonnais distal margin (Late Cretaceous-Paleocene; chapter 5.2.3). Such a normal fault could be compared with the normal fault (subsequently inverted) that separated the Val Maira sub-unit from its former Sampeyre base (Fig. 4). However, significant Upper Cretaceous-Paleocene breccias that could have characterized such faults have not been observed yet, neither in the Val Maira sub-unit nor in the Giulian-Sea Bianca unit, its northern equivalent (Fig. 3; Balestro et al., 2011).

5.2.2 Early Jurassic rifting: the Val Grana record

The Val Grana sequence is well known since many years for its chaotic to bedded breccias partly dated by Sinemurian and Pliensbachian ammonites (Franchi, 1898; Sturani, 1961; Michard and Sturani, 1963; Ellenberger et al., 1964; Michard, 1967). Therefore, this sequence clearly compares with the "Prepiemonte" units repeatedly described since the seminal works of Ellenberger and Lemoine (1955), Ellenberger (1958) and Lemoine (1961) in a more external (less distal) position, i.e., along the internal border of the classical Briançonnais zone (e.g., Deville, 1986; Dumont et al., 1984; Egal et al., 2020; Jaillard, 1987; Lemoine et al., 1986; Pantet et al., 2020). The latter Prepiemonte sequences are generally detached below the Carnian-Norian (Mégard-Galli and Baud, 1977), contrary to the Val Grana unit, which is detached along an older, Middle Triassic incompetent layer. Prepiemonte-type units are also found as isolated mega-boudins or blocks embedded within the Queyras Schistes lustrés (Roc del Boucher, Caron, 1971; Péouvou, Lemoine, 1971; Gidon et al., 1994; Tricart et al., 2003; Fig. 1A). This compares with the position of the Val Grana unit sandwiched between the Queyras nappe and the ophiolite-bearing mélange of the Cima Lubin shear zone. The Val Grana sequence makes a link between the Prepiemonte units of the Western Alps and those of the Ligurian Alps, well exposed in the Arnasco-Castelvecchio overturned fold nappe (Fig. 1A; Boiteau, 1971; Vanossi, 1990; Seno et al., 2005; Decarlis and Lualdi, 2011; Decarlis et al., 2015). In the rootless nappes of the French-Swiss Prealps, it is the Nappe de la Brèche that represents a Prepiemonte-type unit, being thrusted over the underlying Briançonnais units of the "Médianes rigides" (Lemoine, 1967; Lemoine et al., 1986, their Fig. 19; Steffen et al., 1993). The sedimentary facies of the protoliths of the Val Grana breccias (Additional file 1: Fig. S2a, b) perfectly correspond to that of the Liassic part of the Prepiemonte units quoted above. These breccias record the main rifting phase that prepared the formation of the Piemonte-Liguria Ocean. Their formation results from a protracted fault activity that began as early as the Rhaetian-Hettangian. This is clearly shown in the Rochebrune unit southeast of Briançon (Dumont, 1984), and additionally, illustrated in the studied area by the angular unconformity of the Rhaetian-Hettangian beds over the tilted Norian dolostones at M. Bettone (Additional file 1: Fig. S3). Only Sinemurian strata have been dated in the Val Grana breccias, based on (scarce) fossils of ammonite preserved in these blueschist-facies rocks by the phosphatic composition of infilling. In the lower levels, the breccias exhibit fine granulometry and clear vertical grading, typical for turbiditic emplacement. They become coarser in the upper, undated beds, with a clear horizontal coarsening toward the juxtaposed Acceglio-Longet units (Michard, 1967; Additional file 1: Fig. S2a, b).

This kind of lateral coarsening suggests a western source located in the Acceglio *s.l.* domain, consistent with the deep erosion that affected this domain during the Early Jurassic. The post-Sinemurian breccias did not yield fossils, but they compare with the upper beds of the "Lias prépiémontais" from the Rochebrune unit, which yielded a Toarcian ammonite (Dumont et al., 1984). Unfortunately, the Val Grana unit is truncated at its top by the basal thrust of the overlying Queyras nappe. For this reason, the quartz-rich "Formation détritique rousse" (Dogger) of the Roche des Clots-Grande Hoche unit and the overlying polygenic breccias cannot be firmly identified here although the Valliera clastic beds could be their equivalent (see Additional file 1: Fig. 2c). Likewise, the Upper Jurassic-Cretaceous beds described elsewhere (Roche des Clots-Grande Hoche unit: Lemoine et al., 1978; Dumont, 1983, 1984; Grande Motte unit: Deville, 1986, 1990; Nappe de la Brèche and Mont Fort Nappe: Pantet et al., 2020) have not been clearly identified in the Val Grana area (Tibert siliceous limestones? Additional file 1: Fig. 2d).

5.2.3 Late Cretaceous-Paleocene extensional event: the Longet Breccias record

At the western border of the studied area (Fig. 3), a very particular type of polygenic and mostly chaotic breccias

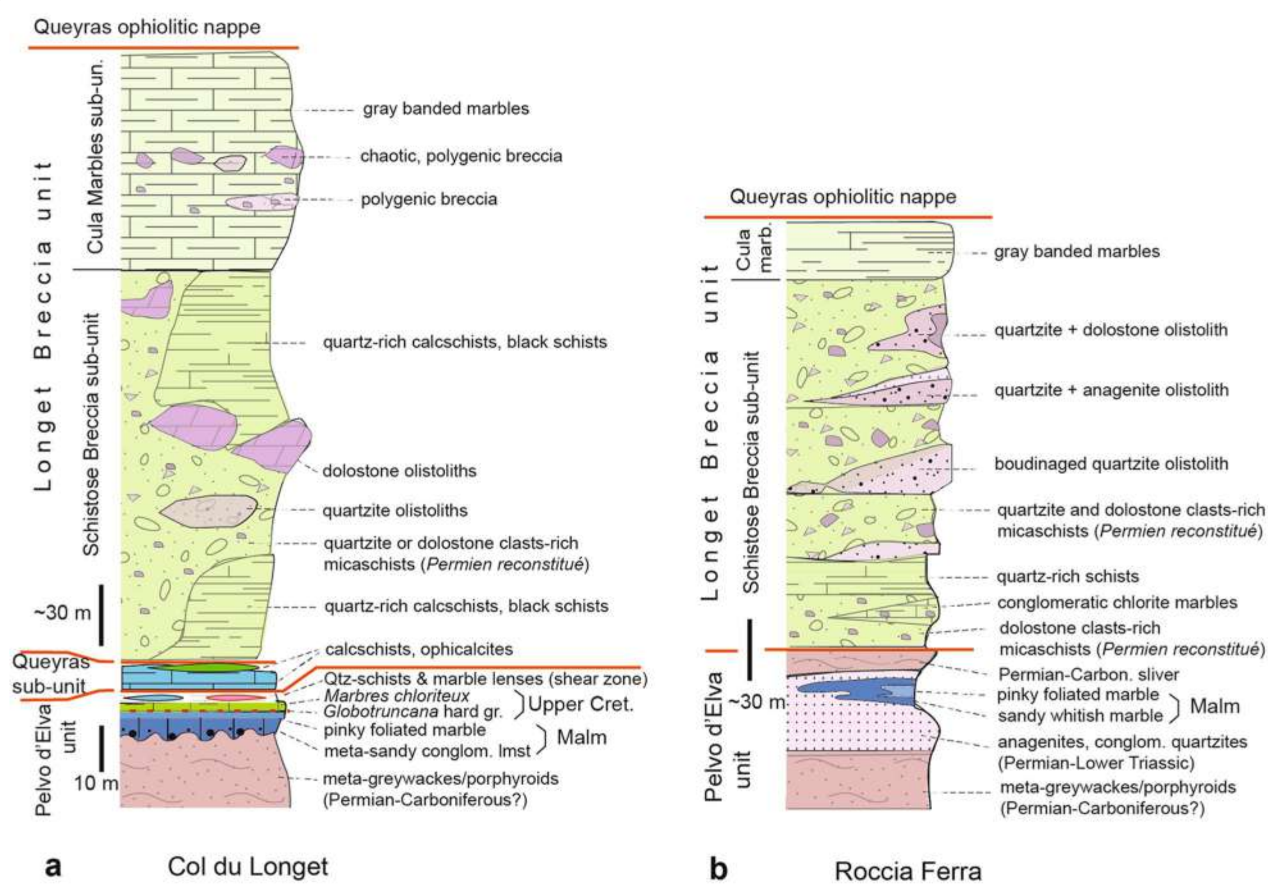

Fig. 15 Lithostratigraphic columns of the Longet Breccia series associated to the Pelvo d'Elva unit. **a** Col du Longet (Upper Ubaye River) outcrops. Mesozoic cover of the Pelvo d'Elva unit after Lemoine (1967) and Lefèvre and Michard (1976); Longet breccias after Lemoine (1960), Leblanc (1962), Gout (1987), and this work. **b** Roccia Ferra (Upper Varaita valley)outcrops, after Michard (1959), Lefèvre and Michard (1976), and Michard (unpubl. observ., 1973–1974). In **b**, the olistolith location and thickness correspond to the observed (inverted) cross-section whereas they are only schematic in **a**

is found, referred to as Longet Breccia(s) since it was described as "Brèche du Col du Longet" by Lemoine (1967). Figure 15 presents the two most representative occurrences of these breccias, both juxtaposed to the Pelvo d'Elva unit, which is the innermost unit of the Acceglio-Longet antiformal band (Fig. 3). The Col du Longet stratigraphic column (Fig 15A) summarizes breccia outcrops exposed on both sides of the Ubaye river, structurally above the hinge of the Pelvo d'Elva F3 antiform and forming a small independent tectonic slice (Leblanc, 1962; Lemoine, 1961, 1963; Lefèvre and Michard, 1976; Gout, 1987), whereas the Roccia Ferra column (Fig. 15B) crops out along the inverted, intensely sheared limb of that fold (Fig. 12; Michard, 1959; Leblanc, 1962; Lefèvre and Michard, 1976), which could partly account for its reduced thickness. Equivalent, although more restricted outcrops of the Roccia Ferra breccias are found beneath the Pelvo d'Elva and neighboring Mt. Maurel summits (Michard, 1959; Lefèvre and Michard, 1976) and further to the SE in Val Marmora (Schumacher, 1972; Gidon, et al., 1977, 1994). A restricted equivalent of the Col du Longet outcrops is observed in Val di Preit above the southernmost tip of the Pelvo d'Elva unit (Fig. 3; Lefèvre, 1982).

The Longet and Roccia Ferra breccias are not directly dated yet, but the following arguments strongly support a Late Cretaceous to Paleocene age: (i) the breccias contain intercalations of "marbres chloriteux", a facies of pale green chlorite-rich calcschists that is commonly observed above the well-dated Upper Cretaceous-Paleocene hard-grounds on top of the Upper Jurassic marbles of the internal Briançonnais *s.str.* units; (ii) a close lithological equivalent of the Longet Breccia is the "Alpet Breccia" that crops out along the overturned sedimentary sequence of the Combrémond unit (eastern margin of the Roure-Combrémond units depicted in Fig. 6, located adjacent to the westernmost Queyras synform). These breccias are in stratigraphic continuity with the thin Acceglio-type cover sequence of the Combrémond unit and its *Globotruncana*-rich hard ground and "marbres chloriteux" (Le Guernic, 1967; Gidon et al., 1994); (iii) the Alpet Breccia contains pinky marble elements showing the classical facies of the Briançonnais Jurassic marbles (Le Guernic, 1967), and (iv) another, well-dated equivalent of the Longet and Alpet breccias is the famous Tsanteleina Breccia of Vanoise that contains fossiliferous "marbres chloriteux" besides of large dolostone olistoliths and "schistes reconstitués" beds (Deville, 1986; Ellenberger, 1958; Jaillard, 1987). Note, however, that the breccias of the Rio Secco window and Eychauda klippe (west of the Chaberton and west of Briançon, respectively; Fig. 1) were formerly also compared to the Longet/Alpet breccias (e.g., Lemoine, 1967), but they are now dated to the Eocene-Oligocene (Barféty et al., 1992, 1995), like those of the Lower Units of Alpine Corsica (Di Rosa et al., 2020; Di Rosa, 2021; Frassi et al., 2022).

The relationship of the Longet Breccia itself with the Pelvo d'Elva unit is ambiguous. At Col du Longet (Fig. 15A), the Pelvo d'Elva unit exposes its typical condensed cover sequence, which yielded Late Cretaceous pelagic foraminifera fossils in hard ground crusts above the Jurassic pinky marbles and below the "marbres chloriteux" (Lemoine, 1960; Leblanc, 1962; Lefèvre and Michard, 1976). There, the Longet Breccia is separated from the Pelvo d'Elva unit by a sliver of Schistes lustrés including calcschists, siliceous marbles and prasinites (Leblanc, 1962; Lefèvre and Michard, 1976; Gout, 1987). Note that this well-documented tectonic setting has been overlooked by Valetti and Mosca (2011). In contrast, from Roccia Ferra to Val Marmora, i.e., along the inverted limb of the Pelvo d'Elva antiform, the Longet Breccia is in direct contact with the Pelvo d'Elva unit, which in places preserves its own thin Upper Jurassic-Upper Cretaceous/Paleocene cover (Fig. 15B). Thus, in view of the Roccia Ferra profile, we consider the Col du Longet setting with an intervening sliver of Queyras elements as a secondary imbrication that resulted from the detachment of the package of breccias during thrusting of the Queyras nappe. According to this interpretation the Acceglio-type Pelvo d'Elva unit formed the former substratum of a basin that was subsiding during the Late Cretaceous-Paleocene, and within which the chaotic breccias were gravitationally emplaced.

The accumulation of this type of breccias involving mega-blocks of quartzite and dolostone with a size often exceeding 10 × 100 m clearly requires normal-fault activity leading to submarine slopes along which the large blocks may fall and slide in the style of a submarine rock fall (Lemoine, 1967). Most of the breccia horizons display a quartz-rich schistose matrix. Thus, the source area displayed large exposures of Middle-Upper Triassic carbonates (mainly dolostones), Lower Triassic quartzites, and weathered Permo-Triassic conglomerates (anagenites) and possibly Permian-Carboniferous volcano-sedimentary series or basement schists. The weathered rocks were eroded and transported as coarse sands forming the matrix of the mega-breccias. This would explain the similarity of the recrystallized matrix with the genuine pre-Triassic rocks, a kind of matrix labeled "Permien reconstitué" by Lemoine (1967). It follows that the source area of the Longet breccias must have been a part of the classical Briançonnais *s.str.* that only became partially eroded before the formation of the breccias, and not an Acceglio *s.l.* unit that lack a significant volume of Triassic dolostones. The Marinet and Aiguilles de Mary units (Fig. 6; Michard and Henry, 1988; Michard et al., 2004) are good candidates for a potential source of the Longet/Alpet breccias, although this raises the problem of the position and role of the Ceillac-Ciappera unit—an issue that is beyond the scope of the present work.

More generally, coarse to chaotic breccias are widespread in the Cretaceous-Paleocene formations of the Briançonnais *s.str.* units, where they are related to normal-fault activity (Bourbon, 1977; Claudel et al., 1997; Tissot, 1955). Likewise, Late Cretaceous-Paleocene faulting is well-described in the Marguareis massif of the Ligurian Briançonnais (Fig. 1 for location), where steeply to moderately dipping, hundreds of meters high paleo-escarpments incise Jurassic to Permian beds, being coated with chaotic breccias (Michard and Martinotti, 2002; Bertok et al., 2012). All these Upper Cretaceous-Paleocene breccias associated with large normal faults record a dramatic, but enigmatic extensional event, which strongly differs from the rifting event of the Early to Middle Jurassic as it accompanies the *closing* of the Piemonte-Liguria Ocean and not its opening. The interpretation of this event is discussed below (chapter 5.2.5).

5.2.4 Restoring the pre-orogenic evolution of the Briançonnais s.l. domain

Several authors have proposed a crustal-scale restoration of the Briançonnais *s.l.* for the time it represented a distal part of the European passive margin separated from the proximal margin by the Valais-Subbriançonnais rift or narrow ocean (cf. Fig. 1B). By "unfolding" the Classical Briançonnais nappes and adding some 20 km more for the Acceglio- and Prepiemonte-type units, Lemoine et al. (1986) suggested a width of ~70 km for the extensionally faulted Briançonnais *s.l.* domain in the Late Jurassic. Schmid et al. (1997) and Stampfli et al. (1998) restored the thinned crust of the north-easternmost and narrower part of the Briançonnais *s.l.* (Tambo, Suretta and Starlera nappes east of the Lepontine culmination; Fig. 1A) to a width of 115 km and 140 km, respectively by the Early

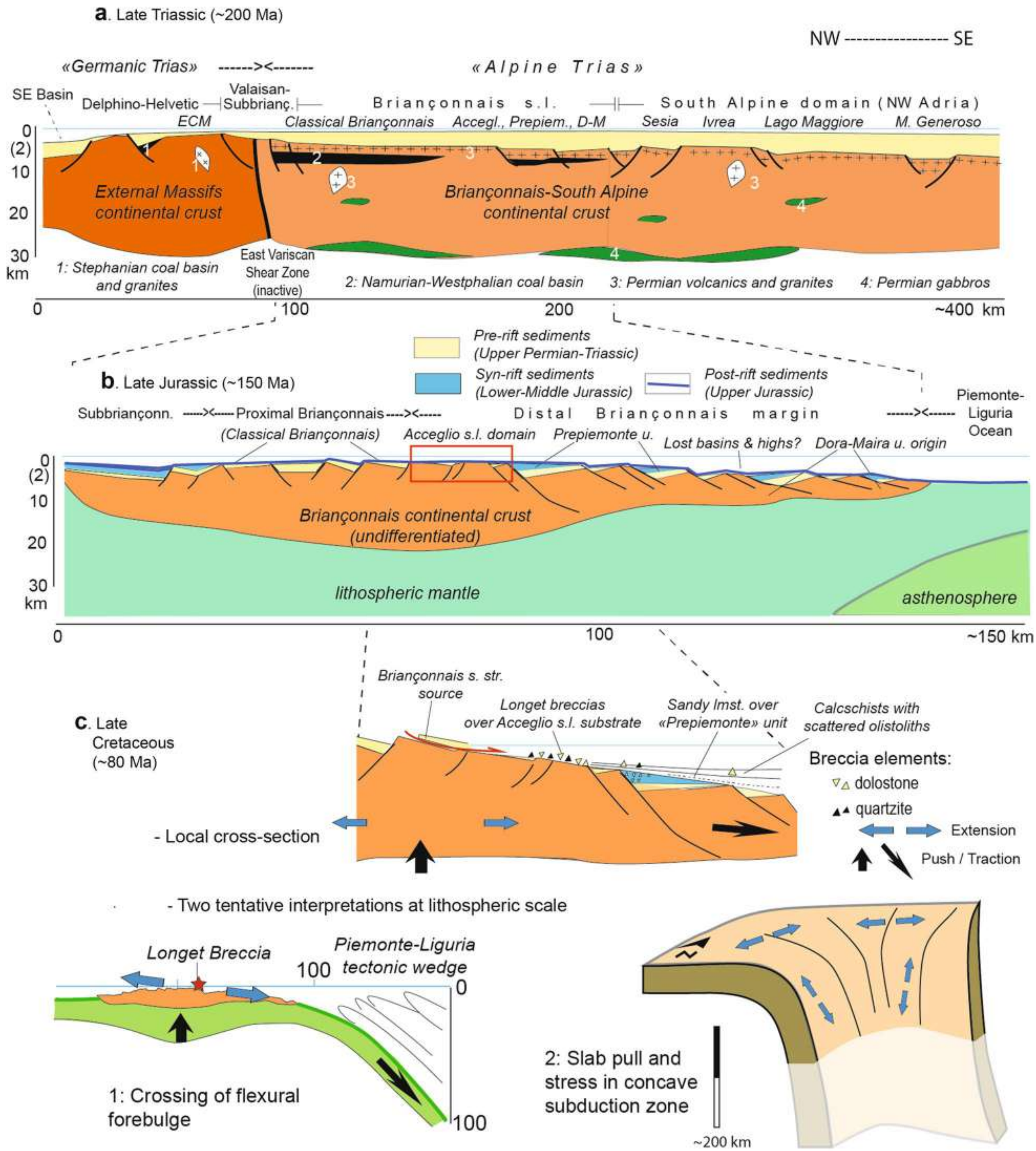

Fig. 16 Pre-orogenic evolution of the Briançonnais domain shown in cross-section (vertical scale strongly exaggerated in the first kilometers depth). **a**: Late Norian (~200 Ma) setting, based on (i) Ballèvre et al. (2018) from the External Crystalline massifs ECM) to Sesia, and ii) Bertotti et al. (1993) for the South Alpine domain (only western half shown). **b**: Kimmeridgian-Tithonian (~150 Ma) stage following the rifting and necking processes, based on (i) Pantet et al. (2020) for the overall aspect; (ii) Mohn et al., (2010, 2012) for mantle and cover features, and iii) this work for the width and thickness of the Briançonnais crust and location of the Briançonnais *s.l.* units of the Cottian transect. **c** Late Cretaceous extensional event (~80 Ma). Local cross-section: restoration of the tectonic-sedimentary setting allowing the Longet-Alpet Breccia to emplace onto the Acceglio *s.l.* units. Two tentative interpretations of the regional extension are shown below; the second interpretation (sketch 2) is favored in this work

Eocene. Recently, Pantet et al. (2020) and Ballèvre et al. (2020) suggested a ~100 km width for the Briançonnais *s.l.* domain in the Swiss Prealps and Classical Briançonnais transects, respectively. Mohn et al., (2010, 2012), however, adopted the smaller width estimated by Lemoine et al. (1986). Hereafter, we discuss the evolution and changing width of the Briançonnais *s.l.* domain during three critical stages: (i) the Late Triassic stage, which offers the presages of the Pangea breakup; (ii) the Late Jurassic stage, which marks the end of rifting and the transition to spreading, and (iii) the Late Cretaceous-Paleocene, which record a new extensional event whose origin is controversial.

5.2.4.1 Triassic period At this time (Fig. 16A), the future Briançonnais domain is characterized by a marine carbonate facies ("Trias alpin") similar to that of the Southern Alps and contrasting with the sandy-evaporitic facies of Western Europe ("Trias germanique" *s.l.*; Baud et al., 1977). In the External Crystalline Massifs, the Germanic-type deposits hardly reach 300 m thickness (Mégard-Galli and Baud, 1977) and include hydrothermally altered alkaline basalt flows; they record the occurrence of a faulted crustal high between the strongly subsiding South-East Basin of Dauphiné-Provence (up to 4000 m-thick Triassic deposits; Le Pichon et al., 2010) and the (more slowly) subsiding Briançonnais-South Alpine domain. The classical Briançonnais Triassic sequence exhibits a maximum thickness of ~800 m (~200 m of Lower Triassic clastics and ~600 m of Middle and Upper Triassic limestones and dolomites; Mégard-Galli and Baud, 1977; Michard et al., 2004). In the "Prepiemonte"-type units, the total thickness of the Triassic series goes up to ~1300 m, cumulating the Lower, Middle and Upper Triassic sequences that generally became tectonically detached one from the other along late "Werfenian" (Olenekian) and Carnian evaporite horizons (Mégard-Galli and Baud, 1977; this work, Fig. 5).

Likewise, in the eastern part of the South Alpine domain, the Triassic sequence is ~1500 m thick whereas it reaches ~4000 m in the faulted Mte Generoso basin and ~6000 m further to the east (Dolomites, not shown in Fig. 16A) where Ladinian volcanism developed (Bertotti et al., 1993). The contrast between the Dauphinois-Helvetic and Briançonnais-South Alpine realms has been analyzed in terms of Variscan inheritance by Ballèvre et al. (2018): (i) granitoids are mainly Late Carboniferous in the Helvetic basement and Permian in the Penninic-South Alpine basement; (ii) small Westphalian D-Stephanian (Gzhelian) basins occur on top of the Helvetic basement whereas a large Namurian-Westphalian (Serpukhovian-Bashkirian) basin extends in the Briançonnais *s.l.* domain ("Zone houillère" and more eastern Pinerolo basin); (iii) Permian gabbros, diorites, granites and volcanic-sedimentary formations are widespread in the Briançonnais-South Alpine domains and virtually absent in the Helvetic domain. Thus, Ballèvre et al. (2018) propose that the Subbriançonnais-Valaisan boundary corresponds to the East Variscan Shear Zone, a major, Late Carboniferous-Early Permian right-lateral strike-slip fault system (Carosi et al., 2020). The minimum width of the Briançonnais *s.l.* continental domain during this time interval may be estimated by "unfolding" its various units from the most external Briançonnais *s.str.* to the most internal, higher-grade Dora-Maira units. Michard et al. (2004) suggested a width of ~100 km, which is retained in Fig. 16A.

5.2.4.2 Liassic-middle Jurassic rifting Tectonics of rifting was repeatedly studied on both sides of the nascent Piemonte-Liguria Ocean. For the northwestern Adria margin, the reader is referred to Bernoulli et al. (1979), Bertotti et al. (1999), Manatschal and Bernoulli (1999), Manatschal (2004), Mohn et al., (2010, 2012), Beltrando et al. (2015), Chenin et al. (2017) among others. Here we focus on the European side (Fig. 16B), e.g., on the Briançonnais domain. After the mild extension that occurred during late Permian (Najih et al., 2019) and Triassic times (Fig. 16A), rifting of the Briançonnais *s.l.* is well recorded by the Rhaetian unconformity and by the Sinemurian-Pliensbachian chaotic to graded breccias of the subsiding faulted basins (Lemoine et al., 1986; see chapter 5.4.3). The Toarcian-Bathonian thermal uplift resulted in subaerial erosion of the proximal Briançonnais platform (Stampfli et al., 1998). Erosion was particularly intense on the crest of the tilted blocks of the rift shoulder, giving birth to the Acceglio *s.l.* units juxtaposed to the Prepiemonte-type units (chapter 5.2.2). Sea-floor spreading started during the Bathonian-Callovian (De Wever and Caby, 1981; Bill et al., 2001b; O'Dogherty et al., 2006; Cordey and Bailly, 2007). The Briançonnais s.str. platform and its shoulder foundered during the Kimmeridgian-Tithonian with renewed normal-fault activity (Claudel et al., 1997) while spreading of the Piemonte-Liguria ocean continued (Le Breton et al., 2020; Lemoine et al., 1986; Li et al., 2013). The structure of the Briançonnais distal margin was mostly achieved by the Late Jurassic (Fig. 16B), waiting for the Late Cretaceous final extensional pulse.

The rift localization between the Briançonnais continental domain (still close to Europe) and Adria was probably favored by a crustal-scale thermal anomaly, established at 215–210 Ma, followed by thermal decay by 200–190 Ma (Beltrando et al., 2015). This thermal anomaly is evocative of the development of the Central Atlantic Magmatic Province (CAMP) further to the west at ~201 Ma (Marzoli et al., 2018). The conjugated Adria and Briançonnais margins that formed subsequently to the thermal anomaly can be compared with those of Northern Atlantic, either with that of western Iberia (Galicia margin), as noticed by Bernoulli et al. (1979) and Lemoine et al. (1986), or with that of Newfoundland, which is directly homologous to the Briançonnais margin (Decarlis et al., 2015, 2017). They all formed as magma-poor passive margins including hyper-extended distal segments (Manatschal, 2004; Sutra and Manatschal, 2012). However, the Piemonte-Liguria case differs from the Northern Atlantic not only in the timing of rifting and drifting (Jurassic *versus* Early Cretaceous, respectively), but also in the major role of the Variscan structures in the shaping of the Briançonnais block (Ballèvre et al., 2020) and in the sinistral component of its displacement along the proximal European margin in relation with the opening of the Valais Ocean (Stampfli et al., 1998), at least from the Early Cretaceous onward. In line with Stampfli et al. (1998) and most recent authors (Mohn et al., 2010, 2012, 2017; Decarlis et al., 2015; Pantet et al., 2020; Ballèvre et al., 2020), we assume that the crust of the Briançonnais domain was some 20–25 km thick in the central (proximal) part of the block, becoming thinner and thinner toward the distal, hyper-extended part in the south-east (Fig. 16B). However, the proposals by the above-cited authors differ from each other concerning the width of the Briançonnais domain, which vary from ~— 50 to 70 km (Mohn et al., 2010; Decarlis, 2015) to ~100–140 km (Pantet et al., 2020; Schmid et al., 1997; Stampfli et al., 1998). We propose a total width of ~150 km, based on two independent datasets, (i) palinspastic restoration of the Mesozoic-Cenozoic units, which leads to ~70–100 km minimum width (Lemoine et al., 1986; Michard et al., 2004), and ii) the peak metamorphic conditions recorded in the subducted units (Fig. 12). Peak pressure estimates in the Brossaco-Isasca unit of southern Dora-Maira implies its subduction down to at least 100 km depth (chapters 2 and 4.1.1; Fig. 14), which means a width of the Briançonnais crust at the maximum of its rifting evolution (Late Jurassic) close to 150 km. This value is in marginal agreement with the 210 ± 60 km width reconstructed by Bonnet et al. (2022) but, as noticed earlier (chapter 5.1.4), we hypothesize that the sediments of the Maira-Sampeyre and Val Grana Prepiemonte-type Allochthons might have been formerly deposited onto the basement of the future Dora-Maira units, before being detached during the subduction process. However, it has to be kept in mind that any paleogeographic restoration of this kind assumes that all parts of the subducted continental margin were subsequently exhumed, excluding those parts that may have been lost in the subduction channel.

5.2.4.3 Late Cretaceous-Paleocene This time interval is heralded by an extensional event that affected most if not all the Briançonnais *s.l.* domain. This event is often overlooked although it is responsible for the formation of breccias scattered in the Turonian-Campanian calcschists of the Briançonnais nappes and frequently associated with large fault escarpments (Tissot, 1955; Bourbon, 1977; Jaillard, 1988; Claudel et al., 1997; Michard and Martinotti, 2002; Tricart et al., 2003; Bertok et al., 2012). In our study area and further to the north along the eastern border of the Classical Briançonnais, the Longet-type coeval breccias exhibit a polygenic, frequently chaotic facies with quartz-micaceous to calcareous matrix (chapter 5.2.3; Fig. 15). In view of the richness of dolostone olistoliths in these breccias, we inferred that they were sourced from a classical Briançonnais unit. We also concluded that the Longet breccias and their equivalents such as the Alpet Breccia were deposited onto an Acceglio *s.l.* substrate (i.e., the Pelvo d'Elva and Combrémond units, respectively). The local cross-section in Fig. 16C schematically depicts the setting of the Longet-Alpet breccias at the internal border of the Briançonnais *s.str.*. However, structural analyses of Late Cretaceous faults are too sparse so far at the scale of the whole Briançonnais domain to allow for an estimate of the total amount of extension of the margin during this time.

Being coeval with plate convergence in the Alpine-Mediterranean area (e.g., Handy et al., 2010), this extensional event is not easy to understand. Crossing of the flexural forebulge linked to the subduction of the European plate (Fig. 16C, sketch 1) was first proposed as a likely explanation by Chaulieu (1992). Stampfli et al. (1998) developed this idea for the northern Briançonnais domain (i.e., the Médianes Rigides of the Prealps) in a semi-quantitative model whereby subduction of the Piemonte-Liguria oceanic lithosphere begins at ~110 Ma beneath Sesia—a timing that is not consistent with the more recent data, which point to a Late Cretaceous onset of subduction (Manzotti et al., 2014). In their study of the Marguareis extensional faults, Michard and Martinotti (2002) also proposed to link the Briançonnais extension active since 75 Ma until the Middle Eocene (some 45 Ma) to its crossing of the forebulge domain. However, the forebulge interpretation is tributary of several parameters, in particular the width of the Piemonte-Liguria Ocean, the distance between the bulge and the subduction trench, and the convergence rate (Allen et al., 1991). Indeed, the forebulge model well applies to infilling of foreland basins that are characterized by generally smooth foreland unconformities devoid of fault scarp breccias (Sinclair, 1997; Yu and Chou, 2001; Mattern et al., 2022). However, this contrasts with the Briançonnais case whose Late

Cretaceous extensional setting rather evokes a more substantial rifting event.

An alternative explanation can be sought by considering the role of (i) slab pull and avalanching, and (ii) curvature of the subduction zone (Fig. 16C, sketch 2). The first mechanism was proposed by Capitanio et al. (2009) to account for the extension of the Sirte basin between Tunisia and Cyrenaica. There, a Cretaceous-Paleocene rifting was superimposed onto the Triassic-Lower Jurassic rifting of North Africa at the southern margin of Neo-Tethys (Frizon de Lamotte et al., 2011). The Late Cretaceous rifting could have been triggered by the rapid growth of the North-African slab pull when the slab avalanched below the 660 km transition zone (Capitanio et al., 2009). The second mechanism results from the extensional fragmentation of the slab linked to the curvature of the trench. Mallard et al. (2016) emphasized that a bend of the trench induces differential motion creating tensile stresses in the concave lower plate, which tends to fragment. The curvature of the Piemonte-Liguria trench increased precisely during the Santonian-Selandian (80-60 Ma), according to current paleogeographic restorations (e.g., van Hinsbergen et al., 2020).

6 Summary and conclusion

This work reviewed older data and combined them with new data collected along the southernmost segment of the Western Alps Schistes Lustrés and associated Prepiemonte units, an area that was left fallow for 50 years despite its crucial interest. The location of this segment is ideal for understanding relationships between more external and non-metamorphic units with internal units metamorphosed up to UHP conditions belonging to one and the same passive margin. In the studied area, which presently overlies the southern tip of the Ivrea body indenter, we have re-defined the large Maira-Sampeyre (Triassic) and Val Grana (Triassic-Jurassic) allochthonous units (MGA units) that both are parts of the former Briançonnais distal margin subducted beneath the Piemonte-Liguria accretionary wedge and Adria margin. The presence of these allochthons allowed establishing a direct link between the classical external Briançonnais units and the internal Dora-Maira crystalline units. We conclude that:

(1) The stratigraphic record in the Val Grana unit dominated by Jurassic strata seemingly completes that of the mostly Triassic series preserved in the Maira-Sampeyre unit. During the subduction process, these units likely detached from the distal Briançonnais crust that was subsequently exhumed to form the Dora-Maira units. They became separated from each other when being accreted to the hangingwall of the subduction channel.

(2) Major shear zones including tectonic mélanges of ophiolites and continental slivers bound the Dronero, Maira-Sampeyre and Val Grana units at their base (Valmala-Piasco, San Damiano and Cima Lubin shear zones, respectively). This yields valuable insight on the early exhumation process that occurred in the subduction channel (Fig. 14).

(3) The orogenic evolution began with the subduction of the Briançonnais distal margin, followed by thrusting and subsequent back-folding and -thrusting, recorded in D1-D2 and D3 ductile structures, respectively. D3 is typically associated with greenschist-facies recrystallizations recording the exhumation of the tectonic-metamorphic wedge.

(4) Blueschist-facies metamorphic conditions affected the MGA during the Late Eocene subduction of the Briançonnais distal margin. The peak thermal conditions have been determined by RSCM geothermometry: T_{RSCM} values range from ~400 °C to >500 °C going from the Val Grana unit and overlying Queyras Schistes Lustrés to the uppermost Dora-Maira (Dronero) unit (Fig. 13).

(5) The MGA geological cross-section (Fig. 5) has been integrated in a larger-scale cross-section (Fig. 6). The Dora-Maira major backfold can be regarded as contemporaneous to the backfolds at the internal boundary of the classical Briançonnais and adjoining Acceglio *s.l.* units (Roure-Combrémond and Acceglio-Longet bands).

(6) We propose a pre-orogenic restoration of the Briançonnais passive margin, based on the palinspastic and metamorphic dataset (Fig. 16). By the Late Triassic, the mildly extending Briançonnais would have a width of ~100 km whereas, after the Early-Middle Jurassic rifting stages, its width would have reached ~150 km. The Prepiemonte-type Liassic breccias of the Val Grana unit accumulated during the rifting event along the internal border of the coevally uplifted Acceglio *s.l.* domain.

(7) The occurrence of a late and substantial rifting event during the Late Cretaceous-Paleocene is emphasized. It is well-recorded in the studied area by polygenic chaotic breccias (Longet-Alpet Breccia) sourced from uplifted classical Briançonnais units and emplaced by tectonic-sedimentary processes over the foundering Acceglio *s.l.* domain (Fig. 16C). This late rifting was superimposed to the early, Triassic-Lower Jurassic rifting of the Briançonnais domain. Despite its importance, the Upper Cretaceous-Paleocene rifting event is not

well understood yet and deserves further structural research.

Supplementary Information

The online version contains supplementary material available at https://doi.org/10.1186/s00015-022-00419-8.

Additional file 1. The "Prepiemonte"-type Val Grana unit; salient aspects and nomenclatural note. Fig. S1. Stratigraphic columns of the Triassic-Jurassic sequences of the Maira-Sampeyre and Val Grana Allochthons, after Michard (1967), modified. **Fig. S2.** Meta-sedimentary facies from the Val Grana unit, extracted from Michard (1967), plates 19 & 20. **Fig. S3.** Rhaetian unconformity onto the Norian dolostones, crest of the Monte Bettone anticline at Colle San Giovanni, after Michard (1967), plate 23. Note: The entangled "Acceglio" and "Prepiemonte" units: a nomenclatural note.

Additional file 2: Metamorphism of Briançonnais rocks. Figure S1. D3 mylonitic structure of the jadeite orthogneiss of Grangie Sagneres (summit 2985), west of Pelvo d'Elva (Acceglio-Longet Band). **Figure S2.** Backscattered-electron images from thin sections of the jadeite orthogneiss of Grangie Sagneres. **Fig. S3.** Macrophotograph of a meta-tuffite sample from the Norian of Rocca Caire anticline (Val Grana).

Additional file 3. Metamorphism of the Mte Plum rocks (klippe of Queyras nappe). Compositional layering in metabasites. **Figure S1.** Composition of the sodic amphiboles from the metacherts and metabasites.

Acknowledgements
We warmly thank Giorgio Martinotti for his generous help in giving us access to useful documents; Giorgio also accompanied us in the field, which he visited with one of us (A.M.) about 30 years ago. Thanks are also due to Francesco Nosenzo who guided us to the best sampling sites of the Valmala wooded slopes. Marco Malusà kindly supplied the CIFALPS profile original figure. Comments on the manuscript by Handling Editor, Jan Pleuger and an anonymous reviewer were much appreciated and led to many worthwhile modifications.

Author contributions
AM, AL, MB, SI and DD took the samples, CC and AM acquired the microscope images, CC the backscattered-electron images, and MB and PM the microscope and microprobe data (M.Plum klippe). AL realized the TRSCM measurements. The paper has been written, and the figures drawn by AM (except for Fig. 6 by SS) and revised by SS, MB, CC, SI, AL. All authors read and approved the final manuscript.

Funding
The field work has been done on their own salary by AM, SS, MB, PM, SI and DD. Funding from BRGM covered AL field work and analytical costs. SI acknowledge support by founds "Ricerca locale" (ex-60) of University of Torino.

Availability of data and materials
All geochemical and geochronological data produced for this paper are available in Additional files.

Declarations

Ethics approval and consent to participate
Not applicable.

Consent for publication
Not applicable.

Competing interests
The authors declare that they have no competing interests.

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Received: 2 December 2021 Accepted: 26 March 2022
Published online: 07 June 2022

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