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Cross-propagation of the western Alpine orogen from early to late deformation stages: Evidence from the Internal Zones and implications for restoration

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A B S T R A C T

The internal zones of the Western Alps arc are derived from an oceanic and continental subduction wedge developed beneath the Adria plate during its paleogene northward drift. Exhumation of the internal zones proceded from early Oligocene onwards due to westward extrusion of the Adria plate. The prominent fold-and-thrust structures which follow the arc shape, either forward or backward verging, postdate the initial nappe stacking and overprint differently oriented older deformations which are relevant to proper retoration of this arcuate orogen to minimise overlap problems. We document this early stacking phase through outcrop-scale structural analysis at 55 sites between the Maurienne and Ubaye valleys, along with larger-scale examples of early structures. They consistently show an initial N- to NW tectonic transport, whose kinematic indicators are overprinted by either forward (W- to SW-directed) or backward (E- to NE-directed) deformation associated with post-nappe transport along the Penninic thrust. Accordingly, restoring the Briançonnais fold/thrust system must incorporate reconstruction of the nappe stack along the initial top N-NW direction of orogenic propagation, with careful consideration of their paleogeographic origin towards the S-SE. This stack was built during the Eocene Adria-Iberia collision, and overthrust the Subbriançonnais-Valaisan trough to the NW before involving the Dauphiné-Helvetic foreland. It includes different types of Paleozoic units, either Permo-Carboniferous sediments towards its base, or polymetamorphic basement above, which can be explained by inversion of a late Variscan basin and of its southern shoulder, whereas the uppermost Prepiedmont units result from inversion of the Tethyan margin toe. Mixed breccia, locally preserved close to the tectonic contact between the latter units and the overlying "Schistes Lustrés" oceanic nappes, are interpreted as olistostromes fed by both units in a very early collision stage. $^{39}Ar/^{40}Ar$ dating suggests that these shallow tectono-sedimentary formations were involved in the subduction wedge during the early Eocene, whereas younger (late Eocene) equivalent olistostromes mark the propagation of the Briançonnais stack over the external (Dauphiné/Helvetic) foreland. The Eocene orogenic wedge was rapidly exhumed during Oligocene westward indentation and radial spreading, in a markedly different tectonic context driven by extrusion around an Adriatic upper mantle indenter, which controlled development of the Western Alps arc in relation with the Ligurian sea opening.

1. Introduction

Despite being one of the most extensively studied mountain ranges in the world, the Western Alps are a very specific part of the Alpine orogen whose kinematic evolution is markedly different from the rest of the chain. Whereas the Alps trend approximately E-W from Austria to Switzerland, a shape easily understandable considering N-S Africa-Europe convergence during the Cenozoic (Rosenbaum et al., 2002), the western arc shows a 180° shift across western Switzerland, SE France and N Italy. This shape was partly inherited from the Mesozoic rifting stage, and mainly developed progressively from the late Eocene onwards (Caby, 1996; Ford et al., 2006; Vignaroli et al., 2008; Dumont et al.,

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2012; Malusà et al., 2015), as an accommodation of westward extrusion, possibly combined with oblique convergence (Laubscher and Bernoulli, 1982; Ricou, 1984) and anticlockwise rotation of the northern part of the Adria plate (Laubscher, 1988, 1991; Malusà et al., 2009; Eva et al., 2020). This non-cylindrical propagation produced a complex and polyphase internal deformation of the subduction wedge preserved between the Adria plate and the European foreland, including fast exhumation and changes in tectonic transport direction through time (Platt, 1986; Ramsay, 1989), which are both characteristic of the Western Alps.

This complex 3D and polyphase deformation history makes the initial architecture of the precursor continental margin, presently involved in the western Alpine arc difficult to restore, particularly because the most prominent structures which define the present-day trend of the arc probably postdate the initial contractional features as they overprint the evidence of the earliest orogenic propagation developed during Eocene times. Moreover, the Adria plate first collided with terranes connected to the eastern Iberia plate, such as the Briançonnais domain (Stampfli et al., 2002; Handy et al., 2010), so that an early part of the convergence history is likely to have been accommodated by oblique contraction and reactivation along the original eastern extent of the Pyrenean orogen, resulting in complex interference between preexisting Pyrenean structures and newly evolving Alpine deformation (Lacombe and Jolivet, 2005; Schreiber et al., 2011; Balansa et al., 2022). Finally, the Alpine structures have experienced more recent fragmentation, in the southern part of the arc, through the development of the Ligurian and Tyrrenian breakups and by the growth of the Apenninic chain, driven by the complex lithospheric motion of various lithosphere slabs (Jolivet et al., 2008; Zhao et al., 2016; Salimbeni et al., 2018).

The surface geology of the Western Alps arc gives a misleadingly simple expression of this history. Radial transects have been regarded as more or less equivalent and comparable with little regard for their relative orientation. However, this approach does not incorporate consideration of the magnitude of oblique to lateral transfer and tectonic transport oblique or parallel to the modern orogenic trend, which were potentially of major importance considering the evidence for oblique-slip motions in the southern part of the western Alpine arc (Butler et al., 1986; Ricou and Siddans, 1986; Laubscher, 1991; Malusà et al., 2009). This current work is focused on deciphering the structural and tectono-sedimentary features related to the early Alpine orogenic stages, which were active before the development of the present-day arcuate shape, and which consist of multi-scale evidence for different tectonic transport directions through time, and possible interference structures. Since the early orogenic propagation is also characterised by surficial interactions between relief, gravity and flexural basin distribution, the potential link between selected tectono-sedimentary breccias and the major tectonic contacts is also examined.

2. Geological setting, overview of the western Alpine arc

The Western Alps (fig. 1a) results from the Cenozoic continental collision between the Adria microplate, a northern portion of the Africa plate (Channell et al., 1979), and the European plate s.l., including the Iberia microplate. The orogen incorporated the late Cretaceous oceanic accretionary wedge (Deville et al., 1992; Schwartz, 2000; Dal Piaz et al.,

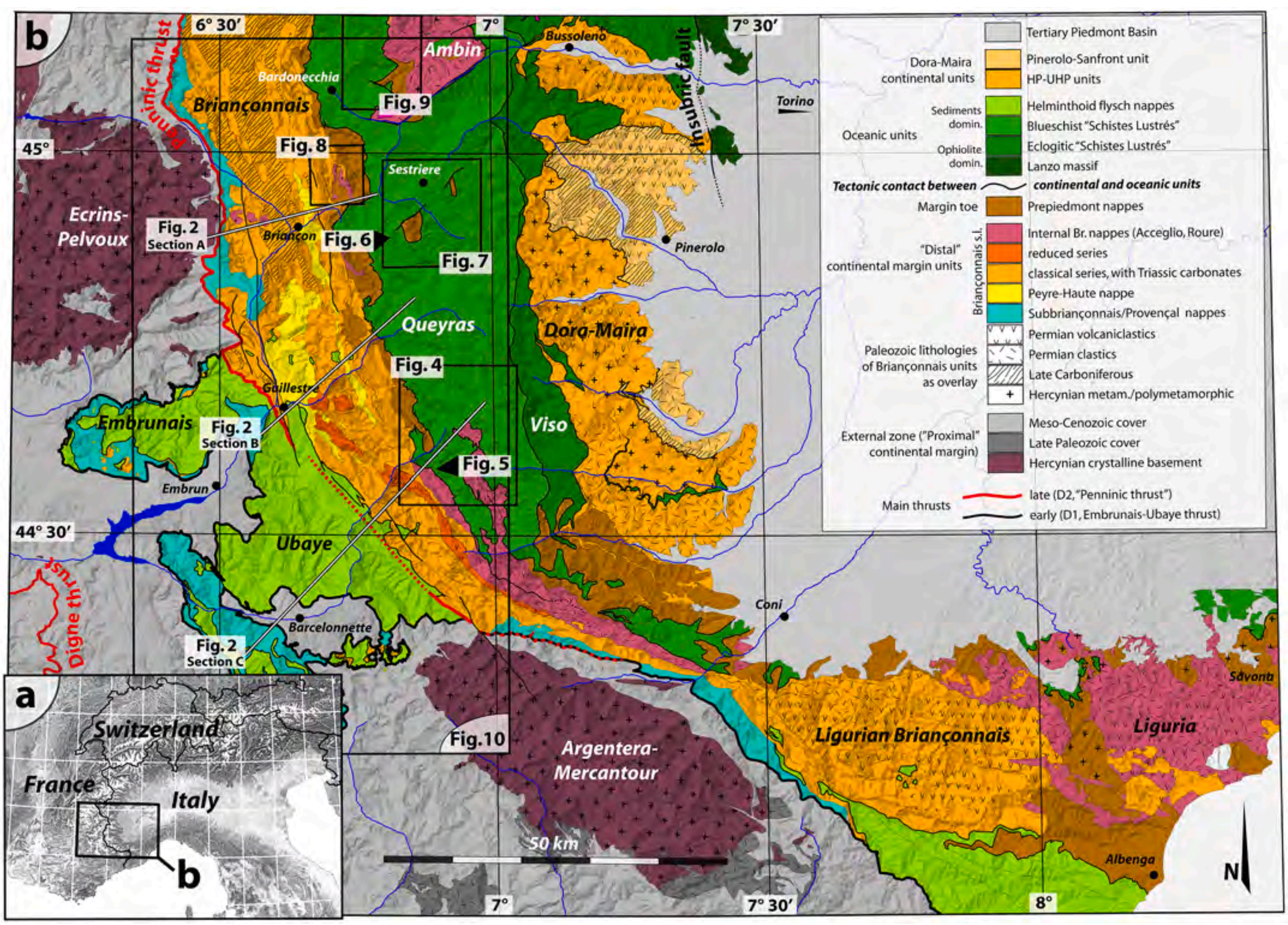

Fig. 1. a- Location of the Western Alps arc in the European framework. b- Geological map of the southern part of the Western Alps arc, with location of the following figures. Colors refer mainly to the paleogeographic origin of structural units, with overlay to distinguish basement and Paleozoic from younger sedimentary cover.

2003; Tricart and Schwartz, 2006; Herviou et al., 2022, and refs therein) produced by the south-verging subduction and closure of the Ligurian Tethys, a small slow-spreading oceanic domain which opened in middle Jurassic times (Bernoulli and Lemoine, 1980; Dal Piaz, 1999; De Graciansky et al., 2011). It also contains parts of the European continental margin of this ocean, variably affected by syn-rift crustal thinning (Lemoine et al., 1986; Manatschal et al., 2007; Le Breton et al., 2021), and an attenuated crust and/or exhumed mantle area, the Valais domain, whose age and extension are still debated (Stampfli, 1993; Bousquet et al., 2002; Beltrando et al., 2007; Pfiffner, 2014). The remnants of units derived from the Adria continental plate margin are relatively scarce in the Western Alps due to syn-orogenic erosion, although their discovery by Argand (1911) laid the foundation of modern Alpine geology.

The Western Alps are conveniently divided into external and internal zones, separated by the « Penninic thrust » (Schmid et al., 2004), a crustal-scale tectonic contact following the arc (fig. 1b) which is E-dipping in the central part of the arc (Guellec et al., 1990; Lardeaux et al., 2006). To the west, the contractional footwall units (External zone) consist of Dauphiné-Helvetic Variscan basement and Meso-Cenozoic series lacking significant Alpine metamorphism. They are locally overlain by exotic sedimentary cover nappes emplaced before the activation of the Penninic thrust (Prealps, Embrunais-Ubaye nappes, Ligurian flysch nappes; Dumont et al., 2012). The External continental crust is deeply subducted eastwards beneath the Internal zones and the Adria plate (Zhao et al., 2015; Nouibat et al., 2022). To the east of the Penninic thrust, the hangingwall nappe stack of the Internal zones is highly heterogeneous both regarding palaeogeographic provenance (De Graciansky et al., 2011) and Alpine metamorphism (Oberhänsli et al., 2004; Bousquet et al., 2008; Agard, 2021), ranging from upper greenschist to eclogite facies metamorphism.

Thus, three main units have to be considered in this area, that is, from W to E, the Extenal zone, the Embrunais-Ubaye nappes, and the Internal zones (Fig. 1b):

- The external zone s.s. is composed of Variscan basement whose local exhumation and uplift due to several thick-skinned shortening events provides some of the major Alpine relief, up to >4km. The top basement is locally buried to ~10km depth in the center of the SE France Basin following several Mesozoic rifting events. The Mesozoic sedimentary cover shows strongly variable thickness and facies across the Jura platform, the Dauphiné rifted margin (Lemoine et al., 1986), the Vocontian basin and the Provence platform. The Mesozoic paleogeographic trends of the External zone are crosscut by the arcuate Alpine structures. The sedimentary cover is locally involved in thin-skinned fold and thrust deformation of different ages and orientations due to Pyrenean and Alpine contractional propagation (Gidon, 1997; Philippe et al., 1998; Espurt et al., 2012; Schwartz et al., 2017). The occurrence of soft evaporitic Triassic layers played a major control on the location of detachments (Lickorish et al., 2002; Espurt et al., 2019; Balansa et al., 2022). Remnants of synorogenic basins record different stages of propagating lithospheric flexure through foreland basin and forebulge development, with flysch and molasse deposition (Joseph and Lomas, 2004; Ford and Lickorish, 2004; Kempf and Pfiffner, 2004; Schwartz et al., 2012; Kalifi et al., 2020).

- The Embrunais-Ubaye nappes were transported in a superficial setting over the flexural basin on the External zone during the late Eocene to early Oligocene (Kerckhove, 1969; Gupta and Allen, 2000). They are dominantly composed of « Helminthoid Flysch », late Cretaceous deep marine sequences detached from the Tethys oceanic floor and transported over the Briançonnais domain whose thin remnant thrust sheets are often observed at their base (Kerckhove, 1969). The Embrunais-Ubaye nappes record only low-grade metamorphism from sub greenschist to lower greenschist facies (Oberhänsli et al., 2004), similar to the Ligurian and Prealps nappe stack. They bear evidence for changes in transport direction (Merle and Brun, 1984) from an original southeastern origin. They are locally preserved over the external foreland in the footwall of the Penninic thrust, which demonstrates the polyphase and non-coaxial character of Alpine orogenic propagation. Their emplacement is dated as late Eocene-earliest Oligocene, with intial NW-directed deformation beneath the Embrunais basal thrust (Dumont et al., 2011; fig. 1b, Fig. 2), and they are overprinted, deformed and crosscut in out-of-sequence mode by the Penninic Thrust propagating towards the SW from early Oligocene onwards.

- The Internal nappe stack includes parts of the distal European margin (detached sedimentary cover and basement of the Briançonnais domain s.l.), exhumed remnants of the Ligurian Tethys ocean (metasediments and ophiolites, the so-called « Schistes Lustrés »), and scarce overthrust pieces of Adria continental crust (Dent Blanche and Cervin units; Dal Piaz, 1999; Schmid et al., 2004). The Internal nappes can be classified according to different criteria: their dominant lithology, either upper crustal basement or sediments, their metamorphic signature (Alpine HP-LT and/or Variscan HT-LP; Handy and Oberhansli, 2004; Bousquet et al., 2008; Von Raumer et al., 2012; Schwartz et al., 2013), and their paleogeographic provenance with respect to the Tethyan framework (Lemoine et al., 1986; Schmid et al., 2004; Handy et al., 2010). In Fig. 1 and 2 we define structural units with respect to their origin, either from the European continental margin s.l. (including parts of the Iberian plate) or from the oceanic domains. Concerning the continental margin units, we follow the definitions of Lemoine et al. (1986), which distinguishes the Subbriançonnais, Briançonnais, internal Briançonnais and Prepiedmont type units. Contrary to Mohn et al. (2010) or Ribes et al. (2019), we have to maintain the distinction between the internal Briançonnais and Prepiedmont units, which have a very different Mesozoic record inherited from their syn-rift history. The ocean-derived nappes still occupy hangingwall locations on both sides of the Penninic thrust (fig. 2), and in the highly metamorphosed core of the arc. The latter include the Monviso and Voltri units, which have been interpreted as remnants of an oceanic subduction channel (Schwartz et al., 2001; Guillot et al., 2004; Federico et al., 2007) but may also result from different intra-oceanic structural inheritance and decoupling processes in a complex plate interface (Balestro et al., 2018; Agard, 2021; Herviou et al., 2022, and refs. therein). The subduction channel process is also involved during the continental subduction stage (Ganne et al., 2006; Bousquet, 2008b, Federico et al., 2005), although alternative processes may explain the exhumation of HP units, such as corner-flow (Polino et al., 1990) or transtension (Malusà et al., 2015). Soon after the Eocene-Oligocene boundary, the initial suture was crosscut by the Penninic thrust and is thus strongly affected by backfolding and tilting due to vertical extrusion of the Internal crystalline massifs from beneath the orogenic wedge (Schwartz, 2000; Rolland et al., 2000; Avigad et al., 2003; Schwartz et al., 2009). At a smaller scale, the « pop –up » structural style of the Briançonnais zone s.l. (fig. 1b, fig. 2) is composed of a stack of nappes involving upper Paleozoic to Cenozoic sediments, probably cored by basement at depth. The exhumation of this « pop-up » and of the associated structures (forward-directed Penninic thrust and backward-directed folds and thrusts), which was initiated during early Oligocene (Jourdan et al., 2013), post-dates the preceding Eocene nappe stacking phase. As emphasised previously, this is consistent with the observed overprint (cross cutting) of the Embrunais basal thrust by the Penninic thrust (fig. 1b, Fig. 2).

Describing the structure of the Western Alps mainly on the basis of the « Penninic thrust » is an oversimplification, because this tectonic boundary crosscuts the initial subduction wedge and developed relatively recently (since early Oligocene onwards; Simon-Labric et al., 2009; Maino et al., 2015), coeval with the formation of the arc (Dumont et al., 2012). Its « out-of-sequence » character can be observed south of Briançon city, where it cuts across an earlier nappe stack involving ocean-derived flysch sediments, the Embrunais-Ubaye nappes (fig. 1b), also represented in the Ligurian and Prealps nappes. Moreover, parts of the oceanic accretionary wedge deformed together with distal European continental margin units are exhumed in its hangingwall (Schmid et al., 2004). It is important to consider that the building of the Western Alps

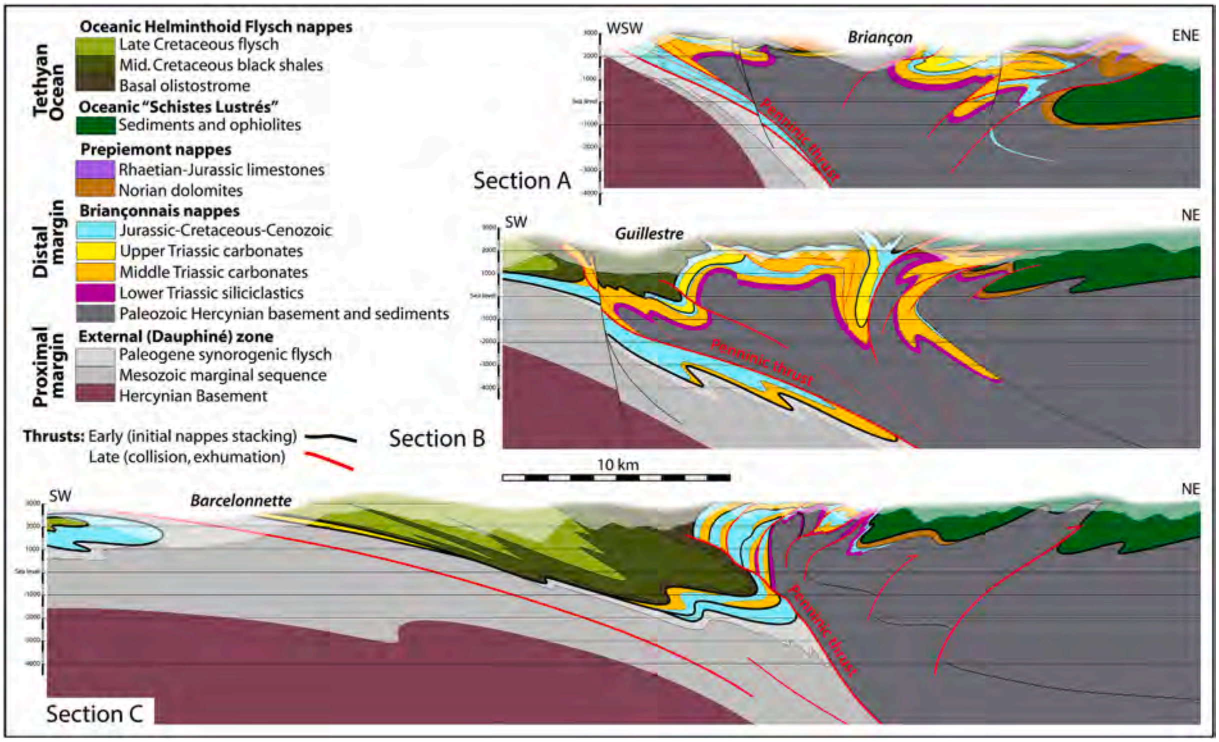

Fig. 2. Radial cross-sections along the Durance (A), Guil (B) and Ubaye (C) valleys, illustrating the double-verging structure of the Briançonnais zone, exhumed in between the Tethyan oceanic nappes (non-metamorphic « Helminthoid Flyschs » westwards, and metamorphic « Schistes Lustrés » eastwards). Such radially oriented sections emphasize the late Alpine structures (red) such as the Penninic Thrust and conjugates backthrusts, which overprint the early structures of the initial nappes stack (black).

arc required polyphase deformation and displacement directions changing through time (Ford et al., 2006) as opposed to a continuous process driven by strain partitioning and maintenance of specifically orientated convergence directions (Fry, 1989).

The structure of the Internal zones of the Western Alps arc underlined by the « Penninic » curved frontal sole thrust and by prominent backthrusting is a possible consequence of Adria mantle indentation, that is particularly evident in the northernmost Western Alps (Schmid et al., 2017; Malusà et al., 2021; Nouibat et al., 2022). Both follow the curvature of the arc. In the study area, radial interpretative cross-sections (fig. 2) illustrate this double-vergent structural style, along with the large-scale characteristics of the Briançonnais nappe stack. Radial sections are commonly used to represent bulk Alpine deformation across the arc (i.e. Schmid et al., 2017) and have been used as the basis for restored models of pre-Alpine paleogeography or shortening estimates (Fry, 1989; Seno et al., 2004; Bellahsen et al., 2014). These existing restorations are not necessarily compatible together, precisely because of their radial distribution over nearly 180°. A key consideration which is often overlooked when examining radial sections is the amount of out-of-plane movement, oblique or perpendicular to cross-section orientations. We provide evidence for complex, non-coaxial deformation histories and geometrically cross-cutting structures, based on examination of numerous field localities.

3. Stratigraphic and structural setting of the study area

Variable stratigraphic characteristics are observed in the Internal nappes, whose outcropping elements are dominantly composed of sedimentary cover. They range from 'pre-rift', rift and starved continental margin sequences, to oceanic sediments and remnants of their slow-spreading oceanic floor (Lemoine et al., 1986; Lagabrielle, 1994). The marginal stratigraphy includes late Paleozoic detrital and volcanoclastic formations which demonstrate the transition from late Variscan foreland basins to Permo-Triassic incipient crustal thinning, Triassic shallow marine, carbonate to evaporitic series, and highly condensed Jurassic to Cretaceous sediments capped by Paleocene to late Eocene flysch. The successive geodynamic settings which have controlled the facies and thickness of these sedimentary sequences are as follows:

- Late Paleozoic: continental clastic and volcaniclastic sequences which may reach 2.5 km in total thickness, were deposited in late Variscan foreland basins in a transtensional setting, in the framework of a major dextral transcurrent shear zone south of the Variscan orogenic belt (Guillot and Ménot, 2009; Von Raumer et al., 2012; Festa et al., 2020). Strong lateral variations indicate the activity of depocenters controlled by extensional faulting (Cortesogno et al., 1993), and widespread Permian volcanic activity, thermal evolution and underplating document the initiation of post-Variscan crustal attenuation in the entire Alpine area (Cortesogno et al., 1998; Rottura et al., 1998; Marotta and Spalla, 2007; Sinigoi et al., 2010).

- Triassic: the classical siliciclastic-carbonate-evaporitic cycle is widely developed in the Internal units of both European and Adria origin, with kilometre-scale thickness. Middle Triassic shallow marine sequences can be traced over the entire Alpine-Carpathian area. In the Western Alps, little evidence for brittle extension has been reported to date despite significant subsidence. More important Triassic rift basins were located further East / Southeast as part of the Neotethyan basin propagation (Stampfli et al., 2002). However, evaporites deposited during the late Triassic played a major role during Alpine orogeny as they controlled the detachment of large sedimentary cover units within the collision wedge.

- Early to Middle Jurassic: all the marginal cover units of the Internal Western Alps show evidence of a rift setting from early Jurassic onwards (Dumont, 1998). Contrasting subsidence patterns during early Liassic suggest tectonic subsidence followed by uplift over the whole Briançonnais domain during late Liassic-middle Jurassic, which produced emergence and continental erosion coeval with extensional block faulting (Claudel and Dumont, 1999). This process, which consists of a long-wavelength uplift with a vertical amplitude reaching about 1 km, could be explained by rift shoulder uplift (Stampfli and Marthaler, 1990). Alternatively, it could result from different processes such as thermal influence linked with the upper mantle boudinage and impregnation during hyperextension (« thermal erosion », Mohn et al., 2012) or extensional ribbon uplift (Tavani et al., 2021). At a shorter wavelength, the maximum amplitude of erosion and uplift is observed in the internal Briançonnais units, initially located close to the paleogeographic boundary with the strongly subsiding Prepiedmont domain. This feature could correspond to flexural uplift (Basile and Allemand, 2002) or to the influence of lithospheric necking (Ribes et al., 2019). The resulting unconformity is a widely recognised characteristic of the Briançonnais marginal plateau, which was emerged and increasingly uplifted towards the incipient Tethyan breakup (Lemoine et al., 1986). The magnitude of the associated erosional gap increases towards the rift, that is from the external to the internal Briançonnais units. Syn-rift continental erosion removed the whole Triassic sequence in the most internal Briançonnais units, allowing the post-rift sediments to rest on the late Variscan volcaniclastics or directly on the basement. Conversely, this unconformity is not recognised in the more proximal rift basins of the External zone, nor closer to the breakup, in the Prepiedmont domain which was fed by turbidites sourced from the emerged Briançonnais shoulder (Dumont et al., 1984). This strongly subsiding domain may represent the most hyperextended part of the distal margin (Mohn et al., 2012) and play a key role to locate the early inversion processes (Tavani et al., 2021).

- late Middle Jurassic to early Cretaceous: consequent to the erosional events described above, a widespread unconformity is spectacularly exposed in the Briançonnais, characterised by deep marine post-rift (syn-spreading) sediments overlying a variety of older (pre-spreading) stratigraphic units. This unconformity has traditionally been described as the "breakup unconformity" (Lemoine et al., 1986; see discussion in Masini et al., 2013). Following the initial breakup in the late Middle to early Late Jurassic (Li et al., 2013, and refs therein), a uniform post-rift series covered the whole margin, including the Briançonnais marginal plateau, preserving a transgressive lag over the erosional surface. These starved post-rift pelagic sediments record thermal subsidence. However, some restricted extensional deformation is reported locally within the Briançonnais area (Claudel and Dumont, 1999), as in Provence (Dardeau et al., 1988), and possibly due to incipient rifting between Europe and areas connected to the Iberian plate.

- Late Cretaceous to Eocene: still in a deep marine setting, several domains of the internal zones record the diachronous onset of flysch sedimentation. The oceanic sediments show evidence of margin sourced turbidites from the Cenomanian onwards (Caron et al., 1989; Durand-Delga et al., 2005; Catanzariti et al., 2007), with increasing clastic input from the Campanian (Helminthoid Flysch fm.) related to the active Adria margin (Di Giulio, 1992; Marroni et al., 1992, 2001). Local sourcing from the European margin (Gottero flysch, Nilsen and Abbate, 1984; Marroni et al., 2010) indicate that denudation occurred before involvement in the Adria-Europe collision, possibly in relation to the onset of Pyrenean orogeny. The continental margin series are also affected by tectono-sedimentary disturbances, for example erosional unconformities and breccias, especially in the Briançonnais units (Gidon et al., 1994). Also recorded in the Alpine foreland (Michard et al., 2010) and in Provence (Espurt et al., 2012), these features can be interpreted either as a response to active transcurrent deformation (Bertok et al., 2012), Pyrenean forebulge propagation (Thum et al., 2015), or Alpine forebulge propagation (Michard and Martinotti, 2002).

Despite the metamorphic overprint, (litho)stratigraphic correlations between many units presently included in the Internal Zones of the western Alpine arc remain possible. These units generally display a sedimentary record markedly different from the External Zone (Dauphiné-Hevetic, Vocontian and Provence domains), both in terms of stratigraphic thickness and environment, from post-Varican to Eocene times (Lemoine et al., 1986). This supports the occurrence of a major lithospheric-scale displacement along the boundary between the Internal and External Zones, whose large paleogeographic areas were likely overthrust from the earliest Oligocene onwards.

As previously stated, the dominant structures following the Western Alps arc (Penninic thrust, Briançonnais zone, metamorphic zonation in Schistes Lustrés, internal cristalline massifs trend) were produced during the westward extrusion stage from the early Oligocene onwards, and they crosscut, deformed and exhumed the initial stack formed during the north- to northwestward Eocene orogenic propagation. Thus, despite this initial phase accommodated important tectonic transports, its structural effects are presently obscured due to further overprint. Some evidence is provided by interference structures at different scales, especially in the Western Alps where the directions of early and late Alpine orogenic propagation are markedly different.

4. Polyphase deformation: large-scale overprint between differently oriented structures in the Internal Zones (interference structures?)

Large-scale superposed deformation due to crossed shortening episodes is suggested in the external zone by the circular shape of the Pelvoux-Ecrins cristalline massif (Dumont et al., 2011). Interference shortening structures are reported from the internal Western Alpine arc (e.g. Jaillard, 1984; Platt and Lister, 1985; Ganne, 2003; Bucher et al., 2004). In the Central Alps, the deeply exhumed Lepontine area shows complex curved shapes which are interpreted to result from superposed deformations with different strain patterns (Merle, 1987). Steck (2008) and Steck et al. (2013, 2015, 2019) document an Eocene-earliest-Oligocene initial stage of N-directed fold-nappe thrusting by ductile detachment of the upper European crust, overprinted by later Oligocene extensional extrusion of the Lepontine dome structure. No occurrences of interference structures are described in the literature from the study area south of the Ambin massif. Here we provide some examples of subperpendicular fold-and-thrust structures in the east of the Briançonnais zone, which are scarce because the older deformation has been largely overprinted by the younger phase which is responsible for the arcuate trend of the modern chain.

4.1. *Superposed fold structures in the internal Briançonnais units of Ubaye valley (a, b, fig. 3)*

In the southern part of the Western Alpine arc, the "Roure" zone (Le Guernic, 1967) and the Acceglio-Longet (AL) zone (Debelmas and Lemoine, 1957; Lefèvre and Michard, 1976) belong to internal Briançonnais. They are composed of late Paleozoic and Mesozoic series, characterized by very thin Triassic-Jurassic sediments compared to the external Briançonnais units. They outcrop in two SSE-NNW oriented strips (a and b fig. 3, respectively) tangential to the trend of the arc, which are formed mainly by ENE-WSW shortening and eastward backfolding. However, in both areas it is possible to detect earlier deformation criteria, roughly perpendicular to the youngest deformation.

At Col du Longet (a, fig. 3), the northern termination of the Acceglio-Longet strip (Schwartz et al., 2000) becomes buried beneath the Rocca Bianca ophiolitic massif and the surrounding metasediments (fig. 4). Outcrop-scale structures consist of a pervasive lineation and metre-scale, top-to-the N to NW overturned folds trending WSW-ENE, both within the Permian to Mesozoic clastic formations of the AL unit and within the oceanic units above (fig. 4; stereogram site 39, see §5; Verly, 2015). These NNW-verging structures postdate the initial stacking of oceanic and continental units, and are clearly deformed by backfolding. The latter occurred close to the Eocene-Oligocene boundary according to U-Th-Pb dating on allanite in this locality (Verly, 2015), implying that oceanic and continental margin units had been stacked earlier during

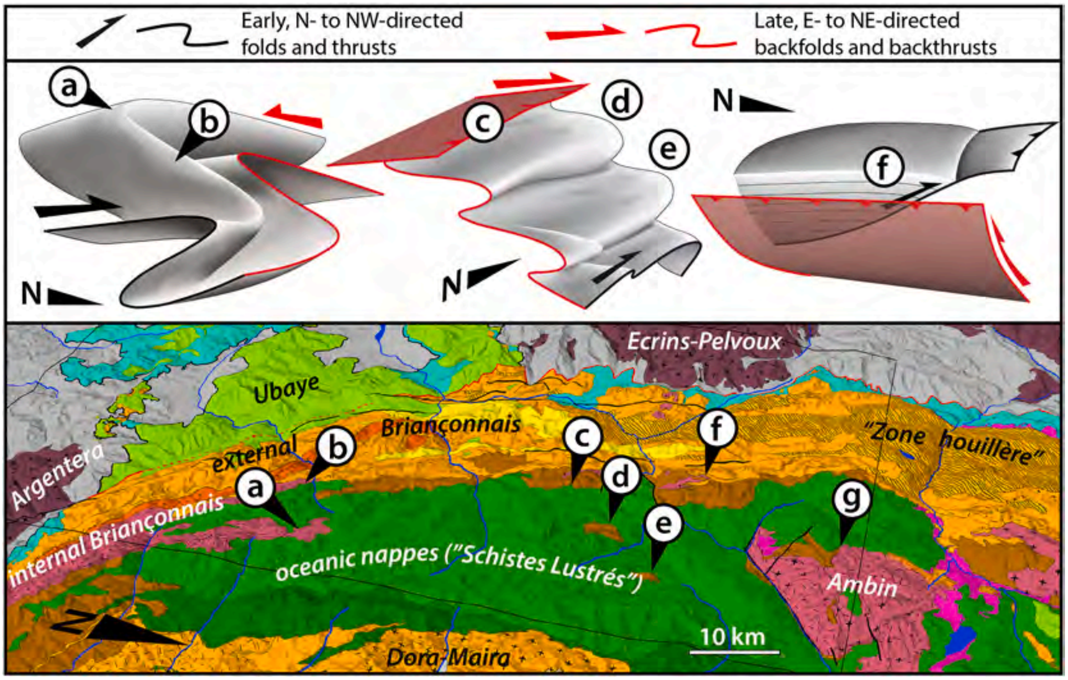

Fig. 3. Schematic situation of the main fold-and-thrust interference structures observed on the internal side of the Briançonnais zone, described in § 4, and their location in the map (perspective viw towards the SW). Early phase: N- to NW-directed; late phase: E- to NE-directed.

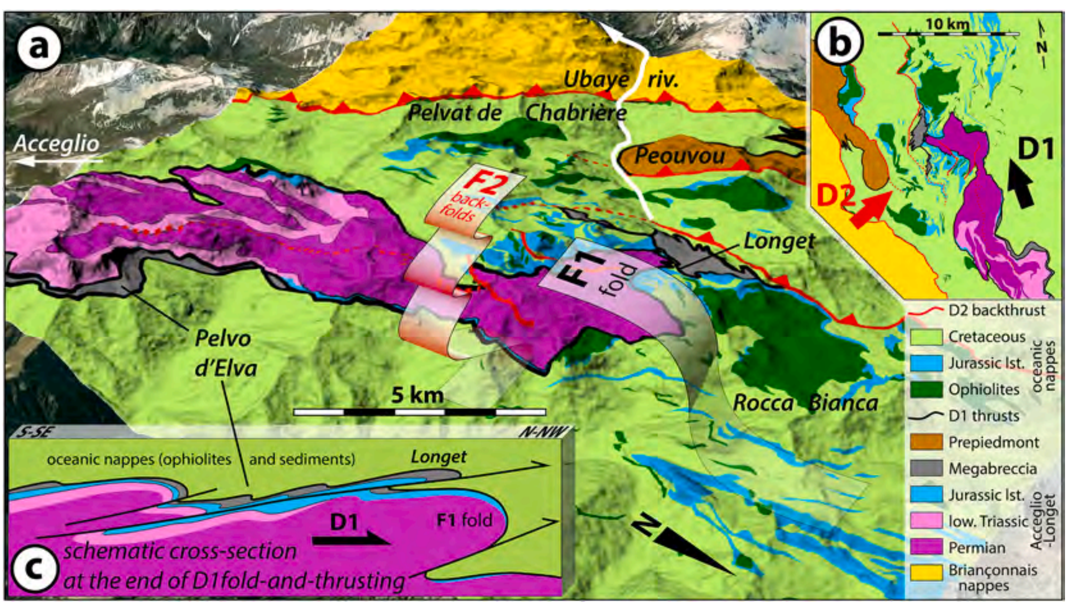

Fig. 4. Structural framework of the northern Acceglio-Longet « ultrabriançonnais » continental margin unit (purple), uplifted from beneath the oceanic Schistes Lustrés nappes (green) to the east of the Briançonnais zone (yellow). Location on fig. 1b. This tectonic window is interpreted as resulting from large-scale interference between early, N-NW directed fold-thrusting (D1, F1) and late E-NE directed backfolding (D2, F2). Note the location of the Pelvo d'Elva and Longet megabreccia slivers (grey) along the boundary between continental margin and oceanic units. a- Perspective view towards the SW of a simplified geological map drapped over DEM. b- Vertical view of the same map with approximative orientation of D1 and D2 deformations. c- Schematic cross-section trending subparallel to D1 tectonic transport, reconstructed before D2 backfolding, and able to explain the northward interruption of the Acceglio-Longet outcrops beneath the Rocca Bianca massif.

the Eocene. We propose that this termination of the Acceglio-Longet strip corresponds to a large-scale hinge of a transverse, recumbent or isoclinal fold trending SW-NE to EW (section c, fig. 4), which was further refolded by NE-ward recumbent backfolding, similarly to the conceptual sketch of fig. 3 (a, upper left cartoon).

Different types of breccia are reported in the Col du Longet area (Lemoine, 1967; Gout, 1987 and refs therein). The significance of these breccia is discussed in §8, but their interpretation and their assignment to either continental or oceanic units is dependant on the structural setting. The two main end-members are (i) thin beds of siliciclastic microbreccias overlying the Permian siliciclastic and volcaniclastic formations of the AL unit, with minor occurrence of dolomitic clasts indicating a post-Triassic age. (ii) mixed siliciclastic/carbonate megabreccia containing metric to decametric blocks of Triassic dolostones together with reworked Permian clasts of various size. They occur as lenses or slices above the Acceglio-Longet unit or within the base of the

Schistes Lustrés oceanic units (grey, fig. 4). The first type (i) is stratigraphically linked with the AL unit, and must be Jurassic to early Cretaceous in age because it is overlain by a Late Cretaceous hardground (Lemoine, 1960a). The second type (ii) is associated with late Jurassic-Cretaceous, strongly folded marbles and calcschists whose continental or oceanic origin is hardly distinguishable (Gout, 1987). However, the megabreccia does not appear to contain any ophiolitic clasts and is dominated by reworking of Permian or Mesozoic formations found in the AL series. This megabreccia is comparable to the Pelvo d'Elva breccia (Michard, 1967; Lefèvre and Michard, 1976) which occurs 10 km to the SE in a cover thrust sheet pinched in between the AL unit and the oceanic units (fig. 4) and which is also sourced from continental margin rocks only (Michard, pers. comm. 2016). The Longet megabreccia occurs within the NW equivalent of the Pelvo d'Elva thrust sheet (c, fig. 4), and its present situation within the Schistes Lustrés can be explained by interference between N to NW directed tectonic transport, isoclinal folding, and further backfolding. The Longet and Pelvo d'Elva breccia occurs presently in the normal and reverse limbs of the AL backfold.

9 km SW of Col du Longet, near Maljasset, the Ubaye river crosscuts the tectonic boundary between the continental margin (Roure) and oceanic (Schistes Lustrés) units (b, fig. 3). This contact is presently dipping southwestwards and the most internal Briançonnais nappe (Roure) is overriding the oceanic units due to backfolding. The asymmetry of the minor folds associated with the initial stacking indicate an apparent southward transport (a, fig. 5, present structure), which is inconsistent with the other regional data. However, taking into account the overall reversal and once restored from backfolding, it becomes N-NW directed (b, fig. 5, restored structure), similarly as in Col du Longet (see §5, site 39).

4.2. Backfold and backthrust overprint structures in the Piemont units east of Briançon (c, d, e, fig. 3)

The Piemont units of Rochebrune (c, fig. 3) and Chaberton are thrust eastwards over the oceanic blueschist units, composed of ophiolites and Mesozoic metasediments (Dumont et al., 1984). This backward thrusting crosscuts an initial stack characterised by oceanic units resting in thrust contact with underlying Prepiedmont units, a situation visible beneath the Chenaillet ophiolitic massif (Gimont tectonic window; Barféty et al., 1995). Some ophiolite bearing slices found along the backthrust beneath the Rochebrune and Chaberton units represent remnants of the sheared backfold. Further east, the Piemont units outcrop in two subcircular tectonic windows surrounded by the oceanic metasediments with rare ophiolitic rocks, the Gran Roc and the Mte Banchetta windows (d, e, fig. 3, respectively). These continental margin units suffered deeper burial than those of the Rochebrune and Chaberton units (Caron, 1971), and we interpret their high elevation (~3000m and ~2800m, respectively) as dome structures resulting from large-scale interference folding similar to the conceptual sketch of fig. 3 (upper sketch, centre). In both cases, these structures are markedly asymmetric, their eastern limb being steeper or overturned.

The Gran-Roc window (d, fig. 3) is composed of, from bottom to top, Carnian gypsum (Megard-Galli and Caron, 1972) which is the detachment layer, a thick pile of Norian dolostones (800m) with dolomitic breccia at their base (Caron, 1971), and Rhaetian to Liassic limestones and shales typical of the Prepiedmont series (Dumont, 1983). The western side of the massif provides a natural N-S cross section showing an asymmetric anticline cored by the Carnian evaporites. This structure may have been initiated as a north-verging ramp anticline (fig. 6). It is furthermore possible that the location of this fold was initially localised by a pre-existing diapir, because of the apparent truncation of dolostones by the underlying evaporites.

Further east, the Mte Banchetta tectonic window (e, fig. 3; fig. 7) is also cored by a thick pile of Triassic dolostones, locally covered by alternating carbonate and shale which have Rhaetian characteristics (site 13, §5, and eastern slopes of Mte Banchetta). A calcschist formation similar to the "Lias prépiémontais" (Dumont, 1984) is exposed to the SW of Mte Banchetta (b, fig. 7; Jouvent, 2017). The occurrence of these three diagnostic lithologies indicates that the Mte Banchetta tectonic window is underlain by a Prepiedmont Mesozoic series similarly to the nearby Gran Roc window.

A 'mixed' megabreccia occurs between these continental margin formations and the oceanic units, first described by Caron (1971), whose significance will be discussed in § 8.3. The 3D map-scale geometry of the Mte Banchetta window is consistent with an interference structure (fig. 3, e). Similarly to the Gran Roc window, the eastern slopes show obvious backward fold-and-thrusting of oceanic and continental unit imbricates (F2, fig. 7), a feature comparable to large-scale backfolding and backthrusting at the eastern edge of the Rochebrune and Chaberton-Grande Hoche Prepiedmont units (Dumont et al., 1984; Tricart and Sue, 2006). To the N-NW of Mte Banchetta, the Prepiedmont unit is buried beneath

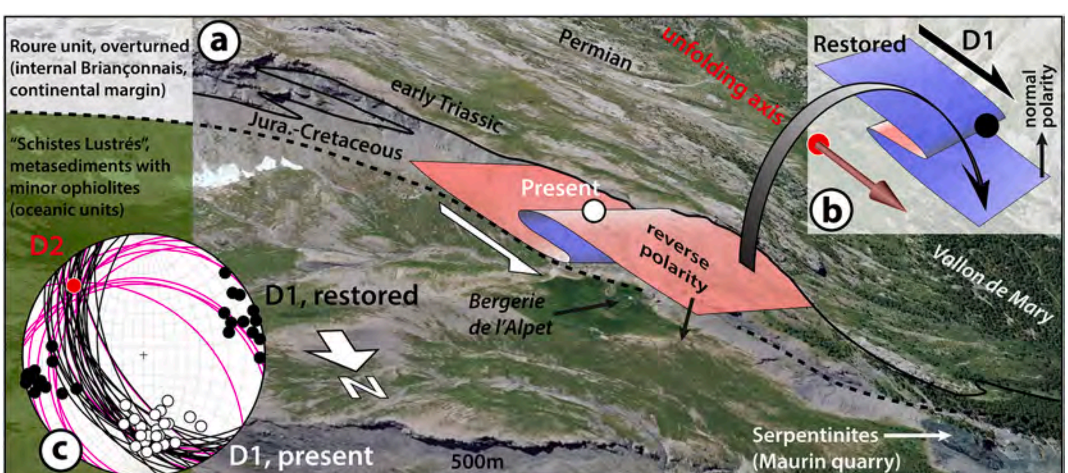

Fig. 5. Fold interference in the Alpet area, southern side of the Ubaye valley (location fig. 1b). a- Synthetic aerial view towards SW of the backfolded eastern limit of the Briançonnais zone, bounding the most internal Briançonnais unit (Roure unit; Le Guernic, 1967) above, from the serpentinite-bearing oceanic Schistes Lustrés nappes below. The contact in reverse polarity is affected by a medium scale isoclinal F1 fold (red) displaying an apparent top-to-the-south asymmetry. b- Once restored from F2 backfolding (red unfolding axis), the F1 asymmetry gets N- to NW-directed, consistently with our other regional observations. c- Stereogram (Wulff, lower hemisphere) showing the unfolding axis (D2 fold), with present (white) and unfolded (black) attitude of the D1 microstructures (fold axis and intersection lineations). Site n°44, location Table 1.

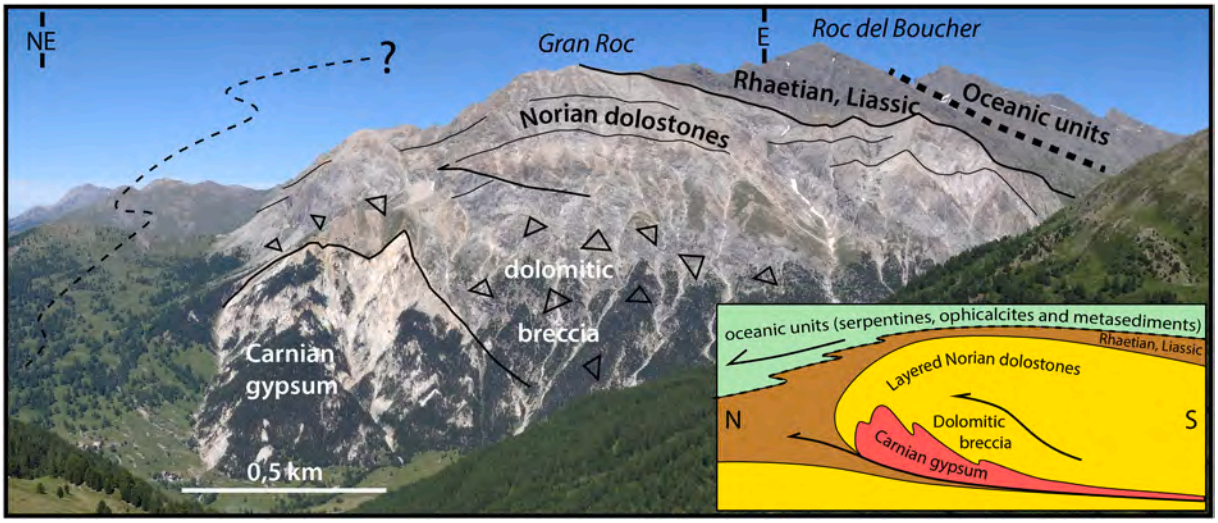

Fig. 6. Western slopes of the Gran-Roc massif, a tectonic window cored by a Prepiedmont upper Triassic-lowermost Jurassic series and surrounded by oceanic Schistes Lustrés nappes (location fig. 1b). It shows asymmetric anticlinal bending of the upper dolomitic layers, likely associated with a NE-directed ramp over the Carnian evaporites detachment layer. We interpret the interruption of the Prepiedmont unit towards the left (NE) as a result of hangingwall cutoff by a D1 NE-directed thrust (cartoon). The eastern side of the massif, behind the crest (visible on fig. 7), is affected by D2 backfolding, producing the rounded shape of this tectonic window.

the oceanic metasediments with no evidence of any normal fault. We explain this structure by the occurrence of a N- to NW-overturned anticline involving both the continental and oceanic units and interfering with backfolds to produce the circular-shape window. Small scale folds consistent with this interpretation are observed (sites 13 and 14, § 5).

4.3. *Hanging-wall cutoff in the Piemont nappes (f, fig. 3)*

To the East of Briançon city, the Montgenèvre pass is located on a Piemont nappe which is thrust over the Briançonnais stack (a, fig. 8). This Piemont nappe is detached beneath the Norian dolostones. The complete thickness of the Norian formation (800m) is present in the Janus massif to the S (b, fig. 8), but it decreases progressively northward to ~300m (Chalvet) as a result of basal truncation by a thrust ramp, and the hangingwall cut-off reaches the Rhaetian and the Jurassic fm. further to the north (fig. 8). The orientation of this cut-off outcropping on both sides of the Chalvet ridge is ENE-WSW. The deformation of Liassic strata in the hangingwall (Col de la Lauze, e, fig. 8) consists of ~N80°, top-to-the N trending folds, whereas stretching lineations in the footwall (Clot Enjaime, d fig. 8) are trending ~N170°. Thus, the large-scale structure and the outcrop-scale deformation consistently indicate a top N to NW tectonic transport of the Piemont units over Internal Briançonnais units of Clot Enjaime-Alpet-Rio Secco, which are in turn resting structurally on the classical Briançonnais footwall. This relatively early thrust stacking is overprinted by both backfolding/backthrusting, duplicating the Piemont units in this area (Chalvet and Chaberton units, separated by the NS trending Rio Secco syncline), and by recent extensional faulting (b, fig. 8).

The Chaberton unit itself terminates northwards, possibly due to a similar hangingwall cut-off truncation of the Norian dolostones near the Acles pass (site 7, § 5). Here the overlying Rhaetian-Liassic strata are affected by ~EW trending deformation similar to the area of Col de La Lauze.

4.4. *Tectonic cover of the Ambin massif (g, fig. 3)*

The Ambin internal crystalline massif is a 20 km-wide dome surrounded by the "Schistes lustrés" oceanic units (fig. 1, fig. 9). It has a Variscan metamorphic crystalline core showing evidence of north-vergent HP early Alpine deformation overprinted by retromorphic E-directed shear (Clarea nappe, Ganne et al., 2004 and refs. therein). This core is covered by a nappe stack including, from bottom to top, late Paleozoic volcaniclastics ("Ambin group", or Ambin nappe), Mesozoic continental derived units (Briançonnais, Piemont), and oceanic units (Schistes Lustrés). The Mesozoic cover nappes pinch-out northwards over the Ambin nappe on top of the dome, as shown in a N-S cross section (fig. 9a, 9c; Jouvent, 2017). The culmination coincides with a shortened fault block with locally preserved Triassic strata, truncated by the cover nappes. The lower Triassic quartzites in the footwall of the nappes display evidence of north-directed shear (stratification/cleavage relationships, fig. 9d; site 3, § 5). The top of the Ambin nappe displays a regional angular unconformity, with northward-younging middle Triassic carbonate strata beneath it, towards a more complete Briançonnais series (Bellecombe area, Jaillard, 1989). In contrast, towards the south, an extremely reduced Mesozoic cover overlies the lowest Triassic to late Paleozoic clastics (Polino et al., 2002), as usually observed in the Internal Briançonnais domains. This regional unconformity together with the fault block (fig. 9a, 9c) suggest that the location of the Ambin dome is partly inherited from the Mesozoic passive margin history, and that the northward propagation of the early Alpine nappes reactivate a transverse (~EW) uplifted rift structure.

5. Outcrop-scale structural record of multistage kinematics in the Penninic nappes of the Western Alps

5.1. *Literature review*

A synthetic outcrop-scale analysis in the southwestern part of the western Alpine arc (Tricart, 1980) demonstrated the occurrence of three deformation episodes in the Briançonnais zone. The second and the third phases (D2, D3) correspond respectively to outward thrusting of the nappe stack associated with the activation of the Penninic thrust, and to inward (backward) fold-and-thrusting. Both are kinematically linked with the formation and the propagation of the Western Alps arc, initiated approximately at the Eocene-Oligocene boundary (Dumont et al., 2011). D2 and D3 sensu Tricart (1980) are overlapping in time because they occurred on both sides of the "Briançonnais fan" whose axis propagated outwards. D1 corresponds to the initial underthrusting of the Briançonnais and Prepiedmont domains beneath the Ligurian-Piemont

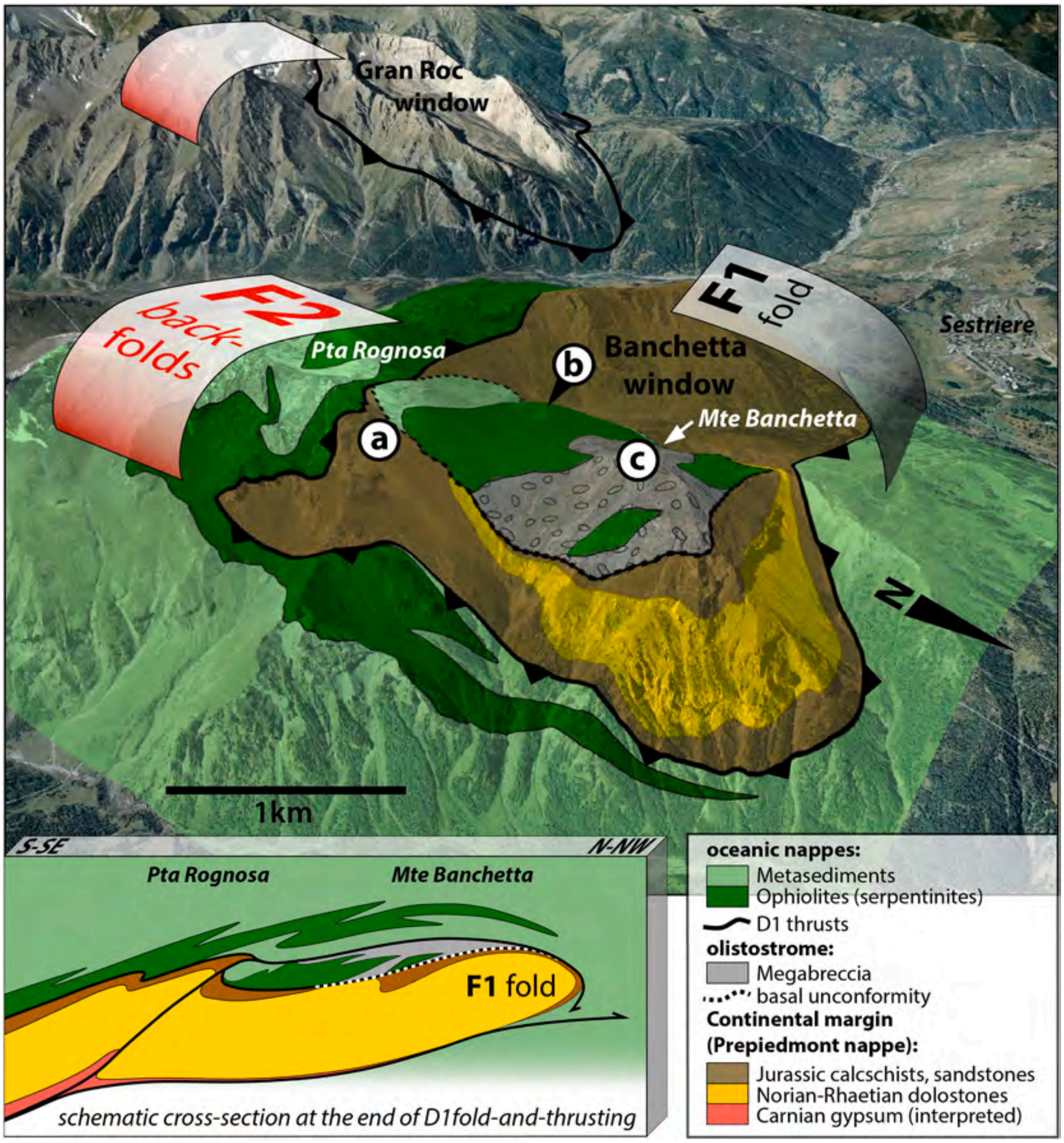

Fig. 7. Mte Banchetta interference structure, near Sestriere ski resort (location fig. 1b): the continental margin series (Prepiedmont nappe) is overlain by a mixed megabreccia described in fig. 14, and outcrop in a dome-shape tectonic window surrounded by the oceanic Schistes Lustrés nappes.
Upper part: Simplfied geological map and satellite view drapped over DEM, perspective view towards SW.
Lower part: schematic cross-section trending subparallel to D1 tectonic transport, reconstructed before D2 backfolding, illustrating our interpretation of the northward interruption of Prepiedmont outcrops as a result of top-to-the north hangingwall cutoff by a D1 N-NW directed blind thrust, similarly as the Gran-Roc structure (fig. 6).

oceanic accretionary wedge. It was previously regarded as late Cretaceous (Caron, 1977; Tricart, 1980). A late Cretaceous age was assigned to the Eoalpine orogeny (85–60 My; Hunziker et al., 1992, and refs therein; Dal Piaz and Lombardo, 1985; Polino et al., 1990) based on K-Ar dating, which is now recognised to have involved substantial uncertainty and scattering due to excess argon analytical problems. HP radiometric ages were still controversial in the 1980's (Lemoine et al., 1984) but were progressively refined to Cenozoic (65–38 Ma; Liewig et al., 1981; Takeshita et al., 1994). From a structural point of view, the discrepancies in the orientation of early small-scale structures with respect to younger deformation phases was reported by several authors (Vialon, 1966; Caron, 1973, 1974; Caby, 1973; Malavieille, 1982; Mahwin et al., 1983; Platt et al., 1989). This well-known feature has been interpreted in different ways, namely as Variscan inheritance (Vialon, 1966), resulting from shear coeval with block rotation (Caron, 1974), as a rotation of early structures (Boudon et al., 1976), or as change in deformation regime (Malavieille, 1982). Conversely, Caby (1973, 1975) interpreted the transverse trends of the early Alpine deformation phase to result from northward translation of the orogenic wedge, having preceded the development of the arc curvature. A northward translation of the early Alpine wedge was also promoted by Maury and Ricou (1983) and Ricou and Siddans (1986), and later documented by Schmid and Kissling (2000) and Ceriani et al. (2001). The latter publication reported initial N to NW-directed tectonic transport in the Internal nappe stack using their microstructural signature in the vicinity of the Penninic thrust.

5.2. Field data and structural interpretation

Here we present a comprehensive field survey of small-scale superposed fold structures which affect various stratigraphic layers of the late Paleozoic-Mesozoic series, over >60km along the southern part of the Internal arc of the Western Alps (fig. 10; fig. 11; table 1). Most of the 55 sites are located in the Briançonnais zone, and some are within the

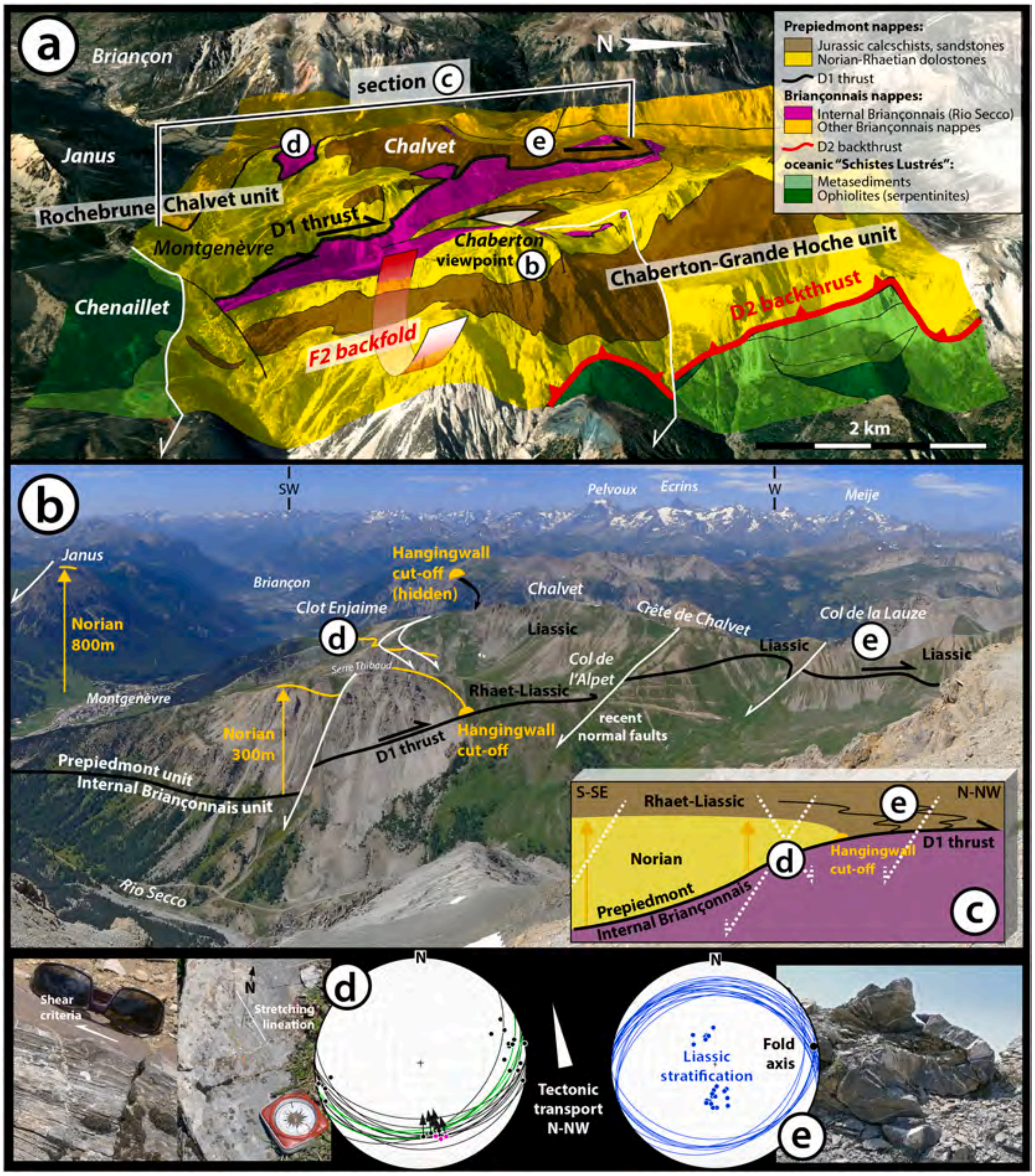

Fig. 8. Fold-thrust interference structures in the Chaberton and Chalvet areas, north of Montgenèvre pass (location fig. 1b). This area shows Prepiedmont series displaying complexe tectonic relationships with both the Briançonnais continental margin units and the Shistes Lustrés oceanic nappes, due to interaction between two subperpendicular fold-thrust deformations D1 and D2. The southern (left) part of the block-diagram shows the Chenaillet ophiolitic massif downthrown along a major late-orogenic normal fault (Tricart and Sue, 2006). a- Simplified geological map and satellite view drapped over the DEM, perspective view towards W. The Chaberton-Grande Hoche massif is backthrust over the oceanic Schistes Lustrés units (green) and affected by D2 backfolding. By contrast, the Chalvet massif preserves more or less the initial D1 stack since the Prepiedmont unit overlains the Internal Briançonnais unit of Rio Secco (§ 4.3) through a D1 thrust. b- Panoramic view of the Chalvet area towards the W-SW, from the top of Chaberton, showing the ramp geometry of this thrust, with basal truncation and hangingwall cutoff of the Prepiedmont Norian dolostones from S to N (right). c- Schematic cross-section trending subparallel to D1 tectonic transport, more or less corresponding to the present outcrops with the exception of late orogenic normal faulting (white). d and e- Kinematic evidence for N-directed transport along the D1 thrust at the bottom of the Prepiedmont Chalvet unit (folds, cleavage and stretching lineation; Wulff stereograms, lower hemisphere). Location in figs. a to c above (respectively sites 12 and 11, fig. 10, table 1).

oceanic units on both sides of it. The deformation criteria measured are outcrop-scale axes, intersection lineations and stretching lineations. At each site, we identified the pro- and/or retro-verging deformations whose trends are broadly following the shape of the arc, and we restored (unfolded) the preceding tectonic features which generally consist of folds and stratification/schistosity intersection lineations. Compared to the chronology of Tricart (1980), the partly overlapping outward and inward deformation episodes referred to as D2 and D3, respectively, are merged in D2 in this study, which aims to focus on the early part of deformation history. D2 outward or inward structures are used to restore the previous kinematic indicators, assigned to D1 deformation.

As an example, a description of site 48, located in the External Briançonnais, is provided in fig. 12 (localisation fig. 10 & table 1). Here, the initial contact between the overthrust oceanic sediments, namely the Helminthoid Flysch nappe composed of early late Cretaceous purple shales, and an underlying continental margin Briançonnais unit (Sautron nappe, Gidon et al., 1994) is observed. This contact is affected by top-to-the SW shear deformation, folding and thrusting, in the footwall of the Brec de Chambeyron Briançonnais unit (a, fig. 12). SW-verging deformation (D2, fig. 12) is disharmonic, characterised by large

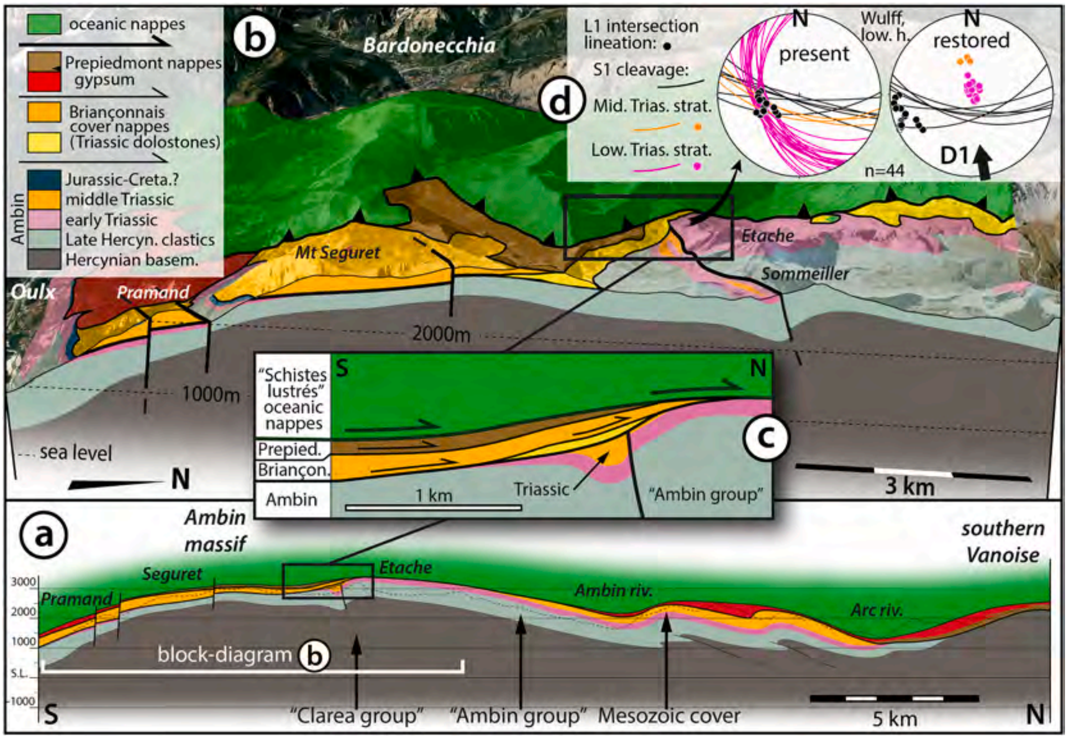

Fig. 9. Nappes structure and tectonic marks on top of the Ambin dome-shape massif, east of Bardonecchia city (location fig. 1b), cored by Variscan basement and late Variscan clastics (Clarea Group and Ambin Group, respectively; Ganne et al., 2004). a- Schematic section across Ambin and the Maurienne valley, illustrating on top of the dome the tectonically reduced character of both the Mezozoic sedimentary cover and the tectonic cover by the Briançonnais-Prepiedmont nappes beneath the oceanic Schistes Lustrés nappes (green).
b- Block diagram combining the southern part of section a with relief, perspective view towards W, with location of critical observations to the W of section a (rectangle, section c below). c- Detailed section on top of the Ambin dome, showing scarse remnants of the Triassic dolomitic cover of the Ambin Group, preserved beneath a northward pinching stack of Briançonnais and Prepiedmont nappes overlain by the Schistes Lustrés. This preservation probably occurred, in spite of intense northward shear, thanks to a pre-orogenic normal fault. d- Kinematic evidence for N-directed tectonic transport in the Triassic cover beneath the nappes (D1 intersection lineation, present attitude and restored from wesward folding; site 3, fig. 10, table 1).

wavelength folds in the Triassic dolomitic limestones of the upper unit, and by tighter folding which displays shorter wavelength in the Cretaceous-Paleogene calcschists (b, fig. 12). These 'D2' folds at different scales trend parallel to the Briançonnais zone in this part of the arc (fig. 10). They overprint an older lineation, trending approximately perpendicular to D2 fold trends, which is systematically observed in the calcschists (c, fig. 12) beneath the oceanic nappe. This lineation consists of intersection between stratification and a pre-D2 schistosity named D1, because it is affected by D2 folds (c, fig. 12, and stereogram). Schistosity S1 is subparallel to stratification, and isoclinal D1 microfolds trending parallel to D1 lineation are also observed. In order to provide a map synthesis of similar observations across the study area (fig. 10; fig. 11), we projected horizontally at each site the mean orientation of D2 fold trends, which may be forward or backward oriented (red or green, respectively), and the mean orientation of D1 lineations or fold axis, once unfolded from D2 deformations.

The map shows a striking consistency of restored D1 trends across the southern part of the Alpine arc. Top-to-the north shear sense is observed in many sites, either indicated by fold asymmetry or by angular relationships between stratification and cleavage. The occurrence of fold trends transverse to the western Alpine chain has long been recognised (Vialon, 1966; Bertrand, 1968 and refs therein; Caby, 1973, 1975; Tricart and Schwartz, 2006), but has been differently interpreted: either as an evidence of N-S shortening and N-directed transport (Caby, 1973; Tricart and Schwartz, 2006), consistent with N-S HP transport identified through high-pressure stretching lineations in the internal crystalline massifs (Choukroune et al., 1986; Ganne et al., 2004; Le Bayon and Ballèvre, 2006; Strzerzynski et al., 2011; Scheiber et al., 2013), or as more or less parallel to tectonic transport, thus oriented radially towards the exterior of the arc (Caron, 1974; Malavieille and Etchecopar, 1981; Malavieille et al., 1984; Philippot, 1990). We argue that these transverse trends can be regarded as approximately perpendicular to the 'D1' transport directions of nappes during early stacking stages, like 'D2' trends are for the late stages.

The D1 structures are regionally consistent, and are locally observed immediately beneath the base of the oceanic nappes (i.e. fig. 10, sites 2, 3, 7, 13, 20, 48). Thus this widespread deformation is most likely associated with the early Alpine collision and with the emplacement of the oceanic accretionary wedge over the distal continental margin units. Like in Provence, the Briançonnais domain must have been affected by older compressional deformation linked with the Pyrenean orogeny, but there is no evidence of deep burial older than early Eocene and these surficial structures were probably largely overprinted by the D1 deformation.

We document the superposition of two deformation stages which can be identified by crossing small-scale structures over the whole area, regardless of the metamorphic grade, from blueschist facies in eastern domains to sub-greenschist facies in the Embrunais-Ubaye areas. Moreover, we observe similar superposed deformations at different scales, from kilometric to outcrop. Many examples of large-scale interference folds have been reported in the Internal Westen Alps (Platt et al., 1989a; Ganne, 2003; Le Bayon, 2005). These observations preclude models involving radial outwards orientations of tectonic transport constant through time, which in addition would make restorations impossible in the core of the arc. We propose that the widespread "transverse" D1 deformation trends result from an early N to NW directed transport and stacking stage, whose propagation appears presently "longitudinal" to the southern part of the Western Alpine arc.

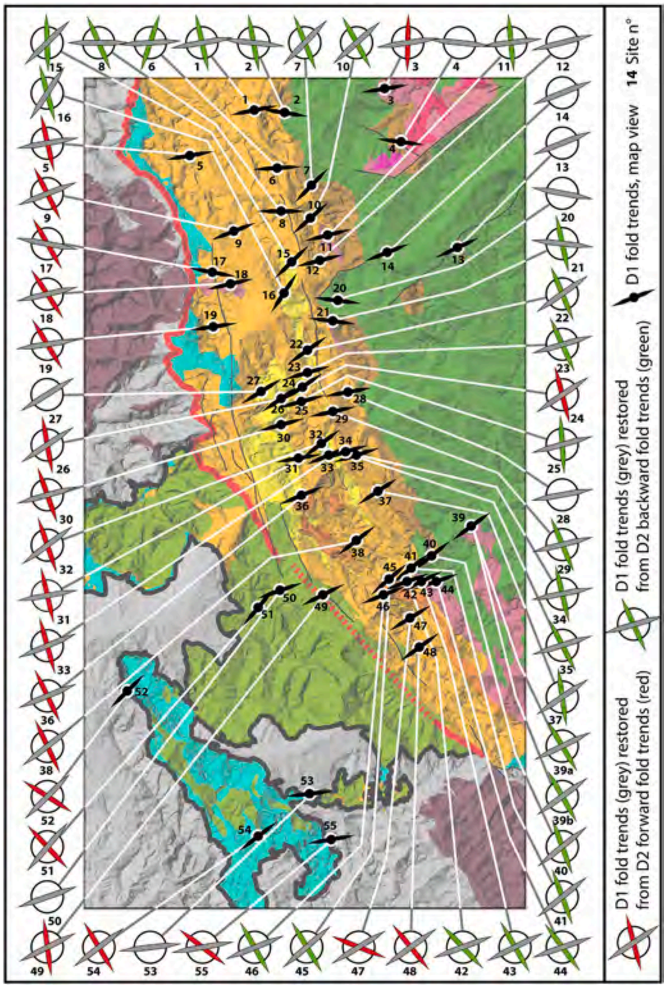

Fig. 10. Geological map taken from fig. 1b, with location of the 55 microstructural sites given in fig. 11 (locations table 1). Over the map are given the orientation of the horizontal projections of early deformation trends (D1) restored from younger overprint, and in the peripheral area the orientation of the late deformation trends (D2), either forward (red) or backward (green) vergent, used to restore the D1 trends. The D2 structures are dominantly W- to SW-ward verging in the WSW part of the Briançonnais zone and in the Embrunais-Ubaye nappes, and E- to NE-ward verging in the ENE part and in the oceanic Schistes Lustrés, demonstrating that the double-verging structure of the Briançonnais was acquired during D2 stage (fig. 2). Note the consistency of D1 indicators, which are trending subperpendicular to the arc shape and varying from E-W in the northern part to NE-SW in the southern part, suggesting that D1 structures predate the formation of the arc.

The restored D1 trends (fig. 10) show a slight discrepancy in orientations from N (Maurienne vally) to S (Ubaye valley), which suggests that these trends were distorted during the formation of the arc. Considering that the formation of the arc is associated with the activation of the post-nappe Penninic thrust (Dumont et al., 2011), and with the pro- and retro-D2 structures, such a distortion is expected from older (D1) structures (Caby, 1973; Bertrand et al., 1996).

Following our interpretation derived from small-scale observations, the most important nappe displacements are expected to have occurred during the early D1 stacking stage, whereas D2 thrusts mostly consist of high-angle, post-nappe structures having accommodated vertical extrusion on both sides of the Briançonnais zone (fig. 2). If so, the internal zones in the Western Alpine arc should display significant along-strike changes as relicts of early stacking process, which is exemplified in the following.

6. Along-strike variations in the structure and internal composition of nappes

Our structural arguments presented in the previous sections show that the D1 early nappe structures are crosscut by D2 folds an thrusts, which follow the shape of the arc, both at an outcrop scale and at a km scale. Since the formation of the arc is a recent feature, similar oblique crosscuting relationships should be observed at a map scale as well. Such obliquity, which may also result from paleogeographic inheritance, can be documented by variations along the strike of the major D2 structures, that is along the different zones following the arc from S to N. To illustrate this, we provide examples from surface geology which demonstrate that the structure of the Western Alpine arc is not concentric in map view. Along-strike variations can be detected within:

6.1. The stratigraphy and structure of the "zone houillère" in the central part of the arc

The "Zone houillère" around the city of Briançon and further north is composed of detrital sedimentary sequences reaching a total thickness of about 2,5 km (Fabre, 1961; Feys, 1963). It shows two superposed main units separated by the Drayères thrust oriented SW-NE (Fabre et al., 1982; Caby, 1996; Barféty et al., 2006; Lanari et al., 2012). The lower unit is found to the N-NE of the study area, from southern Vanoise massif to northern Briançonnais, and includes upper Westphalian (C and D members) to lower Stephanian clastic sequences (Tarentaise fm., Fabre, 1961; Schade et al., 1985) overlain by a thick volcano-sedimentary lower-middle Permian series (Ponsonnière area). The upper unit, dominantly outcropping to the S-SE, near Briançon and further south, contains fining upwards Namurian (south of Briançon) to lower Westphalian (A member) clastics and coal measures, unconformably overlain by middle Stephanian and/or upper Permian coarse clastics (Barféty et al., 1995). These two units were possibly stacked during the latest Variscan orogenic events because both are unconformably overlain by the western Briançonnais Mesozoic cover in the Cerces-Grand Area area (Barféty et al., 2006). They have been subsequently affected by Alpine folds and dominantly east-verging thrusts (Fabre et al., 1982) so that the upper unit is split into several slices. However, it is quite clear that this upper unit, found in the southern area, is less complete than the lower one further north, due to an erosional gap of upper Westphalian-lower Stephanian layers unconformably overlain by middle Stephanian clastics. This southward increasing truncation of the infill of the late Carboniferous basin, also illustrated by Mercier and Beaudouin (1987) and Desmons and Mercier (1993), shows that its southern margin has been deformed and uplifted in relation to late Variscan events. Consistently, a higher late Carboniferous subsidence rate affected the northern part of the Zone Houillère (Manzotti et al., 2014a). The observed fluvial drainage patterns in late Carboniferous formations are dominantly northward directed (Mercier and Beaudouin, 1987; Barféty et al., 1995) and braided systems developed in the vicinity of source areas located to the south (Manzotti et al., 2014a). Thus, in spite of the dominantly N-S trend and the narrow width of the presently outcropping Zone Houillère, there are significant latitudinal changes in its stratigraphy and structure, suggesting that the initial trend of the late Variscan basin was markedly oblique with respect to later Alpine shortening.

6.2. The "Permo-Carboniferous axial zone" in the southern part of the arc

The southernmost constraints on the Carboniferous series are found in a tectonic window across the Ubaye valley, within the lowermost nappe of the external Briançonnais stack (Gidon et al., 1994), structurally equivalent to the « zone houillère » unit near Briançon. Further SE, the Paleozoic formations are only represented by thick Permian volcanics and volcaniclastics, outcropping in both the southeastern extension of this stack, named « Permo-Carboniferous Axial zone » (PCAZ) by Lefèvre (1982), and in the more internal « Roure-Acceglio » zone (RAZ).

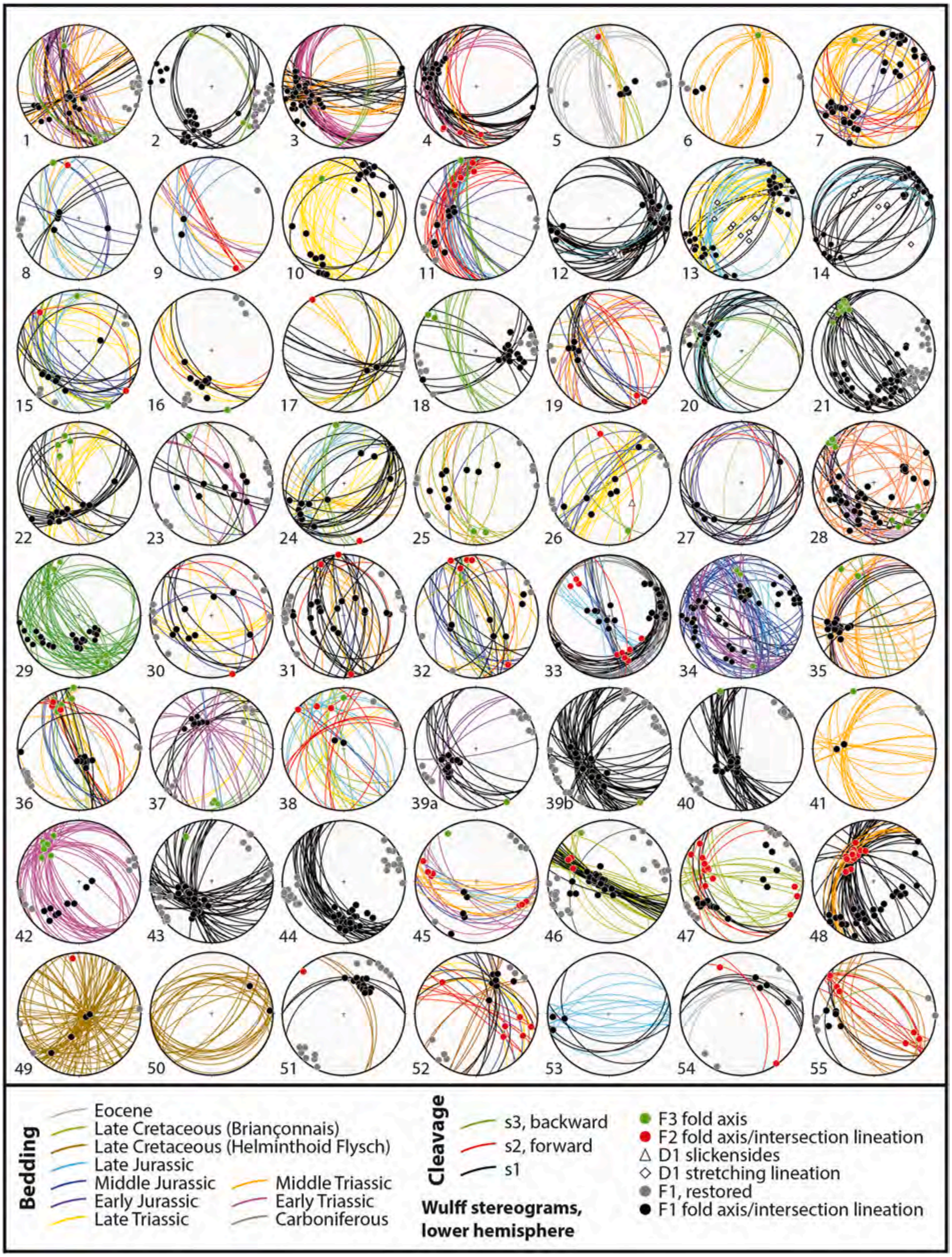

Fig. 11. Microstructural data of the 55 sites (location table 1) investigated in the Briançonnais/Prepiedmont zones and nearby in the Embrunais-Ubaye nappes and the Schistes Lustrés. Red and green dots represent the forward and backward late (D2) fold trends, respectively. Black dots are the early (D1) trends, and grey dots represent the latter restored from D2 deformation (where needed). The median trends at each site are reported on fig. 10.

Table 1
Location of the microstructural data sites listed in figs. 10 and 11.

Site N°	Locality	Latitude	Longitude
1	Col de la Vallée Etroite	45°06'50"	6°36'30"
2	Valle di Rho, NW Bardonecchia	45°06'15"	6°38'42"
3	Valle di Rochemolles, NE Bardonecchia	45°07'35"	6°48'20"
4	Mte Seguret, N Oulx	45°04'30"	6°50'21"
5	Lac des Beraudes, Cerces massif	45°03'34"	6°30'10"
6	l'Aiguille Rouge, NE Névache	45°02'43"	6°38'35"
7	Col des Acles, E Névache	45°01'12"	6°41'45"
8	Forts de l'Olive---Lenlon, S Névache	44°59'10"	6°38'22"
9	Tête Noire, NE Monêtier	44°59'30"	6°32'50"
10	Pointe de Pécé, ESE Névache	44°59'36"	6°41'25"
11	Rio Secco, Col de la Lauze, N Montgenèvre	44°58'00"	6°42'47"
12	Clot Enjaime, NW Montgenèvre	44°56'15"	6°42'18"
13	Mte Banchetta, SE Sestriere	44°56'59"	6°55'20"
14	Champlas Seguin, W Sestriere	44°56'38"	6°48'32"
15	W. slopes of Clarée, N La Vachette	44°55'39"	6°40'06"
16	Fort des Salettes, Briançon	44°54'16"	6°38'54"
17	N. Serre Chevalier, WNW Briançon	44°54'44"	6°32'44"
18	S. Serre Chevalier, WNW Briançon	44°54'22"	6°32'56"
19	Tenailles de Montbrison, N l'Argentière	44°51'09"	6°32'47"
20	Fort du Gondran, Chenaillet, SE Briançon	44°53'32"	6°43'16"
21	Cervières, SE Briançon	44°52'26"	6°43'31"
22	W. Crête des Granges, S Briançon	44°50'15"	6°41'00"
23	Col des Ayes---Beaudouis, S Briançon	44°48'47"	6°41'21"
24	Chalets de Clapeyto, NW Arvieux	44°47'48"	6°40'50"
25	Pic de Balart, NW Arvieux	44°47'03"	6°40'58"
26	Crête de la Moulière, E l'Argentière	44°47'03"	6°39'00"
27	Pic du Bonhomme, E l'Argentière	44°47'30"	6°36'57"
28	Col du Tronchet, NE Arvieux	44°47'41"	6°45'17"
29	Le Coin d'Arvieux, N Château Queyras	44°46'22"	6°43'24"
30	Pic du Grand Vallon, SE l'Argentière	44°45'21"	6°38'40"
31	Col du Lauzet, NE Guillestre	44°43'02"	6°40'08"
32	Col de Furfande, SW Château Queyras	44°43'57"	6°42'42"
33	Col de la Lauze, SW Château Queyras	44°43'12"	6°43'18"
34	Les Escoyères, SW Château---Queyras	44°43'13"	6°44'48"
35	Road to Montbardon, SW Château Queyras	44°43'16"	6°45'57"
36	Gros, NE Guillestre	44°40'38"	6°40'33"
37	Col du Fromage, Ceillac, E Guillestre	44°40'44"	6°47'57"
38	Combe Chauve, E Guillestre	44°39'22"	6°40'30"
39a	W Col du Longet, NE St Paul s/Ubaye	44°38'04"	6°55'53"
39b	E Col du Longet, NE St Paul s/Ubaye	44°38'07"	6°56'52"
40	S. Péouvou, NE St Paul s/Ubaye	44°36'30"	6°53'11"
41	Combe Brémond, NE St Paul s/Ubaye	44°35'53"	6°50'53"
42	lower Vallon de Mary, NE St Paul s/Ubaye	44°35'07"	6°51'27"
43	upper Vallon de Mary, Roure, NE St Paul s/ Ubaye	44°34'36"	6°51'39"
44	Bergerie de l'Alpet, NE St Paul s/Ubaye	44°35'09"	6°52'02"
45	La Barge, NE St Paul s/Ubaye	44°34'25"	6°48'50"
46	S Tête du Sanglier, NE St Paul s/Ubaye	44°33'56"	6°48'29"
47	N Brec de Chambeyron, E St Paul s/Ubaye	44°32'08"	6°50'43"
48	Col de Stroppia, E St Paul s/Ubaye	44°30'18"	6°51'30"
49	Pic de Crevoux, W Vars	44°33'39"	6°38'31"
50	Pic de Chabrieres, W Vars	44°34'03"	6°36'54"
51	Crevoux, E Embrun	44°33'00"	6°36'40"
52	Le Lauzet---Ubaye, W Barcelonnette	44°26'11"	6°26'17"
53	Super---Sauze, S Barcelonnette	44°20'08"	6°41'47"
54	Petit Cheval de Bois, Col d'Allos, S. Barcelonnette	44°17'36"	6°37'01"
55	Col de la Cayolle, S Barcelonnette	44°15'13"	6°43'27"

The PCAZ and the RAZ are separated by a major tectonic contact post-dating the initial nappe stacking and having possibly accommodated left-lateral displacement of the Roure-Acceglio units, which are truncated and pinch-out SE-wards along it (Preit fault zone, Lefèvre, 1984). A significant part of these southeastern Permian-dominated units, especially in the PCAZ, are devoid of their Mesozoic sedimentary cover for tectonic reasons. This detached cover may correspond in part to the Mesozoic nappe stack developed further N in the region of Briançon above the « Zone Houillère ». Another part, mostly observed in the RAZ, bears evidence of Mesozoic erosional truncation having removed, at least in part, the Triassic carbonate and siliciclastic series (Debelmas and Lemoine, 1957; Michard, 1959). The gap is increasingly important from the lower unit (core of the Acceglio anticline) to the upper one, the Pelvo d'Elva nappe (Lefèvre and Michard, 1976), and from SE to NW within the Pelvo d'Elva nappe (Lefèvre, 1982): the Jurassic to Cretaceous strata rest over lower Triassic quartzites near Acceglio (Lefèvre, 1962) and over Permian volcanoclastics near Col du Longet (Lemoine, 1960a).

6.3. *The most internal continental margin units*

The units named "Prepiedmont", (Lemoine et al., 1978) are located at the eastern border of the Briançonnais zone (fig. 1). Around the latitude of Briançon, they are detached along the Carnian evaporites and thus only composed of upper Triassic and more recent sediments (fig. 6). The basal thrust ramps up into the Jurassic section in one northwestern location (Chalvet, fig. 8). These units do not outcrop further north than the Maurienne valley, with the exception of the Grande Motte unit, whose paleogeographic origin is debated (Deville, 1986, and refs therein). To the southeast, in contrast, these units are complemented by older strata, due to a lower position of the detachment in the stratigraphic section. In Valgrana (fig. 1; Michard, 1961a, 1961b, 1967; Megard-Galli and Baud, 1977), the Triassic sequence of the Prepiedmont series include paleontologically dated middle Triassic strata overlain by late Triassic and early Jurassic formations dated by Franchi (1898), with a total thickness of 1 km to 1.5 km (unit III of Michard, 1967). The thick middle Triassic succession found in Valmaira can also be regarded as a Prepiedmont unit (unit I of Michard, 1967). Further southeast, in the Ligurian Alps, the Prepiedmont units include late Paleozoic formations and even polymetamorphic basement (Vanossi, 1991; Seno et al., 2004; Decarlis et al., 2017). This shows that the detachment of the Prepiedmont units climbs up section from southern regions (Liguria, Valgrana) towards the north-northwest along the central part of the arc (Cottian Alps).

6.4. *The Valaisan domain*

The occurrence of an oceanic domain sutured between the Briançonnais zone and the external zone, together with its age of opening and closure, remains a topic of strong debate (Stampfli et al., 2002; Bousquet et al., 2002; Beltrando et al., 2007; Masson et al., 2008; Loprieno et al., 2011; Beltrando et al., 2012; De Broucker et al., 2021). One reason underpinning this famous Alpine controversy is the lack of evidence for any continental breakup south of the Maurienne valley and west of the Briançonnais zone, which is a major difference between the northern and southern parts of the arc. A possible solution is to consider a southwestward transition from an oceanic domain, represented in eastern Switzerland and possibly floored by exhumed subcontinental mantle (Manatschal et al., 2006; Ribes et al., 2020; Le Breton et al., 2021) to a more or less attenuated continental crust domain pinching out towards the Vocontian basin, SE France (Bertrand et al., 1996; Dumont et al., 2012). The disappearance of the Valaisan zone towards the southern part of the western Alps arc would then occur for paleogeographic reasons. To the N(W) of the Briançonnais domain, the occurrence of a transitional paleogeographic domain towards this thinned crust area is testified by pinched units bearing specific "intermediate" Mesozoic stratigraphic signatures, frequently assigned to "subbriançonnais", all along the arc (Galster et al., 2010; Ceriani et al., 2001; Maury and Ricou, 1983; Barbier, 1963; Barale et al., 2017). This transition would have been incorporated within the western Alpine arc, creating complex lateral relationships between the Helvetic-Dauphinois, Valaisan, Subbriançonnais and Provençal units, as an expression of the lateral paleogeographic transitions around the scissors-shape Valaisan incipient breakup.

These lateral variations within the internal nappe stack of the Western Alps arc demonstrate that its structure is not concentric, and that radial profiles are not equivalent laterally and should not be considered as the optimal visualisation of a complex kinematic evolution during the Cenozoic. These features can be due to the obliquity of the

Table 2
Results of step-heating on phengite sample BLA-2016-10 (location site 13 fig. 10, and fig. 14f). ^{40}Ar/^{39}Ar analyses were conducted in the Noble Gas laboratory of Géosciences at the University of Montpellier 2, France, using a multicollector mass spectrometer (Thermo Scientific Argus VI MS) with a nominal mass resolution of ~200 and a sensitivity for argon measurements of 3.55 × 10−17 moles/fA at 200 μA trap current.

Step Nb	36Ar (fA)	37Ar (fA)	38Ar (fA)	39Ar (fA)	40Ar (fA)	Age (Ma)	2σ error	%40Ar rad	%39Ar
BLA 2016---10 phengite J= 0.0032864									
1	0,0289	0,0000	0,0000	0,445	0,940	12,50	31,15	9,91	0,07
2	0,0597	0,0573	0,0184	2,464	17,705	42,20	5,80	50,04	0,39
3	0,1902	0,0000	0,0482	26,281	222,336	49,59	0,60	79,75	4,15
4	0,2135	0,1411	0,0292	37,927	334,679	51,70	0,43	84,06	5,98
5	0,3793	0,0962	0,0000	90,172	815,654	52,97	0,24	87,84	14,22
6	0,4151	0,1050	0,0000	102,988	915,613	52,08	0,22	88,10	16,25
7	0,2991	0,0000	0,0000	71,193	636,032	52,33	0,27	87,72	11,23
8	0,4030	0,0000	0,0000	90,245	815,473	52,92	0,25	87,18	14,24
9	0,3484	0,2458	0,0257	66,026	579,366	51,41	0,28	84,83	10,42
10	0,0914	0,0428	0,0000	19,162	167,565	51,23	0,77	86,04	3,02
11	0,0570	0,0854	0,0000	16,510	145,471	51,62	0,89	89,53	2,60
12	0,0397	0,0000	0,0236	12,081	108,057	52,39	1,22	90,12	1,91
13	0,1751	0,0000	0,0406	55,288	504,260	53,41	0,31	90,61	8,72
14	0,2373	0,1452	0,0213	43,164	538,720	72,69	0,43	88,42	6,81

Total gas age : 53.59 ± 1.29

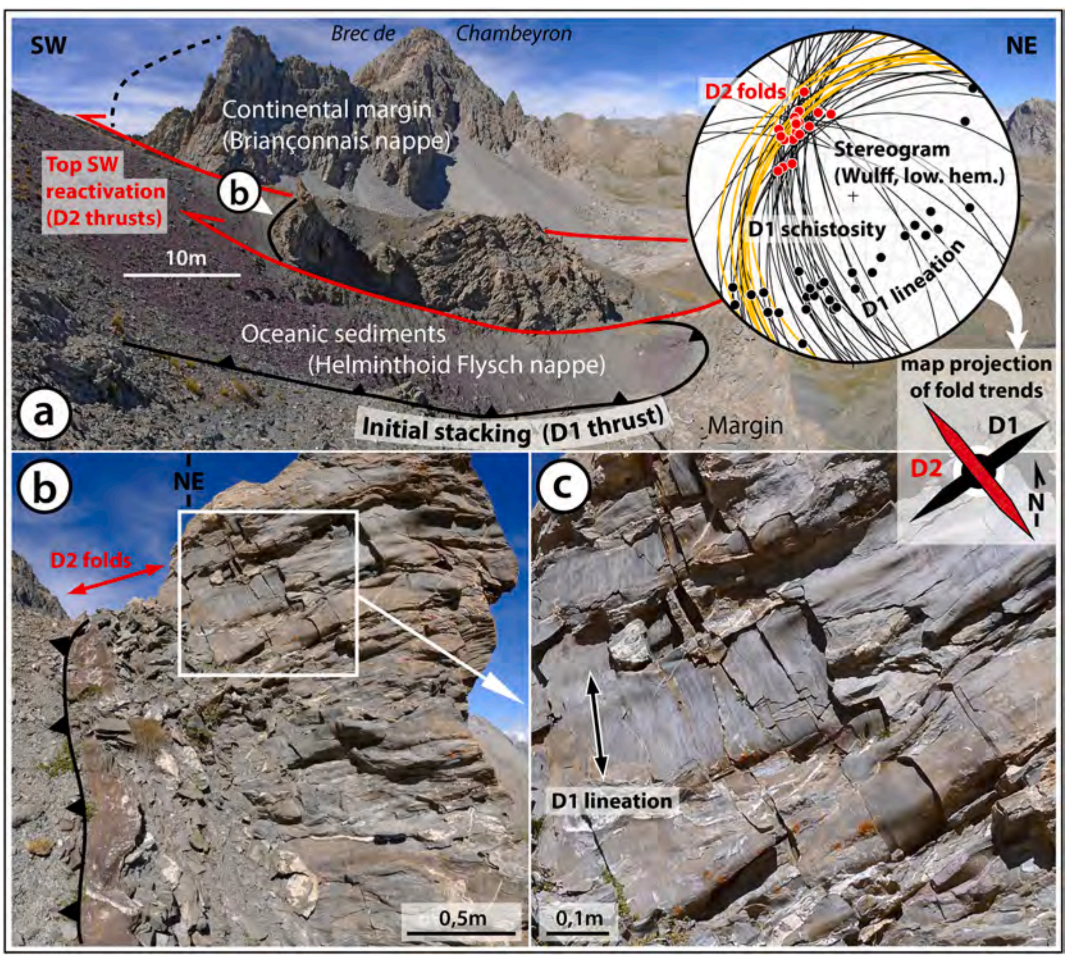

Fig. 12. Field example of D1/D2 microstructural interference in the external Briançonnais zone, south of the Ubaye valley (Stroppia pass near Fouillouse; site 48, location fig. 10 and table 1). a- Overall outcrop view towards NW, showing the early (D1) initial stack with the oceanic Helminthoid Flysch nappe overlying the Briançonnais Chambeyron nappe of continental origin. This early thrust (black) is crosscut by top-to-the SW D2 thrusts (red). The stereogram shows that D1 folds/ intersection lineations and S1 schistosity are scattered and deformed by D2 folding, both beeing perpendicular (projection). D1 and D2 folds are associated with NW-directed and SW-directed tectonic transport, respectively. b- Detail of D2 folds associated with mesoscale folding of D1 thrust. c- Close view of D1 intersection lineation and microfolds involved by perpendicular D2 folds, corresponding to the stereogram above.

orogenic wedge with respect to the paleogeographic trends, but can also be explained by oblique to subperpendicular crosscutting of the early orogenic nappe stack by recent (Oligo-Miocene) collision driven by westward extrusion (Dumont et al., 2012). The superposition order of structural units is a key argument for restoration, but in radial profiles this order is frequently disturbed and modified by recent outward "out-of-sequence" thrusting, i.e. in the Guil valley (Claudel and Dumont, 1999). The knowledge of the structural geometry of the initial nappe stack emplaced during the early stages of thrusting is thus a key criterion for a reliable restoration.

7. Sequence and superposition of nappes in the early orogenic stages

Since the building of the Internal Zones is polyphase, our aim in this section is to enlight some characteristic large-scale structural features associated with the early stacking phase (D1), with due consideration of their subsequent modification by younger deformation pulses (D2). This assists reconstruction of the continental subduction wedge large-scale geometry during the early orogenic stages, as discussed in § 9.2.

7a- Upper part of the continental subduction wedge, overlain by oceanic accretionary units:

A key feature for the understanding the structure of the internal Alpine nappe stack is the distinction between ophiolite bearing series and Triassic-soled series, now referred to as oceanic and continental margin units, respectively. This was recognized several decades ago (Lemoine and Michard, 1963; Lemoine, 1964; Michard and Schumacher, 1973; Bourbon et al., 1979), leading to the identification of imbrications of oceanic and continental units, especially in the southern part of the arc (Michard, 1967; Henry et al., 1993; Balestro et al., 2020). Serpentinite-bearing layers occur, for example, between the main Dora-Maira basement complex containing UHP relicts and the blueschist Permian-Triassic Dronero-Sampeyre unit (Henry et al., 1993; Compagnoni et al., 2012). Some others are found between the middle Triassic series of Val Maira and the Triassic-Jurassic series of Val Grana (units I and III of Michard, 1967, respectively), both derived from the Prepiedmont domain. The imbricated structures can result from different processes: (1) rift inheritance, with inversion of a complicated ocean-continent transition domain featuring continental allochthons separated by exhumed mantle patches (Beltrando et al., 2010b; Festa et al., 2020), or (2) subduction-collision processes, with polyphase deformation of an initially simple ocean/continent thrust contact at the bottom of the accretionary wedge.

While not excluding the first hypothesis, we currently prefer the second option considering our observations and structural data from the literature. Our structural and microstructural data (§ 5; fig. 2) between Briançonnais and Ubaye demonstrate that the internal nappe stack has been crosscut and disturbed by radially oriented forward and backward "out-of-sequence" thrusting. Such multistage deformation is documented in the central part of the arc (Gasco et al., 2011), and also in the southern part of the arc (Val Maira and Valgrana, west of Cuneo city) by Michard (1967) and Schumacher (1972), which emphasize multistage deformation and north directed backthrusting. We argue that oceanic/continental units imbrication is most likely due to a tectonic disturbance of the initial superposition of nappes: In the early stages of stacking, the oceanic nappes, either non-metamorphic (Helminthoid flyschs), or metamorphic sediments (Schistes Lustrés) and oceanic basement (Chenaillet, Monviso), were thrust over the continental margin units (Prepiedmont and Briançonnais). Such a superposition is locally preserved in different locations of the internal zones of the Western Alps: the Chenaillet ophiolites and metasediments overlying the Gondran prepiedmont unit (Lemoine, 1971; Barféty et al., 1995), the Monviso metaophiolites overlying the Dora-Maira massif (Balestro et al., 2011, 2013), the Mte Banchetta tectonic window (§4.2, Fig. 7).

Besides imbrications, other anomalous structures consist of continental margin units resting over oceanic units. Such a reverse superposition is shown for example by the Penninic thrust bringing the Briançonnais nappes above the Embrunais-Ubaye Helminthoid flysch nappes near Guillestre (Dumont et al., 2011), or by the Rochebrune Prepiedmont nappe backthrust over the oceanic metasediments of the Lac des Cordes unit (Dumont et al., 1984; Tricart et al., 1985). In both cases, this reversal is due to a tectonic disturbance of the initial nappe stack by the D2 deformation event (§ 4 & 5) and it must be restored to reveal the top of the continental subduction configuration beneath the accretionary wedge in the early orogenic stages.

7b- Internal structure of the Briançonnais nappe stack:

The Carboniferous "Zone Houillère" and associated Mesozoic cover units were initially located towards the base of the Briançonnais pile of nappes sensu stricto, both in Vanoise and in the region of Briançon (Caby, 1996; Barféty et al., 1995, 2006), as it is the case in the Swiss Alps (Escher et al., 1993). Generally detached from their polymetamorphic basement, these units may have been translated later to a higher structural position by either forward or backward deformation, as for example the Mont Thabor unit around Névache (Barféty et al., 2006; Lanari et al., 2012). In the region of Briançon, the "Zone Houillère" is overlain by a refolded stack of Mesozoic cover nappes detached along Triassic evaporite layers, which possibly represent the initial cover of the "Internal Briançonnais" units of the Permo-Carboniferous Axial Zone located further SE (Lefèvre, 1982).

On top of the Briançonnais thrust sheets scarce evidence of an uppermost polymetamorphic basement nappe are found (fig. 1), which triggered famous debates about identification of thrust-sheets in the Western Alps (Termier, 1899; Termier and Kilian, 1920). These uppermost units locally bear an "ultrabriançonnais" highly condensed type of Mezozoic series (Rio Secco; Lemoine, 1961a, 1964, 1967). To the north, possible equivalents of these units are the Sapey-Ruitor orthogneiss and micaschists in Vanoise, yielding Cambrian-Ordovician ages (Guillot et al., 2002) and which are affected by both Variscan and Alpine metamorphism. They are also located in the uppermost position of the Briançonnais stack. To the southeast, pre-Namurian polymetamorphic basement also occurs in the Internal Briançonnais units of Liguria (Cortesogno et al., 1981). These upper basement nappes which override the Permo-Carboniferous basins are disconnected from the internal crystalline basement massifs. As discussed later, we propose that the thrust sequence involving, from bottom to top, the Carboniferous Zone Houillère and Permo-Carboniferous units, the Meso-Cenozoic cover nappes, and the polymetamorphic basement thrust sheets, may result from tectonic inversion of late Variscan inherited paleogeography.

7c- Olistostromes on top of the early orogenic wedge:

Near Briançon, there is evidence for early and shallow emplacement of a basement nappe in an uppermost position. The so-called "quatrième écaille" (Termier, 1899) is composed of micaschists similar to Rio Secco, but bearing a condensed and specific Mesozoic cover: thin middle Triassic carbonates of more distal environments than the underlying typical Briançonnais nappes (Barféty et al., 1995) unconformably overlain by the Prorel breccia, which is probably late Cretaceous (Barféty et al., 1995). The basal thrust of this nappe is marked by an olistostrome of Bartonian age (Barféty et al., 1992) deposited on top of the Briançonnais stack, and containing various exotic blocks, including oceanic sediments (Helminthoid Flysch). These features underline the former connection of a thrust system inside the early orogenic wedge with the surficial part of the wedge and with the floor of the Paleogene flexural basin in front of the orogen. The N-NW-wards propagation of the orogen and of the preceding flexural basin throughout Paleogene times is well documented in the Western Alps (Sinclair, 1997; Kempf and Pfiffner, 2004; Ford et al., 2006; Schmid et al., 2017). Slope breccia and olistostromes which occur at the transition between the range and the basin are further incorporated as a diachronous marker layer in the footwall of the nappes (Kerckhove, 1969; Ford and Lickorish, 2004). The occurrence of detrital elements of mixed continental (basement, Mesozoic cover) and oceanic provenance (ophiolites, oceanic sediments including Helminthoid Flysch) may allow tracing of the propagation of the obducted accretionary wedge over the Briançonnais foreland in the early stages of continental subduction. However, there is a need to distinguish these thrust-related breccias within the various types of breccia observed in the sedimentary record of the Briançonnais and neighbouring areas.

8. Breccia and olistostromes

The scale and significance of synsedimentary breccias interbedded in the Briançonnais Meso-Cenozoic series and in the adjoining Subbriançonnais, Valais and Prepiedmont domains has been a subject of debate for many decades (Lemoine, 1967; Kerckhove et al., 1980; Chaulieu, 1992 and refs therein; Ribes et al., 2019). Various types of syntectonic breccia are observed in relation to different geodynamic settings, whose sedimentary signatures are difficult to distinguish. Scarp or slope breccia can occur in either divergent or convergent setting with similar sedimentological characteristics (Chaulieu, 1992). Apart from their facies, additional criteria to consider are (1) the occurrence of exotic material, which can reflect syn-orogenic exhumation of source areas or tectonic juxtaposition of oceanic and continental thrust-sheets, (2) their Alpine structural setting, shear fabric and location with respect to the major Alpine thrusts or plate boundaries (Polino et al., 1990). Thrust-related mélanges and associated mass-transport and tectonic processes are described in Festa et al. (2010), which helps to distinguish breccia related to extensional tectonics (type 1) from those related to subduction or collision, and to formalise the transition from precursor olistostromes to tectonic mélanges associated with nappes boundaries.

8.1. Breccia: a review

This section classifies the numerous examples of syndepositional breccia described in the Western Alps literature which respect to the geodynamic context of their occurrence. The following stages are considered: a) Tethyan rifting (early to early-middle Jurassic), b) Tethyan passive margin overlapping with the Atlantic-Bay of Biscay rifting (middle Jurassic to early Cretaceous), c) Initiation of convergence related to the Pyrenean orogeny (late Cretaceous to early Eocene), and d) Collision related to Alpine orogeny (early to late Eocene).

a) During the early Jurassic Tethyan rift stage, the deposition of syn-rift sequences occurred in an extensional setting before the oceanic opening of the Ligurian Tethys (early middle Jurassic). The preservation of such breccia is scarce in the Briançonnais, due to uplift and emergence at that time (Faure and Mégard-Galli, 1988; Claudel and Dumont, 1999), but lateral equivalents are known in the Prepiedmont units of Liguria (Mte Galero breccia, member A; Dallagiovanna and Lualdi, 1984; Decarlis and Lualdi, 2011), in the Cottian Alps (Narbona breccia; Michard, 1967; Michard and Schumacher, 1973; Gidon et al., 1978) and in the Prealps (lower member of the Breccia Nappe: Hendry, 1972; Steffen et al., 1993; lower member of the Evolène series, Mont Fort nappe: Pantet et al., 2020). Local supply, mass flow, debris flow and turbidites, together with local angular unconformities and submarine erosion (Dumont et al., 1984; Deville, 1986) are regarded as evidence of extensional rift context.

b) After the initial opening of the Ligurian Tethys, passive margin sedimentation on the European margin was affected by a second extensional pulse associated with the Atlantic-Bay of Biscay rifting and Iberian plate divergence (late Middle to Late Jurassic; Vergés and Garcia-Senz, 2001). Locally preserved in the early post-rift Briançonnais sediments (Tissot, 1954; Bourbon, 1980; Jaillard, 1987; Jaillard, 1999; Claudel et al., 1997), including the Ligurian Briançonnais (Bertok et al., 2011), and are widespread in nearby domains, both continent-ward (Subbriançonnais) and oceanward (Prepiedmont). They consist of the Telegraphe breccia in the Subbriançonnais units (Barbier, 1963; Barféty et al., 1977; Barféty et al., 2006) and coeval lateral equivalents (Nielard breccia, Barféty et al., 1977; Neyzets and Piolit breccia; Latreille, 1954; Chenet, 1978). Time-equivalent turbiditic breccia in Prepiedmont units are found in Liguria (Mte Galero breccia, members B and C, Decarlis and Lualdi, 2011), in the Briançon region (Lemoine et al., 1986) and in the Prealps (Breccia Nappe, upper member, Steffen et al., 1993). All correspond to rifted margin pelagic environments, indicative of increased transport and deeper erosion of the source areas than during the previous syn-rift stage.

c) The initiation of Africa-Europe convergence is recorded in the European margin sediments through the propagation of the Pyrenean foreland north of the Iberian plate, but before the complete closure of the Tethys and the Adria collision (late Cretaceous to Paleocene). The pelagic early Late Cretaceous sedimentary record in the Briançonnais contains breccia occurrences which, together with local erosional angular unconformities (Bourbon et al., 1976), possibly reflect incipient compressional reactivation of marginal structures (Chaulieu, 1992; De Graciansky et al., 2011): the Cerces breccia (Tissot, 1954; Barféty et al., 2006), the Mélézin breccia (Barféty et al., 1995), the Madeleine breccia (Gidon et al., 1994), and even olistoliths (Bourbon, 1980) are reported within the Cenomanian to early Senonian formations of several external Briançonnais nappes. Significant tectonic activity is also indicated by the occurrence of coarse breccia interbedded in the upper Cretaceous calcschists of the internal Briançonnais units of Vanoise (Fours unit, Deville, 1986; Tsanteleina and Chevril breccia, Jaillard, 1999). Further south, the most internal Briançonnais nappes contain coarse breccia with olistoliths interbedded in calcschistous formations probably late Cretaceous in age (Gidon et al., 1994): the Longet breccia (Lemoine, 1967) and similar breccias in the uppermost Pelvo d'Elva nappe further south (Lefèvre and Michard, 1976). Lateral equivalents are found in the Subbriançonnais nappes: Bachelard and Pelat flyschs: Blanc et al., 1987, Thum et al., 2015; l'Argentière breccia: Chenet (1978). All these clastic formations are usually interpreted as derived from degraded submarine fault scarps, but their compositions frequently suggest a deeper erosion of the source areas and increased transport than the Jurassic breccias. Angular erosional unconformities affecting the post-rift sediments (i.e. Guil lower nappe or Galibier area) with locally high-angle relationships (Tissot, 1954) may result from compressional reactivation of rift structures (i.e. Béraudes fault, Lemoine et al., 2000). Alternatively, Bertok et al. (2012) describe early Late Cretaceous kilometric-scale normal fault scarps in the Ligurian Briançonnais which are interpreted to have formed in a transcurrent regime. This late Cretaceous deformation must be related to the motion of the Iberia microplate relative to Europe (Le Breton et al., 2021) and to the propagation of the Pyrenean foreland within its eastern continuation (Corsica, Sardinia, Briançonnais). Coeval evidence of inversion is known in Provence, the Maritime Alps (Schreiber et al., 2011), further north (Dévoluy pre-Santonian folding; Michard et al., 2010) and in northern Spain (Soto et al., 2011).

d) Detrital sedimentation finally occurred after the complete closure of the Ligurian Tethys, and recorded the propagation of the Alpine orogenic wedge resulting from the Adria-Europe collision (Eocene). Sedimentation ceased in early Oligocene due to involvement of the whole internal Alpine zones within the collision wedge. This propagation developed an underfilled flexural basin over both the Briançonnais domain and the more proximal parts of the margin (Ceriani et al., 2001; Sinclair, 1997). Its sedimentary infill is diachronous, spanning from Lutetian to Bartonian in the Briançonnais (Barféty et al., 1995; Michard and Martinotti, 2002) and from Lutetian to Priabonian in autochthonous Corsica and in the External Alpine zone (Durand-Delga, 1984; Joseph and Lomas, 2004, and refs. therein). The flexural basin was propagating towards the NW throughout the Eocene (Kempf and Pfiffner, 2004; Ford et al., 2006; Dumont et al., 2012). It is generally overlain by a widespread olistostrome layer marking the base of the obducted oceanic accretionary wedge. This olistostrome formation, named "Schistes à blocs" (Kerckhove, 1969) is also diachronous and locally dated as Late Eocene-Early Oligocene beneath the non-metamorphic thrust-sheets which overly the external zone of the Western Alps (Dumont et al., 2012, and refs. therein). In the external Briançonnais nappes, few occurrences of such deposits are observed, always located on top of the tectonic pile.

The most striking example is the "Quatrième écaille" near Briançon, which has been subject to debate since ~1900 (Lemoine, 1961b), and which is now interpreted as a kilometre-scale block which slid above the Briançonnais basin infill in Early Bartonian times (Barféty et al., 1992). Its exotic character is shown by a specific, polymetamorphic basement only found in the most internal zones, that is the uppermost unit of the Briançonnais thrust stack (a, fig. 13; white stars), and by open-marine Middle Triassic facies different from the underlying nappes. It is comparable to distal ramp carbonate facies described in Liguria (Decarlis and Lualdi, 2009). Nearby and beneath this block are found conglomerates and breccia fed by it, thus containing mostly basement (Eychauda breccia, Barféty et al., 1992, 1995). Similar breccias are found in several places at the same structural and stratigraphic level (a, fig. 13, red stars; b, c, fig. 13). Finally, "mixed" breccia consists of a rare occurrence of mafic or ultramafic pebbles within coeval olistostrome layers (d, fig. 13, blue stars 22 to 24), together with more frequent pebbles of Cretaceous oceanic sediments (Helminthoid Flysch). They were fed by both continental and oceanic fragments from the orogenic front. We argue that these "mixed" detrital formations, always located on top of the Briançonnais nappe pile and/or beneath the exotic flysch nappes, can be useful in identification of the earliest stage of propagation of the accretionary wedge.

8.2. *The significance of mixed breccia*

Few occurrences of "mixed" detrital formations, namely containing clastic elements derived from both continental margin and oceanic domains, have so far been reported in the Western Alps, apart from their intra-oceanic precursors emplaced during the late Cretaceous closure of the Tethyan domain (Deville et al., 1992; Balestro et al., 2015). The Late Eocene olistostrome covered by the Helminthoid Flysch nappes of oceanic origin contains frequent Late Cretaceous oceanic blocks (Autapie type Helminthoid flysch, Kerckhove, 1969). Exceptional blocks of oceanic basement (basalts) are observed in two localities (a, fig. 13, sites 23 and 24). Although not dated, the Paneyron mixed detrital formation ("Ophiolite de Sérenne"; Kerckhove, 1969) is probably a lateral equivalent of this layer. Other occurrences are reported from more internal parts of the arc, but remain a subject of debate in the absence of reliable age data. These are, from south to north (1) the Valliera and Tibert megabreccia in Valgrana (Michard, 1967), which contain siliciclastic facies, dolomitic blocks and serpentinite elements associated with kilometre-scale ultramafic lenses. Initially interpreted as possibly interlayered within the Mesozoic stratigraphy, these megabreccias occur between the underlying Prepiedmont Triassic-Liassic units and the tectonically superposed oceanic "Schistes Lustrés" nappes (Michard and Schumacher, 1973), (2) the Cula breccia (Gout, 1987) which must be distinguished from the Col du Longet breccia (Lemoine, 1967; Gidon et al., 1994). The latter, fed by nearby internal Briançonnais areas, is linked with the upper Pelvo-d'Elva nappe (§ 4.1; fig. 4) and does not contain any oceanic clasts (Michard, pers. com. 2019, and personal observations), whereas the Cula breccia which contains basalt and serpentine pebbles is located at the base of the oceanic "Schistes Lustrés" pile of nappes (Cula unit; Gout, 1987) (3) the Prafauchier series (Dumont, 1983), containing continental margin derived clasts (dolostones, micaschists) and serpentinite grains, with locally serpentinite blocks, and interpreted either as part of the Prepiedmont series, or belonging to an overlying unit, (4) the Rocher Renard breccia linked with the Lago Nero unit at the base of the Chenaillet ophiolitic massif (Polino and Lemoine, 1984; Burroni et al., 2003; Principi et al., 2004), (5) the Mte Banchetta breccia (Caron, 1971), containing blocks and pebbles of serpentinite, oceanic sediments (Jurassic marbles and ophicalcite) and continental margin sediments (Triassic dolostones and sandstones), in a shaley matrix. This latter example crops out on top of a folded Prepiedmont unit (§ 4.2; fig. 7), and is described below.

In addition, two occurrences of ophiolite bearing tectono-sedimentary breccia in the western Alpine arc may have a comparable significance: the Montaldo calcschists in Liguria (Dallagiovanna et al., 1991; Decarlis et al., 2013) which contain olistoliths of serpentinites and prasinites, and the Gets flysch (Caron and Weidmann, 1967; Caron et al., 1989; Bill et al., 1997) resting at the top of the Prealps tectonic pile and containing ophiolitic and granitic blocks. Remarkably, all these examples of mixed continental and oceanic breccia occur immediately above the highest continental margin units, generally the Prepiedmont nappes which represent the most distal part of the European continental margin, and are overlain by the lowermost metasedimentary units of the oceanic "Schistes Lustrés" derived from the accretionary prism (Schwartz, 2000). These "mixed" clastic formations are thus a useful marker of the initial geometry of the subduction trench, later deformed by further collision phases. We argue that their formation is related to the translation of the subduction trench over the distal European margin, and that their significance may be different from other mixed breccias deposited on the oceanic floor and related to either subcontinental mantle exhumation near the ocean-continent transition (Meresse et al., 2012) or early subduction of the Tethyan floor (Marroni and Pandolfi, 2007).

8.3. *The Monte Banchetta mixed breccia, an olistostrome at the bottom of the oceanic accretionary wedge?*

A key example of mixed breccia is found in Mte Banchetta, near Sestriere (fig. 3, site e). The structural setting, described in § 4.2, is interpreted to be a cross-fold anticline producing a dome. An uplifted continental margin unit, probably the same as the Prepiedmont Gran Roc unit further SW, outcrops in a tectonic window (fig. 7; Jouvent, 2017). The underlying series was described by Caron (1971) and regarded as analogous to the Prepiedmont series of Valgrana (Michard and Schumacher, 1973). A recent map (Corno et al., 2019), does not follow this attribution, simply calling it "continental succession", but three diagnostic formations of the Prepiedmont series can be identified, namely the Norian dolostones, the Rhaetian dolostones, limestones and schists, and the "Lias prépiémontais" calcschists ("calcschistes siliceux ankéritiques" of Caron, 1971). This series which crops out in the Mte Banchetta area is not overturned, except on the eastern slopes due to backfolding (fig. 7). On the west side of Mte Banchetta (a, fig. 14, site c), the upper part of the calcschists is overlain by the breccia, through a dark grey schist layer containing scarce pebbles (c, fig. 14) and bearing a specific stilpnomelane mineralization (Caron, 1970). The same schist is also found above as a matrix of the breccia (Caron, 1971; b, d, fig. 14). Laterally (Clot della Mutta; fig. 14, south of site c), the Prepiedmont calcschists are overlain by an hectometric-sized serpentinite sliver in apparent tectonic contact, but this contact seems to pinch out beneath the schist layer at site c. This feature can be interpreted in two ways: either this sedimentary layer is inserted along the contact, thus the serpentinite sliver is a larger block belonging to the breccia, or it seals the tectonic contact between serpentinite and calcschists. In both cases, the breccia must be regarded as an olistostrome deposited over the Prepiedmont series.

With the exception of the basal schist layer which contains rounded pebbles, the breccia has a very proximal character based on the size and shape of the blocks. This suggests that it was fed by nearby and active submarine scarps involving both continental margin (Prepiedmont) and oceanic floor units. Corno et al. (2019) propose an interpretation of this polymictic breccia in an hyperthinned marginal setting, in response to the Jurassic rifting. Alternatively, considering that the breccia is overlying the Prepiedmont marginal series and that it is located in the footwall of mixed imbricates (a, fig. 14), we propose that the Mte Banchetta breccia was deposited on the European margin toe in a subduction trench setting, marking the earliest stage of obduction of the accretionary wedge.

Similar olistostromes are known beneath the Helminthoid flysch nappes, the non-metamorphic equivalent to the oceanic Schistes Lustrés ("Schistes à blocs" fm.; Kerckhove, 1969). They locally contain ophiolitic detritus (§ 8.1; d, fig. 13), and we propose that they represent non-

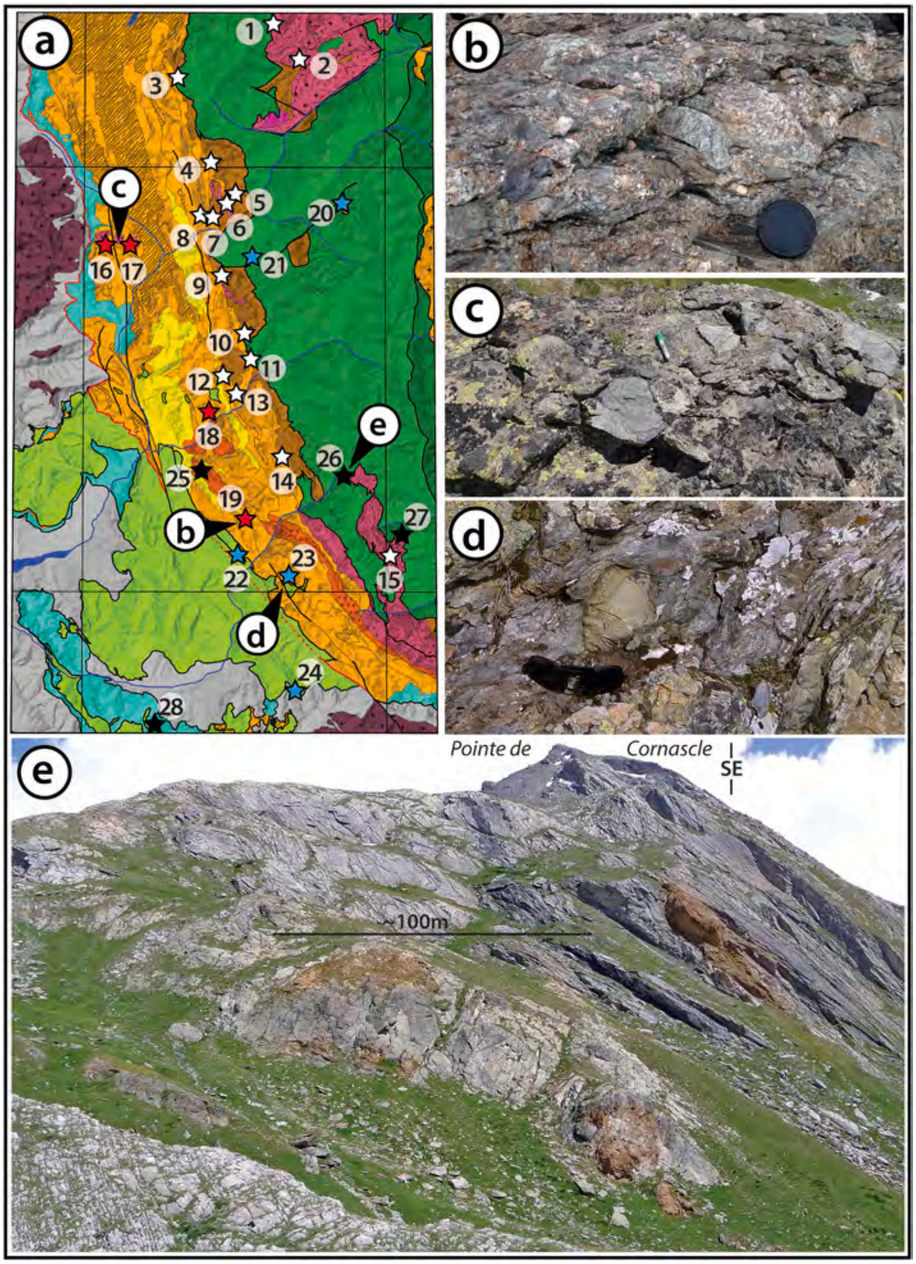

Fig. 13. Location of uppermost basement slices and of synsedimentary breccia and olistostromes on top of the Briançonnais/Prepiedmont nappes stack, emplaced during the surficial propagation of the early orogenic wedge (from literature and personal observations). a- Location map (legend of numbers below). White stars: polymetamorphic basement slices witnessing the widespread occurrence of an uppermost "internal" Briançonnais nappe likely issued from an area devoid of late-Variscan sediments (see § 7). Black stars: Examples of upper Cretaceous (dated or attributed) detrital formations deposited Adria-Europe collision, assumed to result from Pyrenean orogenic propagation (§ 8.1, c). Red stars: Eocene synsedimentary breccia and olistostromes containing pebbles of Variscan metamorphic basement, feeded by this upper basement nappe (examples b and c below); the host flysch sediment is locally dated from Bartonian (Barféty et al., 1992). Blue stars: "mixed" synsedimentary breccia and olistostromes, containing both continental margin and oceanic basement clasts; at 22 to 24, located in external Briançonnais, the breccia belong to the "Schistes à blocs" formation, an olistostrome located at the bottom of the oceanic Helminthoid Flysch nappes (example d below); sites 20–21, east of the Briançonnais zone, could be an equivalent of such olistostrome later involved in the orogenic wedge. b- Polygenic breccia containing pebbles of Variscan metamorphic basement, dominantly gneiss, interbedded in the Eocene "Flysch Noir" formation of the Chatelet Briançonnais nappe, east of Vallon Laugier, below Pic des Houerts (Gidon et al., 1994). c- The Eychauda breccia, same kind of breccia feeded by micachists of the Prorel unit (« Quatrième écaille », § 7 and § 8.1), a Bartonian olistostrome sitting on top of the external Briançonnais nappes stack (Serre Chevalier, near Briançon). d- Mixed breccia with basalt pebbles in the Eocene « Schistes à blocs » fm., SE of Fouillouse village, left side of Ubaye valley. This olistostrome formation is overlain by the Helminthoid Flysch nappe of oceanic origin. e- Col du Longet megabreccia, with spectacular decametric dolostone blocks included in the Cretaceous calcschist formation (§8.1, c).

Location of outcrops:

1- Col d'Etaches (Barféty et al., 2006)
2- Passo dei Fourneaux (Polino et al., 2002)
3- Pian dei Morti fort, Rho valley (Polino et al., 2002)
4- Chalets des Acles (Barféty et al., 2006)
5- Top of Chaberton peak (Barféty et al., 1995)
6- Rio Secco (Termier, 1899; Lemoine, 1960c ou Lemoine, 1961b)
7- Montgenèvre (Barféty et al., 1995; site12, fig. 10 & 11; fig. 8d)
8- Le Rosier, La Vachette (Barféty et al., 2006)
9- Cervières (Barféty et al., 1995)
10- Brunissard, S. Izoard pass (Lemoine, 1961a)
11- Lac de Souliers – Col du Tronchet (Lemoine, 1961a; Debelmas et al., 1966)
12- W. Arvieux (Lemoine, 1961a; Debelmas et al., 1966)
13- E. Villargaudin (Lemoine, 1961a; Debelmas et al., 1966)
14- E. Ceillac (Lemoine, 1961a; Debelmas et al., 1966)
15- Colle delle Sagneres, N. Acceglio (Lefèvre and Michard, 1976)
16- Eychauda conglomerates (Barféty et al., 1992; Barféty et al., 1995)
17- Prorel micaschists (Termier, 1899; Lemoine, 1960b, 1960c, 1964; Barféty et al., 1992)
18- Furfande microconglomerates (site n°32, Figs. 10 & 11)
19- Col des Houerts conglomerates (Gidon et al., 1994)
20- Monte Banchetta, Sestriere (Caron, 1971; fig. 14)
21- Rocher Renard, S. Chenaillet (Burroni et al., 2003)
22- Col de Sérenne, E. Col de Vars (Kerckhove, 1969: Gidon et al., 1994)
23- Fouillouse (Gidon et al., 1994)
24- Les Sagnes, S. Larche (Kerckhove, 1969; Gidon et al., 1978)
25- La Madeleine breccia, Val d'Escreins (Gidon et al., 1994)

26- Col du Longet breccia, high Ubaye valley (Lemoine, 1967; Gout, 1987)
27- Pelvo d'Elva breccia (Michard, 1967; Lefèvre and Michard, 1976)
28- Bachelard and Pelat flyschs (Blanc et al., 1987; Thum et al., 2015)

metamorphic lateral equivalents to the Mte Banchetta mixed breccia.

Some Ar dating was completed from phengites sampled in the Prepiedmont Rhaetian formation beneath the breccia (f, fig. 14; Jouvent, 2017). Variscan inheritance can be ruled out because there is no significant reworking of Variscan material in the late Triassic sediments of the Prepiedmont series (Dumont et al., 1984). Hence the 52.31±1.32 Ma age obtained (g, fig. 14) indicates that the deposition of the Mte Banchetta mixed breccia occurred during early Eocene or before. Although this HP age needs to be supported by further dating studies, the underthrusting of the Mte Banchetta Prepiedmont unit which produced this metamorphic event seems consistent with the early exhumation of the oceanic Schistes Lustrés (Agard et al., 2002; Herviou et al., 2022), and with the slightly younger involvement of the Briançonnais units in continental subduction (Bucher et al., 2004; Berger and Bousquet, 2008; Strzerzynski et al., 2011). This is also consistent with the age of Adria-Europe collision in the Western Alps according to geodynamic models (Rosenbaum and Lister, 2005; De Graciansky et al., 2011; Manzotti et al., 2014a; Pfiffner, 2014; Van Hinsbergen et al., 2020; Le Breton et al., 2021).

9. Discussion and geodynamic implications

Our data allow identification of two main stages during the building and evolution of the internal zones of the Western Alps arc, corresponding to nappe stacking and to westward extrusion, respectively. The orientation of shortening and tectonic transport changed significantly through time, as shown by subperpendicular fold and lineation trends in the study area. The late stage (D2, §4, §5) corresponds to the activation of forward and backward thrust systems which accommodate the exhumation of the Briançonnais stack (fig. 2). It is responsible for the formation of the arc driven by westward migration and indentation of Adria upper mantle (Malusà et al., 2016; Schmid et al., 2017; Nouibat et al., 2022), and the most prominent thrusts and fold trends result from this late stage, such as the Penninic thrust near Briançon. Deciphering the early stage (D1) is much more difficult because the associated structures are overprinted, deformed and crosscut by the second one at all scales. However, proper integration of this early deformation phase is critically important because it recorded the absorption of N-S convergence required by plate tectonics during Alpine orogenesis (Schmid and Kissling, 2000; Handy et al., 2015; Van Hinsbergen et al., 2020). The large amount (several hundred km.) of early N-S shortening is well documented in the Central and Eastern Alps (Pleuger et al., 2007; Scharf et al., 2013; Scheiber et al., 2013; Steck et al., 2015; Handy et al., 2015 and refs. herein), but cannot be kinematically linked with the westward extrusion and radial spreading dynamics of the Western Alps arc, whose formation mainly postdates the initial northward drift of Internal Alpine nappes.

9.1. Formation and kinematic evolution of the western Alpine arc

The Western Alpine arc is a very specific feature of the Alpine chain, because its orogenic dynamics cannot be directly linked with the Africa-Europe motion path, but requires the involvement of an intermediate indenter between the Adria plate and the subducting European plate (Platt et al., 1989b; Rosenbaum et al., 2002; Le Breton et al., 2021; Nouibat et al., 2022) that produced a non-cylindric 3D structure. This complex finite geometry makes restorations more complicated because indentation and extrusion produced a specific lithospheric structure and metamorphic record during collision (Schmid et al., 2004; Bousquet et al., 2008; Beltrando et al., 2010a; Handy et al., 2010; Schmid et al., 2017; Salimbeni et al., 2018). The arc was completed during the more recent stages of Alpine orogenic evolution (Caby, 1996; Schmid and Kissling, 2000; Ford et al., 2006; Maffione et al., 2008; Handy et al., 2010; Dumont et al., 2011; Ring and Gerdes, 2016; Le Breton et al., 2021), accompanied by synorogenic anticlockwise rotations which increase in magnitude towards the south (Thomas et al., 1999). This evolution involved major changes in collision kinematics through time, which have been identified in many localities of the Internal Zones, in the Prealpine and Embrunais nappes, and in the Helvetic-Dauphinois domain (Caby, 1975; Merle and Brun, 1984; Choukroune et al., 1986; Baird and Dewey, 1986; Platt et al., 1989b; Ramsay, 1989; Le Bayon and Ballèvre, 2006; Dumont et al., 2011; Scheiber et al., 2013). Nevertheless, the structure of the arc is often examined on the basis of radial cross sections (i.e. Schmid et al., 2017), which assume that the major displacements occurred perpendicular to the present trend of the chain. This "radial spreading" model, still broadly accepted despite being questioned by Goguel (1963), may be more or less valid in the northern part of the arc, where the change in orientation of tectonic transport through time remains moderate (Escher et al., 1997; Steck, 2008; Steck et al., 2015). An anticlockwise change in translation path is more significant in the central part of the arc and increases southwards (Merle and Brun, 1984; Ceriani et al., 2001; Handy et al., 2010). The total estimate of radial shortening remains significantly higher in the northern half of the arc, including the foreland and the Helvetic nappes (Epard, 1990; Sinclair, 1997; Burkhard and Sommaruga, 1998; Schmid and Kissling, 2000; Bellahsen et al., 2014; Pfiffner, 2016) than in the southern part (Ford et al., 1999; Ford and Lickorish, 2004). In the former part and towards the Central Alps, the total shortening represents the cumulative effects of the early nappe stacking and of the later extrusion with backfolding, both oriented S-N to SE-NW (Escher et al., 1997; Schmid et al., 1997; Steck, 2008; Steck et al., 2015). Conversely, there is an increasing angular discrepancy towards the southern part of the arc between the early (N to NW) and late (SW to S) thrusting stages. Consequently, an increasing part of "along-strike" tectonic transport due to the early phase, which is difficult to detected on radial sections, should be expected towards the southern part of the arc. Our field observations presented above support this interpretation, and provide evidence for "orogen-parallel" (N- to NW-directed) early nappe stacking within the Briançonnais s.l. zone. Moreover, this orogenic stage was possibly responsible for greater lateral displacements than the later extrusion stage.

9.2. Restoration, southern origin of the Internal nappes

Such a perspective may have implications for paleogeographic restorations and for the interplay between pre-Alpine (Variscan and Tethyan) and Alpine structures. In the southern part of the arc, it is difficult to decipher the provenance and thrusting sequence of the internal nappes simply on the basis of their geometrical expression and structural relationships along radial cross-sections. The arc needs to be retrodeformed in several steps, taking into account successive retrotranslation paths (Laubscher, 1988, 1991; Schmid and Kissling, 2000). Any unfolding process must consider an intermediate step of restoration, which reconstructs the superposition of structural units during the earliest orogenic stages, and which must be oriented appropriately in consideration of the early kinematic history, that is ~SE-NW (Michard et al., 2004; Tricart and Schwartz, 2006). Such an attempt is illustrated in fig. 15 (a). This schematic profile deals with the late Eocene situation, during the S- SE-directed continental subduction of the distal European margin. This situation represents the early stage of underthrusting of the Briançonnais and Prepiedmont domains beneath the oceanic accretionary wedge and the Adria plate, consistently with the palinspastic

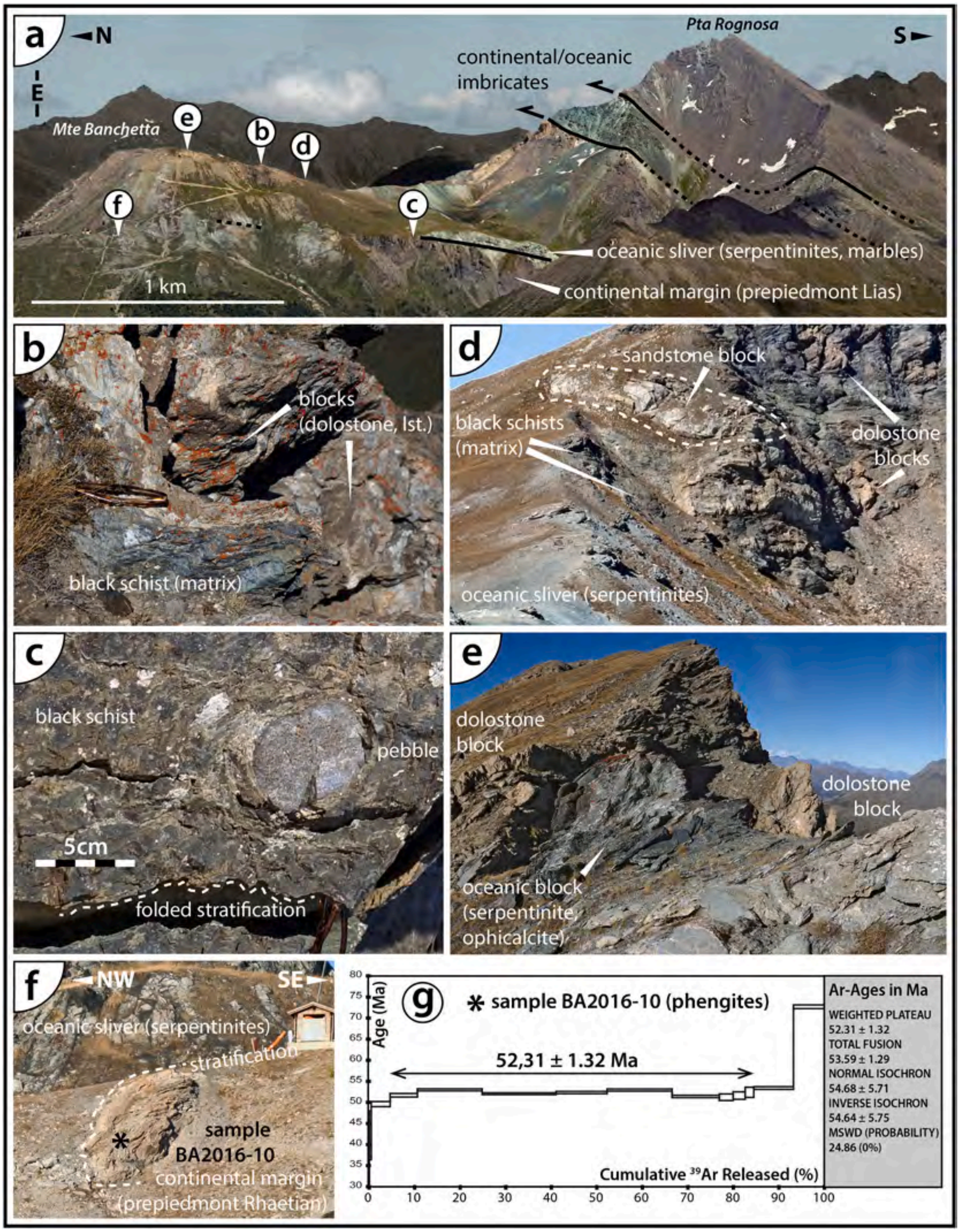

Fig. 14. The mixed megabreccia of Mte Banchetta, near Sestriere, Italy. The structure of this area is illustrated in fig. 7, location fig. 1b. The outcrops overlain a Prepiedmont type series with typical upper Triassic to lower Jurassic formations, and are overthrust by the serpentinite-bearing oceanic Schistes Lustrés nappes, with some mixed slivers in between. The breccia was feeded by both continental and oceanic series. a- Panoramic view towards E, from the top of Chaberton peak, and location of further observation points. The central part shows an oceanic sliver tectonically overlying the Prepiedmont continental margin series. This stack grades laterally to the megabreccia, which interrupts the tectonic contact at point c. These features are overthrust towards N-NW by continental and oceanic thrust-sheets (Pta Rognosa). b- Close view of the block-supported breccia with black shaley matrix. c- Close view of black shaley sedimentary layers resting over the lower Liassic Prepiedmont fm. next to the northern termination of the oceanic sliver, and containing siliceous and carbonate rounded pebbles. This layer, described by Caron (1970), corresponds to the host sediment of the megabreccia (b). d- Outcrop view towards N of the megabreccia resting through shaley matrix over the oceanic sliver, which could be regarded as an olistolith in the megabreccia. The blocks are sourced from the upper Triassic-lower Jurassic layers of the Prepiedmont series. e- Example of mixed oceanic and continental metric blocks (serpentinites/ophicalcite and dolostones) juxtaposed in the upper part of the breccia. f- N-NW directed drag fold affecting the upper Triassic schists and dolostones in the footwall of an oceanic sliver (site 13, fig. 11, table 1), consistent with the N-NW directed thrusting of the Pta Rognosa imbricates (a). g- Preliminary geochonological data ($^{39}Ar/^{40}Ar$ on phengites in upper Triassic schistous dolomitic layer, sample location f) suggesting that the Prepiedmont series which floored the Mte Banchetta megabreccia was involved in the collision wedge during early Eocene. This would provide a minimum age for the deposition of the breccia. $^{39}Ar/^{40}Ar$ dating results in Table 2.

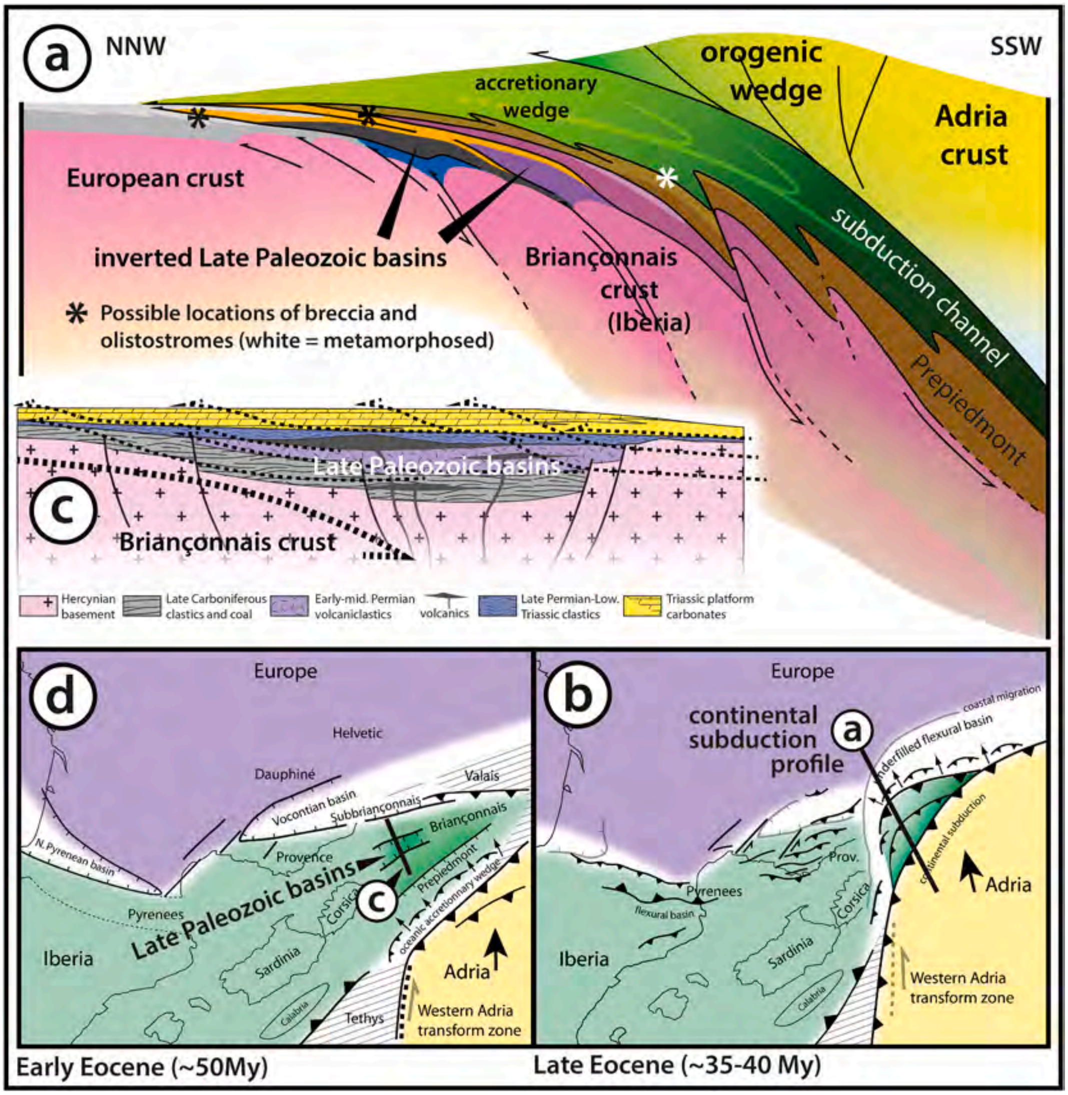

Fig. 15. a- Theoretical profile of the early Alpine wedge during the Adria-Iberia collision stage, trending parallel to the orogenic propagation at this stage, that is NNW-SSE (no vertical scale). This evolution is controlled by uncoupling processes, through the ability of sedimentary cover to be detached and to feed the shallow part of the wedge beneath the accretionnary wedge, whereas basement units are diachronously underplated. The Briançonnais "Zone Houillère" and "Permo-Carboniferous axial zone" Paleozoic basins are represented at an incipient stage of inversion, and should be overlain by thin polymetamorphic basement thrust sheets not visible at this scale. b- Paleogeographic sketch of Adria-Iberia collision stage with location of profile a (after Dumont et al., 2012, adapted). c- Recontructed part of the profile across the Briançonnais Paleozoic basins before beeing affected by thrust sequence and overlain by the Prepiedmont and oceanic nappes. This involves the occurrence of a southern uplifted border with exhumed basement, a potential origin for the polymetamorphic thrust-sheets presently observed on top of the Briançonnais nappes stack (§ 7). Similar reconstruction is provided from Ligurian Briançonnais units (§ 3). d- Paleogeographic sketch before Alpine collision (earliest Eocene), with the postulated situation of the Paleozoic basins within the Briançonnais domain belonging to the Iberian plate (after Dumont et al., 2012, adapted).

reconstruction of Dumont et al., 2012, fig. 15b). Inversion and continental subduction at this stage must have been influenced by rift inheritance, that is by the vertical and lateral distribution of lithospheric thinning in the distal margin as emphasised by recent research (Mohn et al., 2012; Masini et al., 2013; Le Breton et al., 2021; Tavani et al., 2021). Crustal underthrusting likely benefited from the thinned (Briançonnais continental ribbon) to hyperthinned (Prepiedmont margin toe) character of the distal margin. Thin-skinned processes in the upper part of the nappe stack were enhanced by the occurrence of various potential detachment layers in the Late Carboniferous to early Cenozoic sedimentary sequence. Like in the Provençal domain (Decarlis et al., 2014; Espurt et al., 2019; Balansa et al., 2022), the widespread Briançonnais and Prepiedmont Triassic evaporites allowed large fragments of the Meso-Cenozoic cover to detach and remain in a shallow burial environment. Interestingly, this process could not apply to the internal Briançonnais units, derived from the marginal uplift area (§ 3), where all the potential detachment layers had been eroded during rift-related uplift. This may explain in part the higher metamorphic grade experienced by these units. This interpretative profile is also intended to highlight the incipient underthrusting of the European crust s.s. (proximal margin) with the pinched Valais-Vocontian attenuated crust area between the European crust and the Briançonnais subduction wedge.

The nappe stack includes, from bottom to top, (1) Late Paleozoic sedimentary and volcanoclastic Briançonnais units (upper Carboniferous "Zone Houillère" and "Permo-Carboniferous axial zone"); (2) Briançonnais Mesozoic siliciclastic and carbonate sedimentary sequences, either covering the former units, or detached along Triassic evaporitic layers and duplicated. These units contain some remnants of Eocene foreland basins which provide time constraints for thrust initiation. They are derived from a marginal area moderately affected by syn-rift erosion, thus away from the main marginal uplift, (3) Permian to Mesozoic, mostly siliceous nappes, so-called "ultrabriançonnais" or

"Acceglio" type units. These units are derived from the most active rift shoulder area, devoid of Mesozoic platform carbonates due to deep syn-rift erosion, (4) scarce occurrence of Briançonnais polymetamorphic basement thrust sheets with local evidence of overlying unconformable Triassic sediments, featuring inheritance from an elevated late Variscan area devoid of any post-Variscan basins, (5) Prepiedmont nappes transported from an area close to the continent-ocean boundary, beyond the rift shoulder uplift, and providing evidence for northward thrust system propagation. The occurrence of rift shoulder and of megabreccia associated with the uppermost Briançonnais units and with some Prepiedmont series (§ 8) suggests that the hinge zone between both could correspond to a major necking zone within the pre-existing continental margin (Ribes et al., 2019).

The original configuration of these different units cannot be deduced from their present metamorphic signature, which results from the activation of detachments through time and on the depth of their involvement within the collision wedge (i.e. Michard et al., 2004), and also because the initial metamorphic record can be overprinted during younger stages of collision/extrusion (Lanari et al., 2014; Schwartz et al., 2020).

Following our interpretation, the paleogeographic origin of the internal nappes should not be located within the core of the Western Alps arc, which is problematic to restore due to overlap, but more probably in the southeast of their present location. Consequently, unfolding the profile of fig. 15 (a) would restore the Briançonnais nappes far to the south with respect to the Dauphiné-Helvetic domain, especially considering the gap represented by the closure of the Valais-Vocontian basin presently squeezed along the Penninic thrust. Many arguments support an original southern location near to the Provence-Corsica domains (Maury and Ricou, 1983; Stampfli et al., 2002; Handy et al., 2010; Thum et al., 2015). The thick Permo-Carboniferous volcanic and clastic sequences characterise the southern Variscan foreland and are different from the Dauphiné-Helvetic basement, which was closer to the Variscan axial chain (Guillot and Ménot, 2009; Ballèvre et al., 2018) This paleogeographic situation persisted into the Permian (Bourquin et al., 2011). The late Variscan magmatic and volcanic events within the Briançonnais units mark the onset of post-Variscan lithospheric thinning and associated thermal effects (Dal Piaz, 1993), and they are different and more recent than those in the Dauphiné/Helvetic zone (Bertrand et al., 2005; Manzotti et al., 2014b; Ballèvre et al., 2020). Conversely, Carboniferous clastics and coal measures together with Permian calc-alcaline volcanism show similarities to Provence and Corsica (Basso, 1987; Toutin-Morin and Bonijoly, 1992). The Triassic series of the Maritime Alps is very similar to the Briançonnais sequence (Maury and Ricou, 1983; D'Atri et al., 2016) and evaporitic potential detachment layers which controlled the Briançonnais cover deformation are widespread in Provence (Bestani et al., 2015). The northern Provence platform edge, running eastwards to crosscut the Argentera massif cover (Barale et al., 2017), is crosscut and transported northwards beneath the Embrunais-Ubaye nappes (Séolane-Cap unit) and further north in the Prealps nappes (Maury and Ricou, 1983; D'Atri et al., 2016). The Subbriançonnais Late Cretaceous-Paleogene flysch formations can be linked paleogeographically both with the Briançonnais domain and with the Provence Pyrenean foreland (Kerckhove, 1965; Blanc et al., 1987; Thum et al., 2015).

9.3. Pre-Alpine paleogeography and western termination of the Alps

There is thus strong evidence to consider the Briançonnais as the eastern termination of the Iberia-Sardinia-Corsica microplate, and the Prepiedmont and Subbbriançonnais domains as transitions towards the Ligurian Tethys ocean and the Valais basin, respectively. However, the paleogeographic pattern of the margin should not be interpreted as linear. The occurrence of lateral transitions from the Briançonnais marginal plateau to the Provence platform westwards (e.g. Decarlis et al., 2017), or towards its eastern termination within the Tethyan oceanic floor (Handy et al., 2010) is likely to have complicated the thrust sequence even in the early stages of inversion. As an example, some specific units displaying late Jurassic platform carbonates, which are observed locally beneath the oceanic Helminthoid flysch nappes (Dumont et al., 2012), must not be interpreted as being derived from the most distal part of the margin but from its lateral transition to the Provence domain. These lateral variations may be either progressive, resulting from oblique opening and/or a scissors-shape margin, or sharp, due to the occurrence of continental transform zones (Lemoine et al., 1989). Moreover, the lateral termination of the western Alpine orogen coincides with a flip in subduction polarity of the converging Tethyan lithosphere (Lacombe and Jolivet, 2005; Vignaroli et al., 2008; Argnani, 2009), possibly reactivating an oceanic transform zone (Dumont et al., 2011, and refs therein). This feature, which may have localised the western termination of the Alps near to the end of the south-dipping slab, may also explain why the Helminthoid Flysch oceanic sediments, which were deposited near the European paleomargin and thrust northwards or NW-wards (Merle and Brun, 1984; Marroni et al., 1992; Mueller et al., 2019; Mueller et al., 2020), could escape metamorphism because they had been deposited to the west of this transform boundary, that is in an upper-plate position. Conversely, their time equivalent Schistes Lustrés, deposited over the southward subducting domain in lower plate position, were affected by blueschist metamorphism within the accretionary wedge (Agard et al., 2002).

9.4. Early stages of continental subduction and inversion of the distal margin structures

Following the initiation of subduction at the Adria margin and the intra-oceanic stacking stages (Pleuger et al., 2007), the Adria-Europe continental collision was achieved in two stages due to the occurrence of the Briançonnais continental ribbon belonging to the Iberian microplate (Le Breton et al., 2021, and refs therein). The first one, from the Lutetian (possibly late Paleocene in the easternmost Briançonnais areas; Bucher and Bousquet, 2007) to Priabonian, is not recorded in the European foreland s.s. (Dauphinois-Helvetic domains; Boutoux et al., 2016), apart from lithospheric flexure propagation (Ford et al., 2006) and from peripheral consequences of the Pyrenean foreland propagation from Provence (Dumont et al., 2011). This "Adria-Iberia" collision only affected the Briançonnais marginal plateau together with its transition towards Tethys, the Prepiedmont margin toe (b, d, fig. 15; a, b, fig. 16). S-verging subduction of the margin and northward motion of Adria driven by Africa-Europe convergence (Rosenbaum et al., 2002) was probably facilitated by the thinned character of the Briançonnais crust (Mohn et al., 2010; Le Breton et al., 2021) but the rift uplift geometry near to the necking zone, facing the orogenic wedge propagation, played a major role. Progressive detachment within the sedimentary cover and the upper crust fed the early collision prism (Scheiber et al., 2013; Tavani et al., 2021), whose kinematics were controlled by the northward motion of Adria as demonstrated in the Central Alps (Escher et al., 1993; Schmid et al., 1997). This tectonic stack, whose top-to-bottom stacking order was likely representative of the paleogeographic distribution, pinches out westwards along a sinistral transfer zone presently incorporated and distorted in the southern part of the arc (Ricou and Siddans, 1986; Schmid and Kissling, 2000; Schmid et al., 2017). The incipient stages of inversion in the Briançonnais domain were probably analogous to the present structure of Provence, where Triassic evaporites played a major role as a detachment layer (Bestani et al., 2015), but the topographic surface of the Briançonnais orogen also shows evidence of olistostromes and large-scale gravity induced deposits in a deep flexural basin setting. The thick late Variscan Briançonnais basins, which possibly trended oblique to the propagation considering the late Variscan paleogeography (Guillot and Ménot, 2009; Pfiffner, 2014), were also detached (Zone Houillère) and overthrust by their southern margin (most internal Briançonnais units), allowing the development of a top-to-the N-NW thrust sequence (c, fig. 15). Below the detachments, most

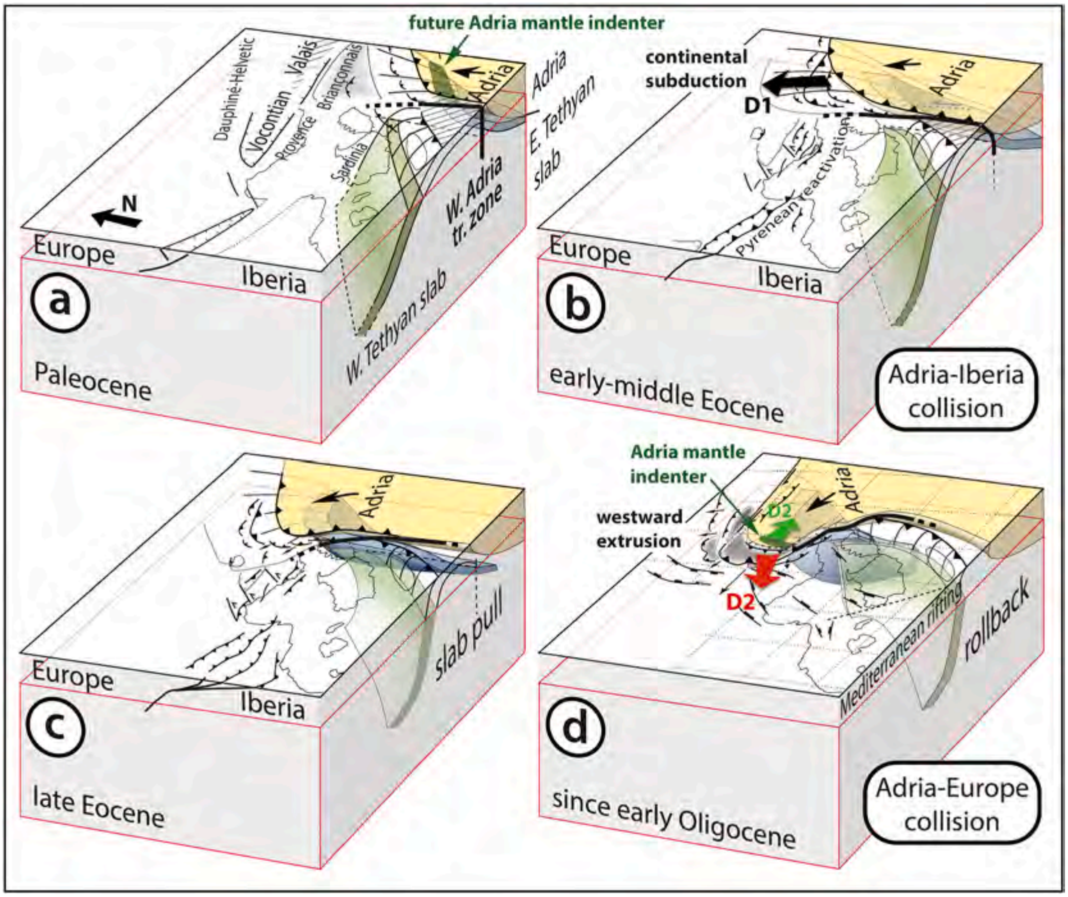

Fig. 16. 4D sketch from the initiation of the Adria-Iberia collision (a) to the westward extrusion atage (d), illustrating the behaviour of the Tethyan slabs and the key role of the subduction flip across the western Adria transform zone. a- The northward drift of Adria (yellow) and the southward subduction of the E-Tethyan slab (blue) are bounded westwards by a transform zone, and the accretionnary wedge is reaching the easternmost margin of the Iberian plate. The northwestern part of Adria lithospheric mantle will become the uplifted Ivrea body in the late stages of extrusion. b- The easternmost part of the Iberian plate (Briançonnais and Pre-piedmont domains, respectively) are undethrust beneath the northward moving Adria plate, leading to detachment, inversion and nappes stacking processes in the subducted upper crust, marked by deformation D1. The Alpine orogen is preceeded by the development of a flexural basin also propagating northwards and pinching westwards along the transform boundary. This stage is also responsible for the progressive closure of the Valais trough. By that time, convergence transmitted through Iberia plate also activates or re-activates the Pyrenean orogen and foreland. c- The N- to NW-directed propagation ceases the end of Eocene, after complete closure of the Valais trough, and a sharp kinematic change occurs with both initiation of extrusion of the northern part of Adria, and development of slab rollback beneath the eastern part of Iberia (green), marked by the western European Cenozoic rifts. d- The westward propagation of the Western Alps, accomodated in the Central Alps by orogen-parallel extension (Simplon fault) and dextral strike-slip faulting (Insubric fault), leads to the formation of the arc with radial spreading deformation. The Internal Zones, containing the accretionary and continental subduction wedges built during the previous stages, are exhumed and pinched between forward and backward structures (D2, red and green, respectively), among which the Penninic Thrust. This stage is driven by indentation at depth by the Ivrea body, a shallow isolated piece of Adria lithospheric mantle in the core of the arc, well identified by geophysics (Malusà et al., 2021). The abrupt southwestern boundary of this lithospheric indenter could derive from the western Adria transform zone having accomodated its northward drift during the previous stages.

of the Variscan and older metamorphic basement was diachronously underthrust beneath the accretion-collision wedge (Bucher et al., 2004; Malusà et al., 2005; Berger and Bousquet, 2008; Strzerzynski et al., 2011; Lanari et al., 2012; Scheiber et al., 2013; Pfiffner, 2014), experiencing HP metamorphism spanning from ~50 Ma to ~35 Ma, and marked by N- to NW-directed tectonic transport criteria (Dumont et al., 2012). The coeval northward drift of the Adria/Briançonnais continental subduction wedge (b, fig. 15; b, fig. 16) caused the closure of the Valais-East Vocontian basins, translated the Adria lithosphere to a higher latitude and shallower depth, and eventually resulting in juxtaposition with the Dauphiné/Helvetic lithosphere.

9.5. Late stages of continental collision

The second part of Alpine orogenic evolution, from the Eocene-Oligocene boundary onwards (Dumont et al., 2012), is dominated by westward extrusion (WNW to WSW) of the previously elevated Adria upper mantle towards the European foreland (c, d, fig. 16), causing both steep subduction of the European lithosphere (Zhao et al., 2015; Malusà et al., 2021) and indentation of the previous continental subduction wedge by the Ivrea body (Schmid et al., 2017), shaping the arc of the Western Alps and exhuming the HP-LT evidence of the early stage. This kinematic stage was accommodated by strike-slip motion and orogen-parallel extension (Mancktelow, 1992; Steck, 2008; Campani et al., 2010; Ring and Gerdes, 2016) with local thermal overprint (Bousquet, 2008a; Wiederkehr et al., 2008) in the Central and Eastern Alps. In the Western arc, radial spreading and forward propagation involving newly formed crustal-scale thrusts facilitated the exhumation of the external foreland until recently (Schwartz et al., 2017). This "modern" Alpine orogenic stage accommodates the Apenninic dynamics and the Corsica-Sardinia breakoff (Gidon, 1974; Laubscher, 1991; Maffione et al., 2008), suggesting that coupled driving forces should be sought in the Mediterranean and Alpine slab dynamics (Jolivet et al., 2008; Vignaroli et al., 2008; Faccenna et al., 2014; Salimbeni et al., 2018; Eva et al., 2020) rather than in Africa-Europe convergence, and should incorporate consideration of gravitational spreading. The "Adria-Europe" collision stage, which is responsible for the most obvious structures of the Internal Zones, in particular the arc and the "Penninic thrust", disturbed the initial stacking order and overprinted the metamorphic record of the internal units. The dome shape of the main internal cristalline massifs (Gran Paradiso, Vanoise, Ambin, Dora-Maira) could partly result from crossed shortening episodes between early and late stages of continental collision, as proposed in the external zone (Dumont et al., 2011). Despite this overprint, we argue that the early stage structures remain preserved at all scales.

9.6. Insights for large-scale 3D structure

The main geophysical transects imaging the finite lithospheric structure in the Western Alps are oriented perpendicular to the trend of the arc (NRP 20: Pfiffner et al., 1997; ECORS: Guellec et al., 1990; CIFALPS: Malusà et al., 2021). Following our interpretation, the geometrical expression of convergence must be different depending on their location and orientation. In the northern part of the arc (NFP 20), the NW-SE to NNW-SSE orientation is able to take into account most of the early stage nappe stacking together with later backfolding. However, such a profile is not adequate to observe the late stage of dextral shear coupled with extrusion. On the other hand, a NE-SW oriented profile in the southern part of the arc (CIFALPS 1) shows a nice expression of the late westward extrusion phase but fails to see the earlier N- to NW-directed nappe stacking. Consequently, any attempt of restoration using such radially oriented lithospheric profiles must follow a sequential approach, first based on the southern profiles to retrodeform extrusion, then considering northern profiles to restore the main part of stacking due to N-NW-ward convergence.

A key feature of our model is the location of the western termination of the Alps along a lithospheric sinistral strike-slip boundary named the "Western Adria Transform Zone" (Dumont et al., 2012; fig. 16a), which was subsequently involved and distorted in the arc (Malusà et al., 2015; Schmid et al., 2017; fig. 16d). Despite the overprint, some relicts of such a feature should still be observed in the present lithospheric structure, which has recently been investigated with increasing resolution (Hetényi et al., 2018; Malusà et al., 2021). The shape of the Ivrea Body indenter, representing Adria lithospheric mantle (Zhao et al., 2020) is now relatively well constrained (Schmid et al., 2017 and refs. therein) and its regular trend is sharply interrupted southwards to the west of Cuneo city. This sharp discontinuity at depth contrasts with the curved form of the Penninic units at surface in the southern part of the arc, which suggests that the latter are decoupled from the indenter. We propose an interpretation of the sharp southern termination of the Ivrea body as an anticlockwise rotated relict of the Western Adria Transform Zone (d, fig. 16). This interpretation is consistent with the occurrence of anticlockwise rotation of the southern Tertiary Piedmont Basin since the Oligocene (Maffione et al., 2008). The shallow location of the Ivrea Body, especially to the south (Lardeaux et al., 2006) could be either inherited from the early stage of Tethyan rifting or from the main exhumation stage of (U)HP continental units in the Eocene (Malusà et al., 2021), bringing the Adria upper mantle and subduction wedge into position before westward indentation. Finally, considering the western boundary of the Alps as inherited from a transform boundary between two opposed-dip oceanic subduction zones allows for an early initiation of the asthenospheric counterflow though the tear zone, subsequently enhanced by Apenninic rollback. Such an asthenospheric counterflow is documented by mantle anisotropy beneath the Alps-Apennines junction (Salimbeni et al., 2018).

10. Conclusion

The occurrence of an early phase of along-strike tectonic transport criteria in the southern part of the Internal Western Alps arc is indicative of an early stage of N- to NW-directed nappe stacking associated with the involvement of the easternmost domains of the Iberia plate (Briançonnais, Prepiedmont) in continental subduction beneath the Adria plate since early Eocene. We propose that this early stage had a major impact on both metamorphic imprint and translation of nappes, through the control of delamination processes within the upper crustal section, and that it is chiefly responsible for "inversion" of not only Mesozoic marginal rift structures, but also of late Variscan foreland structures (Scheiber et al., 2013; Ballèvre et al., 2018). The associated structures were later (early Oligocene onwards) overprinted and distorted during the formation and bending of the arc, and were crosscut by the Penninic Thrust, an expression of the westward extrusion and exhumation of the previously formed continental subduction wedge. Westward extrusion was accommodated in the Central Alps by dextral displacement along the Insubric line as part of the Periadriatic fault zone (Laubscher, 1991; Schärer et al., 1996), by ductile shear and extension along the Simplon fault zone (Mancktelow, 1992; Escher et al., 1997; Steck, 2008; Campani et al., 2010), and by exhumation of the Lepontine dome (Wiederkehr et al., 2008; Steck et al., 2013, 2019). The Western Alps result from a succession of orogenic phases: firstly, during the Eocene, S to SE subduction of the easternmost part of the Iberia plate, namely the Prepiedmont and Briançonnais domains, beneath the Adria plate and the oceanic accretionary wedge; secondly, from the early Oligocene onwards, WNW to WSW extrusion of the previous orogenic wedge over the subducted European plate.

The first stage was accommodated by a major sinistral transcurrent boundary between Corsica-Provence and the Briançonnais, which was possibly inherited from a tear boundary between two opposed-dip subduction areas within the residual Tethyan oceanic domain (a, fig. 16). The development of this early orogenic wedge was controlled by the northward drift of Adria (b, fig. 16), beneath which the Prepiedmont and Briançonnais "distal" continental margin units were diachronously involved, likely activating crustal uncoupling processes similar to those described in Provence (Bestani et al., 2015) and in the Pyrenean foreland (Lacombe and Mouthereau, 1999), together with detachment of the upper Paleozoic and Mesozoic sedimentary cover, partly controlled by evaporites (Michard et al., 2004). The European domains s.s. (Dauphiné-Helvetic) were only lightly affected by this deformation, mainly through reactivation of the Pyrenean-Provence structures at the northern margin of the Iberian plate s.l. However, this N-NW propagation of the early Alpine orogen, with a minimum translation of approximately 200km (Schmid and Kissling, 2000; Ford et al., 2006), accommodated the closure of the eastern part of the Vocontian-Valais basin. This propagation was fringed to the north and NW by the development of a flexural basin, over which the surficial record of this early orogenic wedge is locally preserved (Swiss and French Prealps, Embrunais-Ubaye nappes, Ligurian flysch nappes). It consists of various tectono-sedimentary breccias and olistostromes, sometimes reworking mixed oceanic-continental material in the vicinity of the sole thrust of the lowermost oceanic nappes.

At the initiation of the second stage, close to the Eocene-Oligocene boundary, some parts of lithospheric mantle of the Adria upper plate had been brought to a shallow depth in front of the Dauphiné-Helvetic crust, due to underplating of the Briançonnais crustal elements and in response to its steep transcurrent western boundary (c, fig. 16). Thus, the western Adria lithospheric mantle was suitably located to indent the European crust in the southern Western Alps. This stage was driven by westward extrusion of the northern Adria plate, accommodated by the dextral activation of part of the Periadriatic fault zone (Insubric line) and extension in the Simplon-Lepontine areas. The western Alpine extrusion occurred coeval with rifting and breakup of the Ligurian, then Thyrrenian oceanic domains and with the propagation of the Apennine orogen in a slab rollback framework (Jolivet et al., 2008), suggesting the occurrence of an asthenospheric counterflow responsible for coupling these opposite dynamics (Salimbeni et al., 2018). The expression of extrusion in surface geology consists of exhumation, forward and backward thrust-folding and distortion of the initial stack along the arc, activation of the Penninic Thrust and radial outward propagation of thin- and thick-skinned deformation in the external foreland (d, fig. 16). This largely overprinted the initial structures in the Internal zones, although the amount of horizontal displacement was possibly a lower order of magnitude than during the first stage.

Declaration of Competing Interest

None.

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