Evolution of the Central Pyrenean Mérens fault controlled by near collision of two gneiss domes

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Zusammenfassung:

Die Mérens-Scher- und -Störungszone (MSFZ) ist ein bedeutendes orogen-paralleles Strukturelement innerhalb der variszischen Kernzone der Zentralpyrenäen. Entlang einer Gesamtlänge von 70 km verläuft sie durch Gesteine unterschiedlicher Festigkeit. Die Ausbildung der MSFZ wird beeinflusst durch den Kompetenzkontrast der durchschnittenen Gesteine, sowie der Existenz zweier großer Gneisdome, dem Aston- und dem Hospitaletdom. In Bereichen minimalen Abstandes (ca. 300 m) zwischen den Orthogneiskernen der beiden Dome haben sich mylonitische Bänder in subvertikalen metasedimentären Gesteinen geringer Festigkeit ausgebildet. Westlich der Orthogneise befindet sich eine Übergangszone, in der die MSFZ in eine nördlich einfallende Aufschiebung übergeht, die amphibolitfazielle Glimmerschiefer im Norden von Phylliten mit spröd-duktilem Gefüge im Süden trennt. Der Schersinn innerhalb der duktilen Scherzone und der spröden Störung weist auf eine generelle dextrale, südwärts gerichtete Aufschiebung des Astondoms über den Hospitaletdom hin. Lokale Scherrichtungen stehen unter dem Einfluss von Verformungspartitionierung, befinden sich aber im Einklang mit dextraler transpressiver Tektonik. Im Bereich des westlichen Endes der MSFZ ist die lokale Verformung minimal, wobei sie sich über mehrere Kilometer verteilt und in südvergenten Falten manifestiert. Das zu beobachtende Gefüge weist auf eine kontinuierliche Entwicklung von duktiler Scherung zu sprödem Bruch hin, anstatt zweier unterschiedlicher Deformationsereignisse. Die Ausbildung der mylonitischen Scherzone und der Störungszone ist das Ergebnis der Beihnahekollision zweier großer Gneisdome.

Abstract:

The Mérens shear and fault zone (MSFZ) is a major orogen-parallel structure within the Variscan crystalline core of the Central Pyrenees, passing through rocks of various strength along its 70 km length. The nature of the MSFZ is strongly dependent on the competence contrast of the rock assemblages transected by it, and the presence of two large gneiss domes, the Aston and Hospitalet. A high-strain mylonite zone is developed in narrow subvertical bands of weak metasedimentary rocks where the spacing of the orthogneiss cores of the domes is minimal (ca. 300 m). In a transition zone west of the orthogneisses, the MSFZ grades into a northerly dipping reverse fault, separating amphibolite-facies mica schist with mylonitic bands to the north from phyllites displaying brittle-ductile fabrics to the south. Sense of shear along the ductile shear zone and the brittle fault indicates an overall dextral reverse motion, thrusting of the Aston dome over the Hospitalet dome to the south. Local shear directions respond to strain partitioning and are in accordance to an overall dextral transpressive regime. Near its western termination, local strain is minimal, deformation spread across several kilometres resulted in south-verging folds. Observed fabrics indicate a progression from ductile to brittle deformation, rather than two separate deformation events resulting in ductile shearing followed by brittle faulting. The development of the high-strain shear and fault zone is the result of the near collision of two large gneiss domes.

1. Introduction

The development of fault and shear zones is commonly associated with weakening of rocks by reducing rock cohesion (fault breccia), grain size (mylonitization), or the development of distinct planes of anisotropy, e. g. foliation, mylonitic foliation (Ramsay & Huber 1987, Twiss & Moores 1992). Such zones of relative weakness, compared to adjacent stronger wall rocks, can be reactivated later under favourable stress orientations (Passchier & Trouw 2005). Large-scale (> 10 km) regional fault or shear zones change their expression along strike in response to variations in rock strength and competence contrast. In crystalline core zones of orogenic belts, characterized by multiple deformation and metamorphism, younger thrust faults are most common and easy to recognize. Less obvious, and possibly obscured by younger events, are high-angle fault and shear zones with a long history that may include more than one orogenic cycle. They can possess a complex geometry with co-existing brittle and ductile fabrics that result from either separate deformational events, progressive deformation at different crustal levels, partitioning between zones of different strain, or a combination of these factors. Reconstruction of the deformational history along such fault and shear zones is a difficult task, but crucial for the understanding of the orogenic evolution (Alsop & Holdsworth 2004).

The Pyrenees are an east-west trending mountain range resulting from partial subduction of the Iberian underneath the European plate during the late Cretaceous to late Oligocene, coeval with the formation of the Alps (Verges et al. 2002). This convergence initially created sinistral strike-slip faults along the plate boundaries (e.g. North Pyrenean fault) and subsequently a biverging fold and thrust belt (Fitzgerald et al. 1999). The Axial zone is the crystalline core of the Pyrenees and vestige to an older tectonic period: pre-Mesozoic metasedimentary and magmatic rocks were deformed during the Variscan orogeny. The majority of faults within the Axial Zone have a strong vertical displacement component, and are thrust faults of Variscan or Alpine age. Faults with prominent lateral motion are restricted to high-strain shear zones in the eastern Pyrenees, the Albera and Cap de Creus massifs (Carreras 2001), or regional fault zones developed in the vicinity of gneiss domes of the Central Pyrenees. These steeply dipping fault zones are oriented parallel to the present-day orogenic trend, and display signs of multiple phases of deformation, the youngest having a dominant southerly directed thrust component, corresponding to late Variscan or Alpine thrusts. Their presence at the contact between orthogneisses bodies and metasedimentary rocks reflect the significant competence contrast between strong orthogneisses and weaker schists and phyllites.

One of the most extensive fault zones in the Central Pyrenees is the E–W-striking Mérens shear and fault zone (MSFZ). It can be traced along 70 km from the Noguera-Pallaresa valley in Spain eastward through the French Soulcem valley, along northern Andorra, to the Têt valley, where it abuts the Confluent basin (Fig. 1). The expression and the style of deformation of the MSFZ change along strike and reflect the different rock types it transects, as well as its activity periods. In northwestern Andorra, near its western termination, the fault splays into several smaller strands, separating amphibolite facies schists in the north from lower grade phyllites to the south. In the central part, spectacular subvertical mylonitic bands mark a shear zone that separates the large orthogneiss bodies of the Aston and Hospitalet domes. Along the eastern section, south of the Quérigut massif, the MSFZ is poorly exposed.

Previous researchers (Zwart 1958, Carreras and Cirés 1986, McCaig 1986, Denèle et al. 2008) have focused on specific parts of the fault and shear zone, but without correlation of these different regions our understanding of this major Pyrenean fault zone remains vague. This study is an attempt to develop an evolution model of the MSFZ by comprehensive mapping of a large section (ca. 40 km) of its well exposed western part. The main problems that are being addressed pertain to the along strike variation of deformation, sense and timing of movement, metamorphism and rocks affected. For the first time, the different aspects of the Mérens fault are described and summarized in one comprehensive study, intended not to be the definite model, but rather provide the base for future research.

2. Geological Setting

Elongated mantled gneiss domes, consisting of orthogneiss cores with enveloping amphibolite-facies metasedimentary rocks, are among the most spectacular structures in the Axial zone of the Pyrenees, preserving multiple ductile deformation and metamorphism of the main-phase Variscan deformation (Mezger 2009). The orthogneiss complexes of the Aston and Hospitalet domes are comparable in size (28–34 km length, 9–14 km width) and are composed mainly of monzogranitic augengneisses. Metaleucogranite, metagranodiorite, metadiorite and migmatitic orthogneiss are also common in the western Aston orthogneiss. Intrusions of the protolith are predominantly of middle Ordovician age (ca. 470 Ma, Denèle et al. 2009), while parts of the Aston orthogneiss are vestige to a Cadomian magmatic event (545 Ma, Mezger 2010). The orthogneisses are intruded by granites of uncertain Carboniferous age. Both

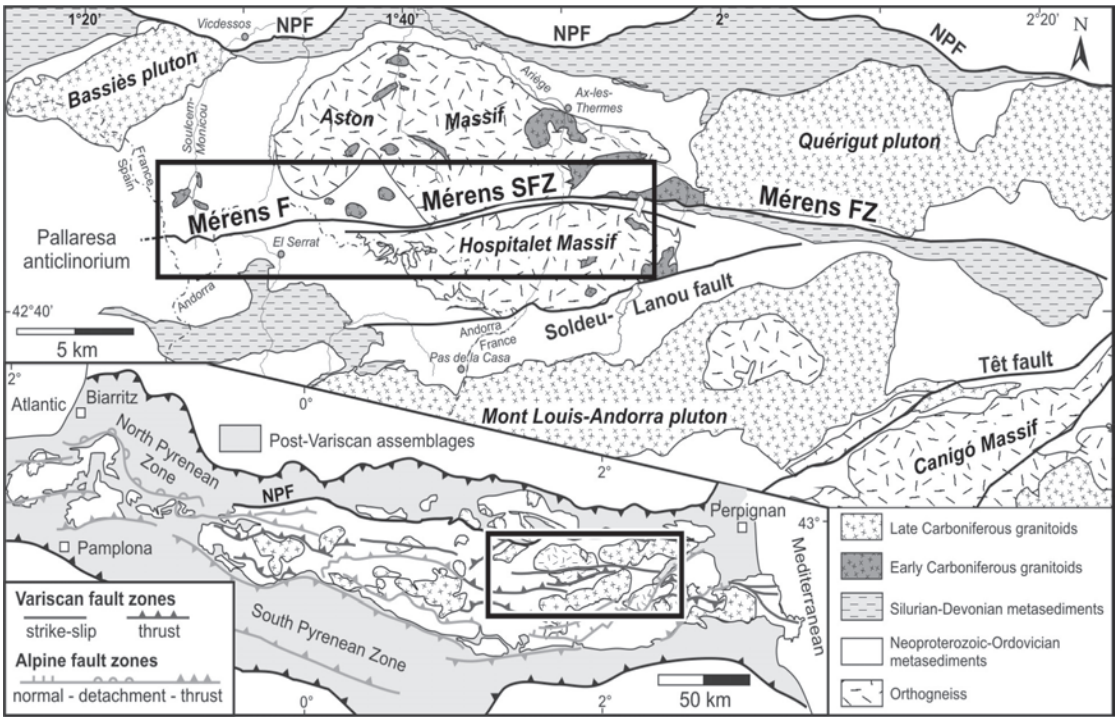


Fig. 1 Geological overview map of the Central Pyrenean Axial zone with location of the Mérens shear and fault zone. The Major faults within the Axial zone and the post-Variscan cover rocks are shown on a simplified geological map of the Pyrenees. Study area is indicated by blue rectangle. NPF denotes North Pyrenean Fault. Modified after van den Eeckhout (1986), Autran & Garcia-Sansegundo (1996) and Mezger (2009).

orthogneisses have a laccolith shape, the thickness of the smaller Hospitalet gneiss is estimated to be 2.5 km (Denèle et al. 2007). At their eastern termination, the gneiss cores dip underneath the metasedimentary mantle, while high-grade mica schists, migmatitic in the Aston dome, structurally underlie the orthogneiss in the west. Similar attitudes of schistosity in metasedimentary rocks and orthogneisses suggest that the main deformation post-dates initial magmatic emplacement in the Ordovician (Mezger 2009). Metasedimentary rocks consist predominantly of metapelites, with minor micaceous quarzites, marble and quartzite layers, of Cambro-Ordovician age (Zwart 1965, Casteras 1969) and possibly latest Neoproterozoic age (Ediacaran, Mezger, unpublished data). Their metamorphic grade decreases from upper amphibolites facies near the contact with the orthogneiss to greenschist facies (mica-chlorite phyllite) with increasing distance.

Formation of the Aston and Hospitalet domes is the result of N–S directed compression, subsequently reducing the distance between the orthogneiss cores to a few hundred meters in the central region (Denèle et al. 2007, Mezger 2009). Within this narrow zone, high-strain deformation of metasedimentary rocks produced mylonites and ultramylonites of the Mérens shear zone (MSZ), subsequently overprinted by the semi-brittle Mérens fault (MF), named after the village of Mérens-les-Vals in the Ariège valley. To the west of the orthogneisses, the shear zone passes through less competent metasedimentary rocks and splays into several strands of mylonite and deformation bands with gradually lower strain, eventually to diminish west of the French-Spanish border (Carreras & Cirés 1986). In these low-strain zones, which also cut fine-grained greenschist facies phyllites, distinction between low-strain shear zones and younger semi-brittle faults becomes difficult. Schnapperelle (2010) named the whole composite structure the Mérens shear and fault zone (MSFZ).

Reverse faults are also developed along the northern and southern margins of the orthogneisses, accommodating uplift of the orthogneiss complexes relative to the adjacent low-grade metasediments. However, only the Soldeu-Lanou fault along southern margin of the Hospitalet orthogneiss forms a distinct fault zone (Fig. 1, Denèle et al. 2007).

3. Previous Studies

Despite its length of approximately 70 km, its good exposure along the main Pyrenean chain (800–2000 m elevation above sea level) and its accessibility through the Ariège and northern Andorran valleys, the MFSZ has not been the subject of many detailed studies. Its major characteristics were described by Zwart (1958) and subsequently published on 1:50,000 scale maps (Zwart 1965, Besson 1991), though the lithology in many sections of the MFSZ is not differentiated. Carreras & Cirés (1986) and Carreras & Debat (1995) have shown that the distinct mylonitic zone of the MSZ in northern Andorra splays into two mylonite bands near the French-Andorran border, passing into several lower strain deformation bands which disappear in the Pallaresa anticlinorium west of the Port de Roumazet (Port Vell) at the Spanish-French border. They relate the deformation bands to intense late folding without significant offset that, consequently, do not represent a fault. Thus, they refute previous studies which postulate a continuation across the Pallaresa anticlinorium further to the Cauterets massifs, 125 km to the west (Oele 1966, Soula et al. 1986).

McCaig (1986) mapped a small area of the MF and adjacent Aston dome in the Rialb valley of northern Andorra. Although the main focus of the study is fluid-rock interaction and metasomatism in shear zones (McCaig 1987), structural analyses suggested that the MF was active together with smaller WNW-trending shear zones (McCaig 1986, McCaig et al. 1990). These mylonites are related to widespread oblique shear zones that cut through all Variscan structures associated with Alpine orogenesis (Soula et al. 1986). ^{40}Ar-^{39}Ar ages of 100-50 Ma from whole rock and muscovite of mylonitic metapelites are interpreted as the timing of movement along the MF, although McCaig & Miller (1986) do not rule out older Variscan movement.

East of Mérens-les-Vals, Denèle et al. (2008) studied mylonitized plutonic rocks (granodiorites to gabbro-norites) within the MSZ, interpreted to have been deformed in sub-solid state during emplacement. The authors concluded that the MSZ served as a feeder zone to a now eroded pluton during the main Variscan magmatic phase.

Most authors agree on reverse motion with relative movement of the northern block (i.e. Aston dome) over the southern Hospitalet dome, with a minor dextral component (Zwart 1958, Carreras & Cirés 1986, McCaig 1986, Mezger 2009), except for Soula et al. (1986) who postulated a major sinistral offset of 10 km based on presumed displacement of metamorphic mineral isograds.

4. Methodology

The incentive for this study is to connect the study areas of Carreras & Cirés (1986) and Denèle et al. (2008) and develop a comprehensive evolution model of the MSFZ by mapping it in its entirety along the best exposed section, 40 km between the Port de Roumazet at the French-Spanish border to the Étang de Naguille in the east. Excellent accessibility allowed continuous mapping, leaving only small gaps (<1 km) due to steep ridges. The 30 km east of the Parc Natural d'Orlu, where the MSFZ passes through regions of relative poor exposure, were excluded from this study. Data were collected in the course of several mapping and diploma projects conducted by the authors from 2000 to 2008. Aside from structural mapping with special emphasis on kinematic analyses, the lithology of the MSFZ rocks was mapped in detail at a scale of 1:25,000 along a 1–2 km wide strip, covering the shear zone rocks and well into the adjacent areas of the Aston and Hospitalet dome not affected by the MSFZ (Rölke 2009, Schnapperelle 2010). Structural and geological data from the adjacent areas are obtained from Mezger (2009). In addition, microstructural analyses from 112 samples were made.

5. The Mérens Shear and Fault Zone (MSFZ)

5.1 Overview

Although the rocks within the narrow high-strain zone between the Aston and Hospitalet orthogneisses are mechanically weaker, the MSFZ does not control the morphology. Instead of E–W trends, the main valleys of Soulcem-Monicou, Seignac, Aston and Ariège are aligned north to south, reflecting the more recent Alpine orogenic and exhumation history. The valleys that run parallel to the MSFZ are less distinct and catch the headwaters at the foot of the main Pyrenean mountain chain, the watershed between rivers flowing to the Atlantic (north) or the Mediterranean (south). Even within E–W oriented valleys, the MSFZ does not necessarily pass through the valley floor, but is located along steep mountain slopes (e.g. Tristaina valley, northern Andorra).

Based on structural characteristics, the mapped 40 km segment of the MSFZ can be divided into three domains. A high-strain shear zone prevails along a distance of 27 km, from the Étang de Naguille to the Port de Soulanet at the French-Andorran border. That section is dominated by mylonitic rocks with a subvertical foliation. The shear zone width ranges from less than 250 m (Laparan valley) to 1.5 km (Mérens-les-Vals, Étang de Naguille, Fig. 2). Following the western termination of the orthogneisses near Port de Soulanet, coexisting brittle and plastic deformation mark a 5–6 km long transition zone where the MSFZ passes through metasedimentary rocks. The width of the transition zone ranges from 30–150 m. Near the Arcalis ski station in the Tristaina valley, the MSFZ is marked by a fault zone (Mérens fault zone, MFZ) that splits into two deformation bands, the major striking westward, and a minor splaying to the southwest, terminating after 3 km at the French-Andorran border (Carreras & Cirés 1986). The main MFZ passes the border at Port de Rat, a narrow zone few tens of meters wide with steep northerly dipping fault planes, tightly folded schists

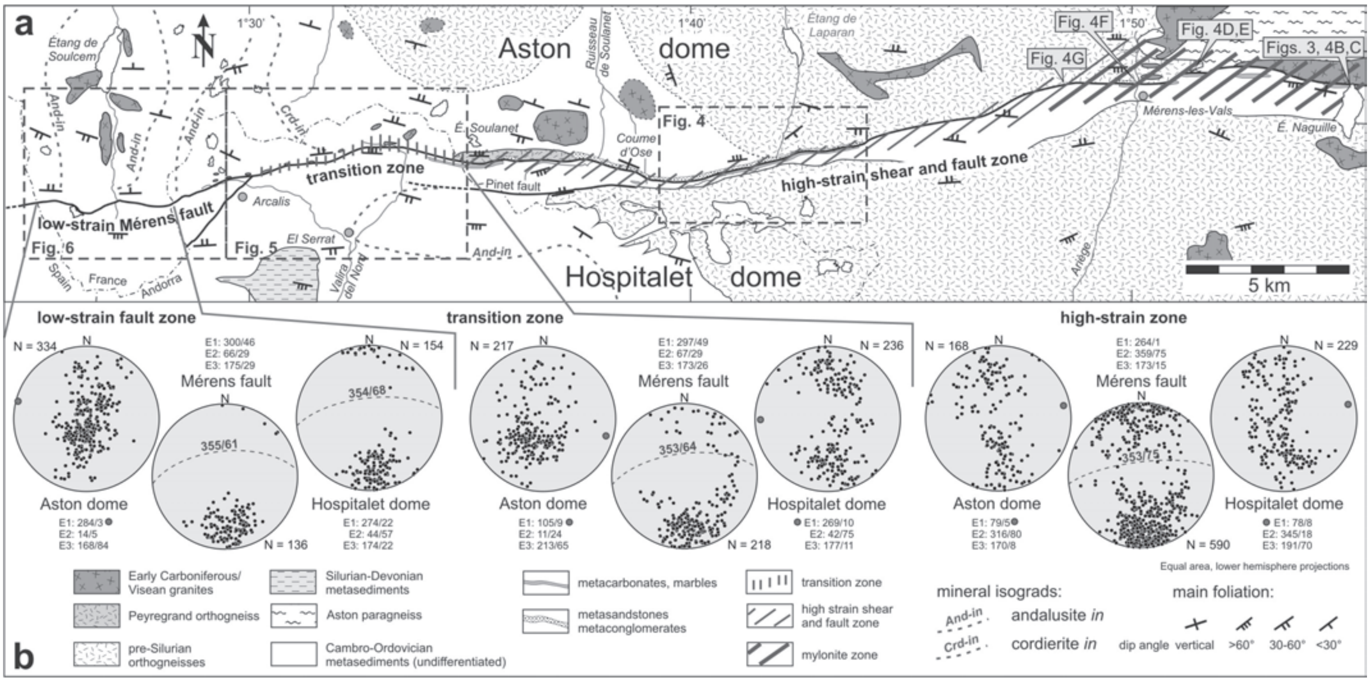


Fig. 2 Geological map of the western part of the Mérens shear and fault zone (MSFZ) subject to this study. Based on observed strain magnitude, the MSFZ is subdivided into three sections, a low-strain, a transition and a high-strain zone. Attitudes of the main foliation and location of selected mineral isograds are shown; B: Equal area stereonets of poles to foliation within the MSFZ and adjacent areas of the Aston and Hospitalet domes in all three zones. E1–3 represent the calculated eigenvectors. Geology modified after van den Eeckhout (1986), Denèle et al. (2008), Mezger (2009), Rölke (2009) and Schnapperelle (2010).

and decimetre-scale duplex structures. Along the final 8 km, strain within the MFZ diminishes gradually and the metamorphic contrast between the metasedimentary rocks of the northern and southern blocks disappears. At its end near Estany de Port Vell, the MFZ is merely reduced to a few deformation bands without obvious offset (Carreras & Cirés 1986).

5.2 Lithology of the MSFZ

Located between the Aston and Hospitalet domes, the lithology of the MSFZ reflects the geology of the adjacent domes. Metasedimentary rocks within the MSFZ include phyllite, quartzite, meta-microconglomerate, metasandstones, metabreccia, meta-arkose, metamarl, calcsilicate and marble. Heterogeneous metacarbonates, varying from pure marble, calc-silicate to marls with interbedded phyllitic layers, form bands up to a thickness of 50 m which can be traced for up to more than one kilometer. Three individual long carbonate layers have been observed at Port de Soulanet, in the Laparan valley and east of Mérens-les-Vals. The eastern band is assigned to the middle to upper Devonian (Denèle et al. 2008, BRGM &d IGME 2010), while van den Eeckhout (1986) correlated western marble bands with the Cambrian Ransol member. Pure quartzite banks are very common in the western Hospitalet dome (van den Eeckhout 1986), but in the MSFZ only a lone metre-thick bed has been encountered in the Rialb valley. Psammitic metasedimentary rocks include meta-microconglomerates with small (<1 mm) clasts of quartz, plagioclase and minor potassium feldspar and muscovite that suggest granitic or granodioritic origin, as well as metasandstones that still preserve individual quartz grains. Metaconglomerates occur in the Laparan valley north of the MF, where they form a prominent 6–7 km long and 50–250 m thick band structurally overlying the orthogneisses of the Aston dome. Isolated small lenses are found in the Rialb valley further to the west, but are probably more widespread, since strongly deformed metaconglomerates are barely distinguishable from mylonitic metagranites with the naked eye. South of the metapsammites, the MSFZ contains fine-grained biotite-muscovite-chlorite phyllites, structurally overlying the Hospitalet orthogneiss. They vary in thickness from tens of meters (western Laparan valley) to more than 2 km west of Mérens-les-Vals. There, however, only the northernmost 500 m lie within the shear zone. The absence of porphyroblasts characteristic of amphibolite facies (e.g. andalusite, staurolite) and the fine-grained texture of less deformed metapelites outside the MSFZ strongly suggest that the metamorphic grade of the phyllites did not exceed greenschist facies. These phyllitic rocks are in strong contrast to migmatitic schists, andalusite and/or sillimanite schists that underlie the western Aston dome north of the MSFZ westward from Port de Soulanet. In the eastern Rialb valley, these coarse high-grade schist are mylonitized and incorporated into the shear zone, easily distinguishable from the adjacent phyllites to the south of the MF.

Magmatic rocks within the MSFZ occur as lenses of orthogneiss and granitic pegmatite varying in size from tens to hundreds of meters in the NW Andorran section between Port de Soulanet and Port de Rat. Their occurrence is restricted to the area north of the MF. In the eastern section, predominantly east of Mérens-les-Vals, calc-alkaline granitoids and mafic plutonic rocks (diorite, gabbro, tonalite) form a 10 km long band interpreted by Denèle et al. (2008) as a deformed magma feeder zone of a now eroded pluton. The only other known occurrence of mafic rocks within the MSFZ west of Mérens-les-Vals is the central Laparan valley, where a dark fine-grained metatonalite forms a lens of possible 100 m width within the Aston orthogneiss.

The main bodies of the Aston orthogneisses have only marginally been affected by deformation along the MSFZ, most notably augengneisses in the Laparan valley and the leucocratic granitic Peyregrand gneiss east of Port de Soulanet. The deformational imprint only extends for a few tens or at most hundred metres. Hospitalet orthogneisses basically lie outside the deformation zone of the MSFZ.

Within the MSFZ, the mineral composition of the metasedimentary rocks has seen an increase in chlorite abundance, primarily due to alteration of biotite, which also produced ilmenite, and muscovite. The new chlorite preferentially grows in shear bands crosscutting the previous schistosity. Chloritization of biotite is not restricted to the MSFZ, but occurs in schists of the Aston dome as far as one kilometre away from the fault zone. In orthogneisses, feldspar is altered to calcite and sericite. McCaig (1987) suggested that metasomatism is related with enhanced fluid flow in the shear or fault zones, accompanied with a significant volume increase.

5.3 The eastern high-strain and mylonite zone: the Mérens shear zone (MSZ)

Squeezed between the orthogneisses of the Aston and Hospitalet domes, the eastern MSFZ is characterized by mylonitic magmatic and metasedimentary rocks forming a shear zone up to two kilometres wide. The maximum width is observed east of the study area, between the southwestern margin of the Quérigut pluton and the Étang de Naguille, where calc-alkaline granitoid rocks have experienced high-strain deformation (Denèle et al. 2008). West of Naguille, the shear zone width is reduced to 1.5 km near Mérens-les-Vals, and to 0.5–1 km up to the termination of the orthogneiss complexes near Port de Soulanet, where the lower strain transition zone commences (Fig. 2). At its narrowest part in the Laparan valley, the Aston and Hospitalet orthogneisses are separated by a zone of metasedimentary rocks less than 300 m wide (Figs. 4A & 7). In the eastern section, up to approximately 1.5 km west of Mérens-les-Vals, the MSFZ affects a significant portion of the Aston orthogneiss. Further west, the majority of the strain is taken up by the metasedimentary rocks within the MSFZ. There, only the southernmost 100–200 m of the Aston orthogneiss has experienced deformation that can be attributed to the MSFZ, while the Hospitalet orthogneisses to the south appear not to be affected at all. In areas where the spacing between the orthogneiss complexes is less than one kilometre, phyllites and meta-microconglomerates, and a minor (5–10 m) marble-marl layer, are completely incorporated in the MSFZ. The brittle Mérens fault passes through the rheological weakest layers, the carbonate band. With increasing space between the Aston and Hospitalet orthogneisses, only metasedimentary rocks north of the fault zone are affected by high-strain deformation. This suggests that the controlling factor for development of the MSFZ location is the southern margin of the Aston orthogneiss.

Structural analyses show that poles to the main foliation within the MSFZ form a broad cluster with a mean around a steep (75°) northerly dipping plane (Figs. 2B & 7). The orientation of the foliation in the adjacent Aston and Hospitalet domes partly overlaps with that of the MSFZ, forming nearly identical girdles around E–W-trending subhorizontal fold axes. They mirror the overall attitude of the main foliation within the two dome structures (Mezger 2009).

The best exposed section of the high-strain MSFZ is located along the shore of the Étang de Naguille, where it cuts through steeply north dipping metasedimentary and adjacent magmatic rocks of the MSFZ (Fig. 3B). The highest strain is recorded in a few tens of meters wide mylonite zone near the dam, at the contact between northern granodiorite and southern phyllites (Fig. 4B). South of the zone of intense shearing, localized mylonitization indicates partitioning into several high and low strain zones. Along the Ariège valley, characteristic rocks of the high-strain MSFZ are mylonitic orthogneisses of the Aston massif with centimetre-sized feldspar augen (Figs. 4 D, E, F). These augengneisses can be traced across the Ariège valley to the upper ski station of the Plateau de Bonascre (Ax-3-Domaines). West of the Ariège valley, gradually

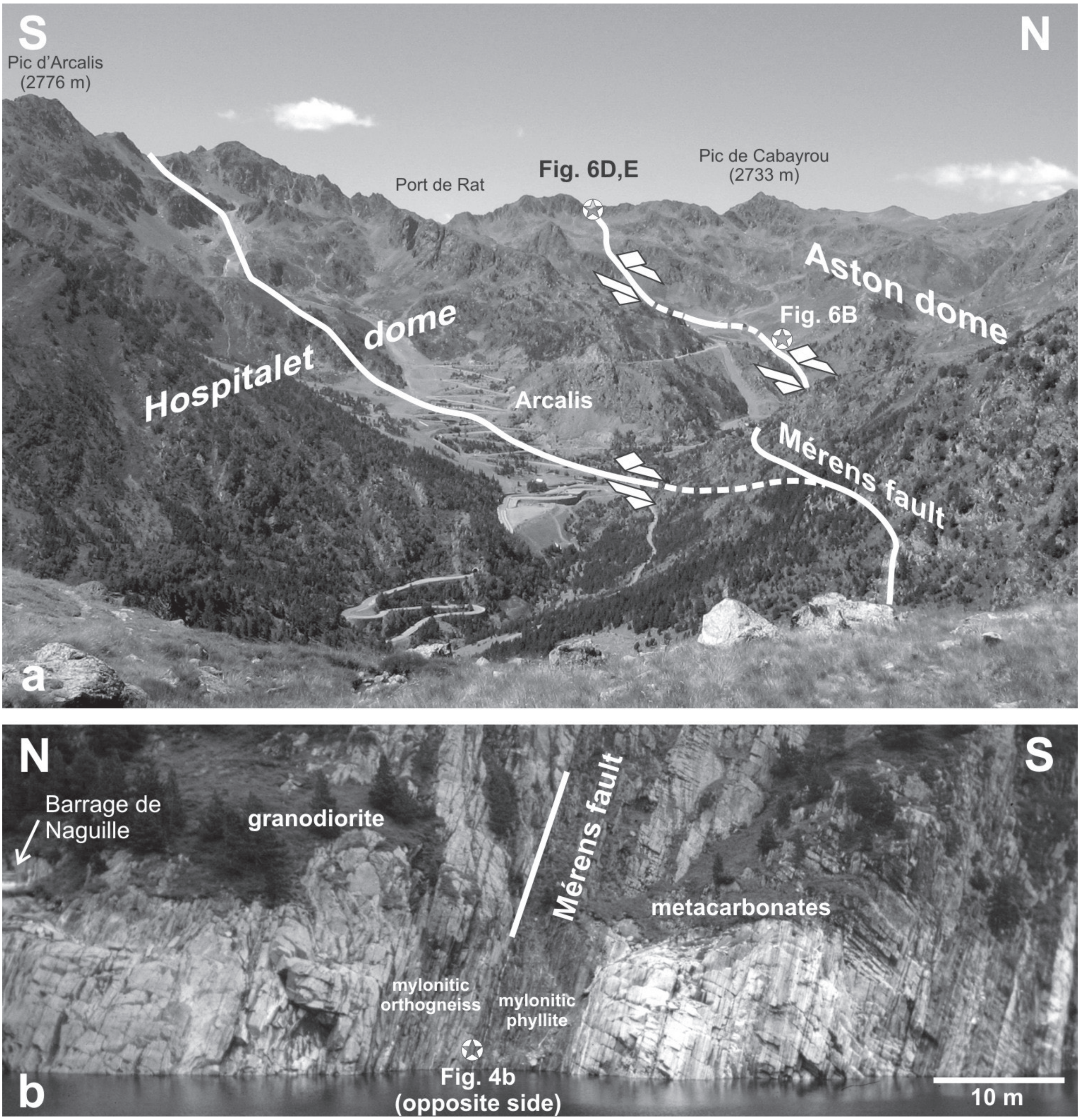


Fig. 3 View westwards across the Tristaina valley in northwestern Andorra towards the border with France at 5 km distance. White lines mark the surface trace of the Mérens fault (MF) and the southwest-striking deformation band passing through Pic d'Arcalis. Arrows indicate top-to-the-south thrusting; B: 80 m wide N–S section of the Mérens shear zone near hydroelectric dam at the northern end of the Étang de Naguille. The MF passes through the centre of the photograph, along the contact between mylonitic granodiorites of the MSFZ in the north and mylonitic phyllites and carbonate to the south. A photomicrograph from a thin mylonitic tonalite band at the base of the orthogneiss is shown on Fig. 4B. The mylonitic foliation is commonly parallel to the lithological contacts, i.e. it has transposed original layering, and dips steeply (> 70°) to the north. Reactivation of the fault zone by brittle stuctures is evident in photomicrographs of Fig. 4C.

Fig. 4 Geological map and rocks from the high-strain MSFZ; A: Detailed structural map of the upper Laparan valley where width of the MSFZ is at its minimum, generally less than 500 m. Most kinematic indicators, slickenlines, show a hanging wall-up, top to the SE motion, i.e. reverse dextral movement of the northern block. However, some fault planes also reveal normal, top to northerly direction, motion. Topography based on the 1:25,000 scale maps Vicdessos and Bourg-Madame (I.G.N. 1992b); B: Thin section scan of a 5 cm thick band of ultramylonitic tonalite from the Mérens fault at the Naguille dam (Fig. 3B). Elliptical sigma-type clasts of hornblende and plagioclase (Pl) indicate dextral sense of shear. Matrix consists of quartz and abundant biotite. Adjacent leucocratic granitic layers are less mylonitised, suggesting that strain was localized in the mica-rich tonalite; C: Detailed photomicrograph from B reveals a millimeter thick breccia layer parallel to the mylonitic foliations, implying reactivation of the older fabric during brittle conditions. Randomly oriented angular fragments of the mylonitic rock are outlined by red arrows. Plane polarized light (PPL); D: Thin section scan of a fine-grained mylonitic granitic augengneiss from the Aston dome near the Mérens fault, one kilometer east-northeast of Mérens-les-Vals; E: Detailed photomicrograph from D zooms in on rounded feldspar clasts mantled by recrystallized feldspar wings with a characteristic delta-type geometry, indicating sinistral sense of shear. Crossed polarized light (XPL); F: Field photograph of a mylonitic two-mica granitic augengneiss with cm-sized feldspar augen (red arrow) from a road outcrop along the RN 20, 500 m north of Mérens-les-Vals. Rock hammer for scale; G: Photomicrograph of an orthogneiss from the Aston dome, Pic de Savis, 2 km west of Mérens-les-Vals. Central feldspar porphyroclast is cut by a sinistral shear band. Red arrows point to domino-type fragmentation of the clast near the fracture. Plastic deformation of quartz crystals in layers within the matrix is indicated by subgrain rotation recrystallization (SGR). Strain experienced by this granite is lower than in D and E (XPL); H: Photomicrograph of a meta-microconglomerate from the Coume d'Ose valley with plastically deformed quartz layers and clasts (SGR) overprinted by younger fold (white line) (XPL); I: Photomicrograph of a crenulated meta-microconglomerate from the western Laparan valley (PPL); J: Close-up of I reveals aligned feldspar clasts and SGR in quartz layers indicative of ductile deformation preceding crenulation folding (XPL).

decreasing strain is evident from a larger average matrix grain size and fractured feldspar clasts (Fig. 4G).

In the western part of the high-strain zone, a narrow microconglomerate band immediately north of the MF records two ductile deformation phases: an earlier high-strain deformation preserved as subgrain rotation recrystallization of quartz and aligned feldspar clasts (Figs. 4H, J). This protomylonitic fabric has been overprinted by younger millimetre-scale tight and southerly verging folds (Figs. 4H, I). Fault planes and slickenlines in adjacent metacarbonates and phyllitic rocks mark the site of the brittle MF. Thin (1 mm) fault breccia zones parallel to the mylonitic foliation of a tonalite at the Naguille dam are the result of brittle reactivation of ductile fabrics within the MSFZ (Fig. 4C).

5.4 The central transition zone

West of Port de Soulanet, the MSFZ leaves the confinement of the Aston and Hospitalet orthogneisses to pass through metasedimentary rocks, mechanically weak phyllites overlain by thin carbonate beds and sillimanite schist to the north (Fig. 5A). In a zone of intense faulting tens of metres wide, thrust planes, duplex structures and northwesterly plunging slickenlines indicate south- to southeast-directed thrusting (Figs. 5B, C). On average, the fault zone foliation dips slightly shallower (64°) northward than in the high-strain zone (Fig. 2B). In contrast, the main foliation of phyllites south of the fault zone and schists to the north has shallower dip angles. Immediately north of the fault zone, a parallel 100–150 m thick protomylonitic band cuts through sillimanite schist, metaconglomerates and granitic lenses along the first 2.5 km of the transition zone up to the Portella de Rialb (Figs. 5A, 7). Observed sense of shear is similar to that of the fault zone, reflecting south to southeasterly thrusting. The overall strain in the protomylonitic zone is less than in the mylonites of the eastern high-strain zone (Fig. 5 D-F).

Several minor faults exist on both sides of the main MF, with similar orientation and sense of motion, e.g. at Port de l'Albeille, east of the Estany Tristaina. Their formation and activity most likely coincides with motion along the main fault, and reflect widening of the deformation zone. These faults are difficult to trace over a wider range, and some may be blind thrusts, such as the one 500 m south of the MF in the Riab valley, that could represent the termination of the Pinet fault (Fig. 7, van den Eeckhout 1986) or a south-verging syncline (Clariana & García-Sansegundo 2009). Near the Arcalis ski station, a several meter wide deformation band splays off the main strand towards the southwest, passing north of the Pic d'Arcalis ending after 3 km near the French-Andorran border (Fig. 3A). Along the valley floor, tight folding, local crenulation foliation and metamorphic contrast –phyllites in the north, staurolite-cordierite schist to the south– are evidence for thrusting.

5.5 The western fault zone: the Mérens fault zone (MFZ)

The main characteristic of the MFZ is a discrete, few tens of meters wide zone of intensely folded and

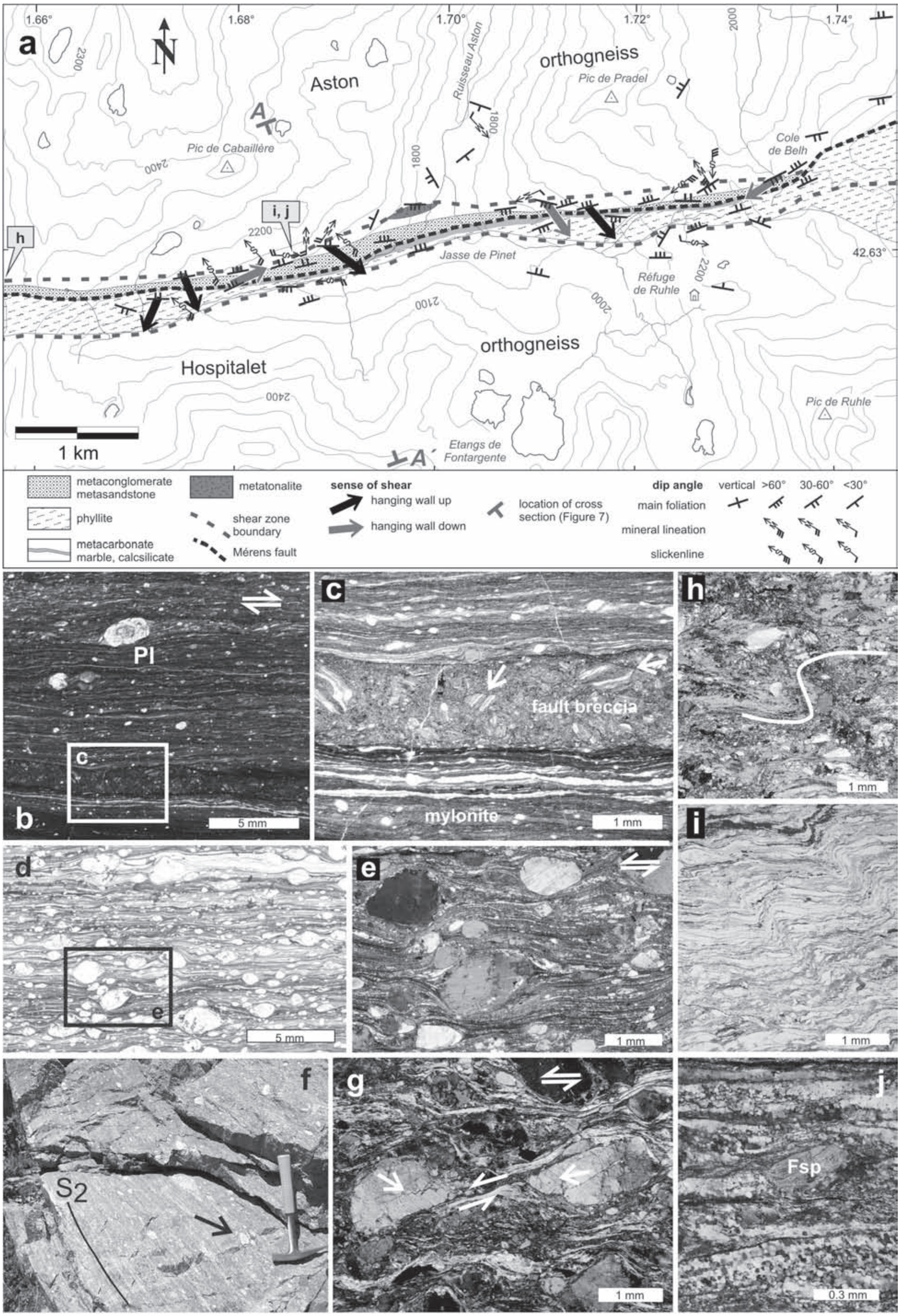

a
Aston
orthogneiss
Pic de Pradel
Pic de Caballière
Cole de Belh
Ruisseau Aston
Jasse de Pinet
Réfuge de Ruhle
orthogneiss
Hospitalet
Pic de Ruhle
Etangs de Fontargente
1 km
h
i, j
A
A´
42.63°
1.66°
1.68°
1.70°
1.72°
1.74°
metaconglomerate metasandstone
phyllite
metacarbonate marble, calcsilicate
metatonalite
shear zone boundary
Mérens fault
sense of shear
hanging wall up
hanging wall down
location of cross section (Figure 7)
dip angle
vertical
>60°
30-60°
<30°
main foliation
mineral lineation
slickenline
b
Pl
c
5 mm
fault breccia
mylonite
1 mm
h
1 mm
d
e
5 mm
1 mm
i
1 mm
f
S2
g
1 mm
j
Fsp
0.3 mm

Fig. 5 Geological map and rocks from the transition zone; A: Detailed structural map with major mineral isograds of the transition zone in the Rialb and Tristaina valleys, northern Andorra. Topography based on the 1:25,000 scale map Vicdessos (I.G.N. 1992a); B: Northerly dipping marble beds overlying phyllites at Port de Soulanet, along the French-Andorran border; C: Duplex structures in the phyllites indicate top-to-the-south hanging wall up movement; D: Photomicrograph of a quartz-rich metaconglomerate in the Mérens shear zone in the Rialb valley. Ductile deformation is indicated by shape preferred orientation of quartz subgrains that define an oblique foliation (XPL); E: Photomicrograph of a metagranite from a metre-sized lens within the Mérens shear zone displays brittle fracturing of feldspar clasts and ductile deformation in horizontal quartz bands (SGR) (XPL); F: Photomicrograph of meta-microconglomerate in the Mérens shear zone of the eastern Rialb valley shows a late brittle shear band, diagonal black line, overprinting plastically deformed (SGR) quartz clasts (PPL).

faulted metasedimentary rocks. Decimetre-scale duplex structures, northerly dipping thrust planes with NW- to NNW-oriented slickenlines and southerly verging tight folds are indicative of southward directed thrusting with a dextral strike-slip component (Fig. 6B, F). North of the fault, andalusite-bearing mica schists possess a moderately northerly dip, while south of the fault, finer grained phyllites dip steeply northwards. The discrete nature of the MFZ is reflected in the tight clustering of the foliation (average attitude of 355/61), compared to the scattering of shear and fault planes in the transition and the high-strain zones. The orientation of the foliation of the Hospitalet phyllites is also more uniform than further east, dipping slightly steeper to the north (354/68). Foliation poles of the schist of the western Aston dome north form a girdle around a horizontal E–W trending fold axis (Figs. 2B, 7). The MFZ cuts mineral isograds of the southern part of the Soulcem contact aureole and the main Aston dome, thus juxtaposing sillimanite-andalusite schist with lower grade phyllites of the Hospitalet dome.

While the attitude of the foliation does not change towards its eastern termination, the overall strain recorded in the MFZ diminishes. Up to the eastern slope of the Soulcem valley, the zone is thrust-dominated, with brittle fracturing of feldspar clasts, plastic deformation of quartz, shear bands in both schist and phyllite (Figs. 6C-E, G, 7). Further west, thrusts give away to southerly verging folds with steeply northerly dipping axial planar cleavages (average attitude: 359/82). Up to a distance of 1.5 km from the MFZ, vertical displacement and strain is distributed over south-verging folds, rather than concentrated on discrete faults. Tight folding, the development of a new cleavage and pressure solution induced by shortening, and transposition of the older fabrics characterize the deformation band west of the Étang de Soucarrane (Fig. 6H,I). Less than one kilometre into Spanish territory, these last vestiges of the MF disappear at Estany de Port Vell (Carreras & Cirés 1986).

The southwesterly striking deformation band and fault zone that splays off the main Mérens fault near the Arcalis ski station in northwest Andorra displays SE-verging tight folds with NW-dipping axial planar cleavage and slickenlines. The direction of displacement, top-to-the-SE thrusting, is similar to that along the main MFZ.

6. Kinematics within the MSFZ

Along the surveyed 40 km, from the high-strain zone at Étang de Naguille to the termination near Estany Port Vell, the average strain within the MSFZ gradually decreases. The overall displacement direction, however, is the same throughout the MSFZ: a relative upward motion of the northern Aston dome onto the southern Hospitalet dome (Fig. 8A). In detail, brittle and ductile fabrics of the eastern high-strain zone differ from those of the western transition and fault zones. Mylonitic foliations of the high-strain zone are subvertical and strike E–W, while the orientation of shear planes in the transition and fault zone is more scattered and possesses shallower northerly attitudes (Fig. 8B). On the other hand, fault planes and slickensides have similar attitudes, dipping steeply in northern directions (ca. 345/75, Fig. 8B). Slickenline and mineral lineation orientations are subparallel, mirroring the differences between high- and low-strain zones: plunging to the NW-WNW (306/65) in the eastern fault zone, and wider scattered around a westerly plunging mean in the high-strain zone (276/57). The mineral lineation and slickenline trends in the MSFZ are very similar to those in the adjacent dome structures. The plunge angles, however, are significantly steeper in the MSFZ (55–80°) than in the domes where they are shallow to subhorizontal (Mezger 2009). Denèle et al. (2008) report similar lineation attitudes for the metaplutonic rocks of the MSZ.

Shear sense indicators in the eastern shear zone mostly record relative upward motion of the northern block, while both hanging wall up and down motion along northerly dipping shear planes can be observed in the western low-strain zone (Fig. 8A). Given the west-northwestern to northwestern attitudes of most stretching lineations and slickenlines in the eastern part of the MSFZ, a top-to-the-SE motion implies a dextral component to thrusting, as noted before by Zwart (1958), Carreras & Cirés (1986), McCaig & Miller (1986) and Mezger (2009).

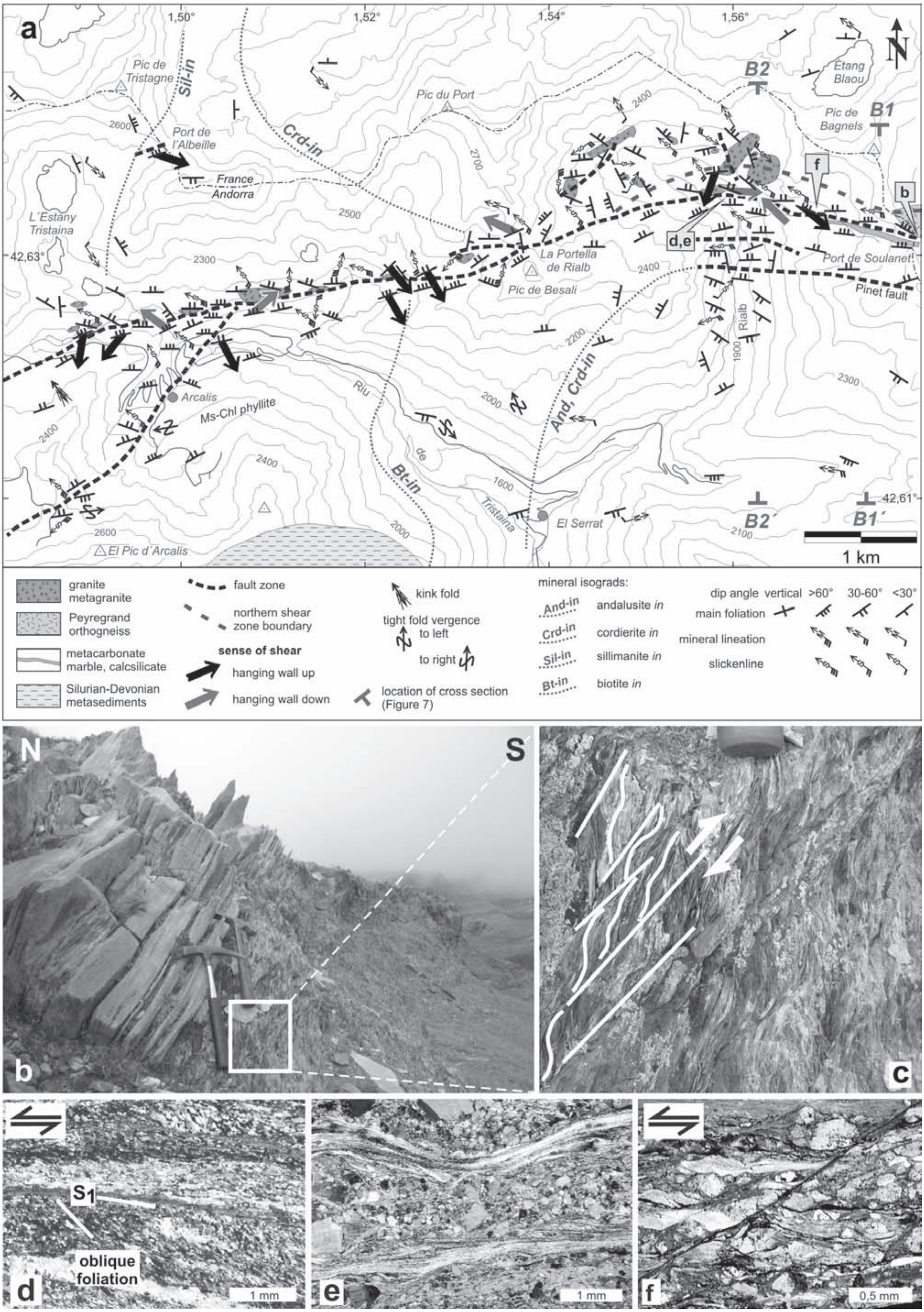

a
1,50°
1,52°
1,54°
1,56°
N
Pic de Tristagne
Sil-in
Port de l'Albeille
France
Andorra
Crd-in
Pic du Port
L'Estany Tristaina
42,63°
La Portella de Rialb
Pic de Besalí
Étang Blaou
Pic de Bagnels
B2
B1
f
b
d,e
Port de Soulanet
Pinet fault
Rialb
Arcalís
Ms-Chl phyllite
Riu
de
Tristaina
And, Crd-in
Bt-in
El Serrat
El Pic d'Arcalís
B2´
B1´
42,61°
1 km
granite metagranite
Peyregrand orthogneiss
metacarbonate marble, calcsilicate
Silurian-Devonian metasediments
fault zone
northern shear zone boundary
sense of shear
hanging wall up
hanging wall down
kink fold
tight fold vergence
to left
to right
location of cross section (Figure 7)
mineral isograds:
And-in
andalusite in
Crd-in
cordierite in
Sil-in
sillimanite in
Bt-in
biotite in
dip angle vertical >60° 30-60° <30°
main foliation
mineral lineation
slickenline
N
S
b
c
S1
oblique foliation
d
1 mm
e
1 mm
f
0,5 mm

Fig. 6 Geological map and rocks from the western low-strain zone of the MSFZ; A: Detailed structural map of the Tristaina and Soulcem valley, where the MF is expressed as a steep reverse fault with predominant top-to-SSE motion. Topography based on the 1:25,000 scale map Vicdessos (I.G.N. 1992a); B: Folded granite pegmatite in the upper Tristaina valley, north of the Mérens fault. Adjacent mica schist possesses second order folds and an axial planar cleavage, which is also poorly developed within the granite. Folds and new cleavage verge to the south. Yellow notebook (19 cm) for scale; C: Photomicrograph from B reveals brittle fracturing of feldspar, while subgrain rotation in quartz bands is indicative of plastic deformation (XPL); D: Port de Rat, immediately south of the MF: thin section scan of a fine-grained phyllite with late shear bands; E: Port de Rat, immediately north of the MF: photomicrograph of a coarser grained garnet-bearing mica-quartz schist outlines the metamorphic contrast to the phyllites south of the fault. A continuous S1 foliation defined by recrystallized quartz grains is cut by younger shear bands (S2) formed by fine-grained white mica (XPL). Sense of shear is identical in D and E, and related to motion along the MF; F: Outcrop photograph of a mica schist in the MFZ at the eastern Soulcem valley. South-verging tight and sheared-off fold and duplex structures indicate top-to-the-south motion; G: Thin section scan of the mica schist at F shows late shear bands overprinting plastically deformed (SGR) quartz layers (XPL); H: Field photograph taken at Port de Roumazet (Port Vell) along the French-Spanish border, one kilometer east of the termination of the Mérens fault. Here, the only trace of the MF is a deformation band characterized by tight folding of quartz-rich mica schist with prominent new subvertical foliation. Scale bar equals 8 cm; I: Photomicrograph from H reveals a distinct spaced S2 schistosity developed as an axial planar cleavage.

7. Discussion

7.1 Deformation variation along strike

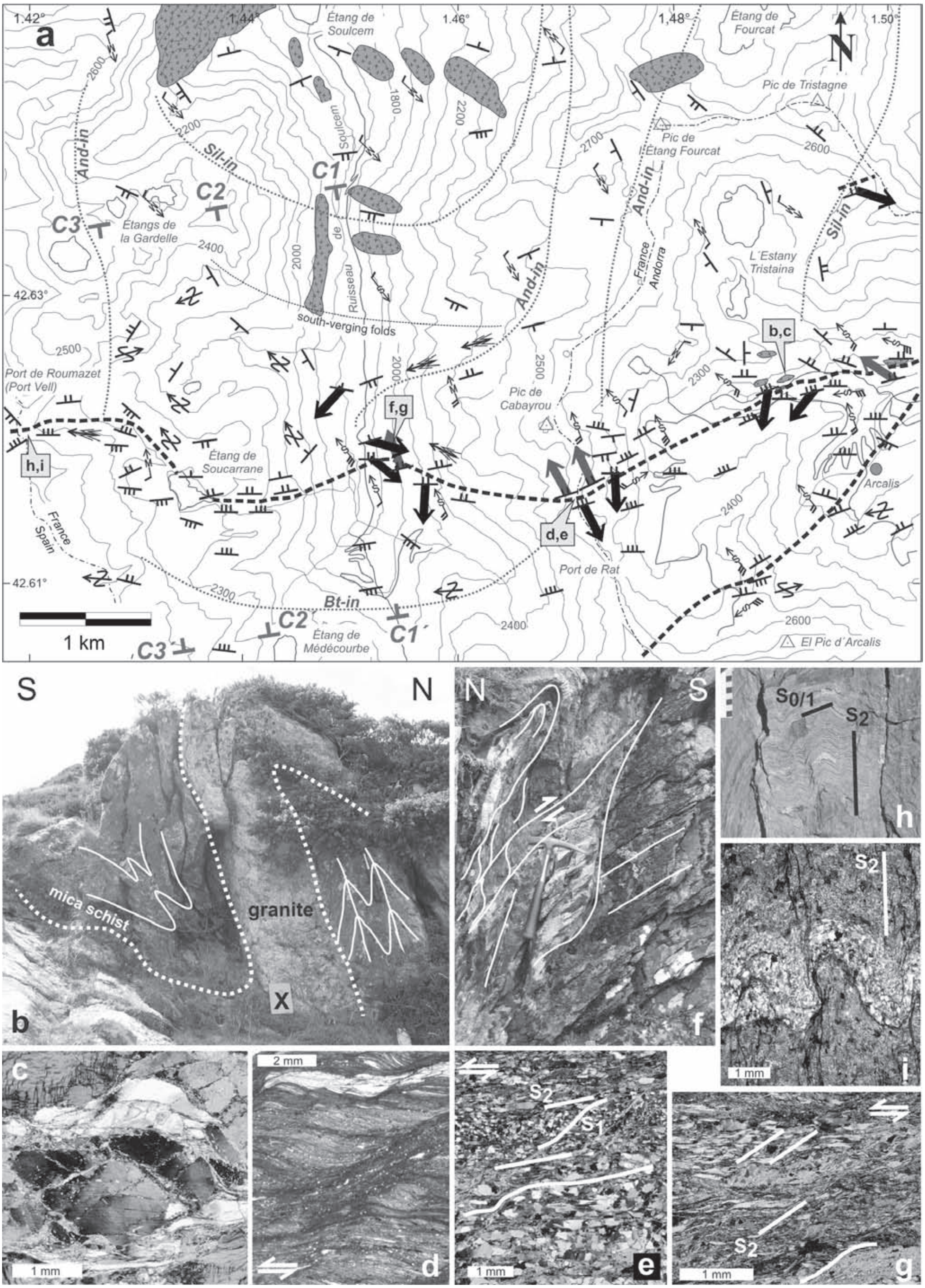
Rock strength, competence contrast and the shape of the gneiss cores control the development of the MSFZ. High-strain deformation is prevalent where the orthogneiss cores of the Aston and Hospitalet domes are in close proximity. The maximum difference in rock strength is reflected in mylonitization of narrow bands of weak metasedimentary rocks wedged between strong orthogneisses, well exposed in the Laparan/Jasse de Pinet section (Fig. 4A). East towards Étang de Naguille, however, an increasing volume of magmatic rocks, mostly calc-alkaline granites, diorites and gabbros, is involved in high-strain deformation. According to Denèle et al. (2008), gabbroic and successively more felsic magma intruded along large tension gashes in a dextral transpressive regime, forming the root zone to a now completely eroded pluton. Periodically ascending magma is thought to have enhanced movement along the fault, at the same time thermally weakening adjacent older Aston orthogneiss, so that it experienced plastic deformation throughout a wider zone (ca. 1 km) than elsewhere along the MSFZ. High-strain deformation is concentrated in the northern part of the MSFZ, close to and partially incorporating the Aston orthogneiss, while the Hospitalet orthogneiss is exempt. The concave shape of the northern Hospitalet orthogneiss margin west of Mérens-les-Vals served as a strain shadow for weak metasedimentary rocks, which retained their low-strain deformation (Fig. 9).

Overall strain decreases as the MSFZ passes west out of the narrow confinement of the orthogneisses. While the Hospitalet orthogneiss gives way to softer phyllites, the western Aston dome is underlain by relatively strong migmatitic mica schists. In the transition zone, competence contrast is less than in the high-strain zone, but enough to produce thin mylonitic bands. Further reduction of the competence contrast occurs westwards as the metamorphic grade in the Aston dome decreases and migmatitic gneiss is succeeded by weaker andalusite schist. As a result, deformation is spread along branching faults and deformation bands and southerly verging folds across a width of more than two kilometres.

The brittle MF which has overprinted ductile fabrics shifts its site relative to the shear zone, and is variably located to the north, south or within the previously developed shear zone. Its location is controlled by the strongest competence contrast, e.g. at the base of the Aston orthogneiss, or the presence of mechanically weak rocks, kilometre-long marble layers, in general following a path of least resistance.

7.2 Metamorphic grade within the MSFZ

Chloritization of biotite, sericitation of feldspar, nucleation of chlorite in shear bands and fractures indicate that plastic and brittle deformation within the MSFZ took place under greenschist facies conditions. Fine-grained phyllites within the eastern higher strain zone near Naguille show that they were never subjugated to amphibolite facies metamorphism. Within the MSFZ, there is no indication of mineral assemblages, neither fresh or retrogressed, that can be related to deformation under amphibolite grade conditions. Cretaceous-Paleogene ^{40}Ar-^{39}Ar ages of mylonites either indicate post-Variscan deformation, or more likely, loss of argon caused by surpassing the closure temperatures in biotite and muscovite at 400-500°C (McCaig 1986). These observations refer to the MSFZ exposed at the present surface. The fact that sillimanite isograds of the western Aston dome are cut by the MSFZ implies that it affects


a
1.42°
1.44°
1.46°
1.48°
1.50°
42.63°
42.61°
N
Étang de Soulcem
Étang de Fourcat
Pic de Tristagne
Pic de Étang Fourcat
Étangs de la Gardelle
L´Estany Tristaina
Ruisseau de Soulcem
south-verging folds
Port de Roumazet (Port Vell)
Pic de Cabayrou
Étang de Soucarrane
France Spain
France Andorra
Arcalis
Port de Rat
Étang de Médécourbe
El Pic d´Arcalis
And-in
Sil-in
Bt-in
C1
C2
C3
C1´
1 km
b,c
f,g
h,i
d,e
S
N
mica schist
granite
X
b
f
S0/1
S2
h
i
1 mm
2 mm
c
d
e
g
S1

the middle crust as well. Denèle et al. (2008) inferred from microtextures that emplacement and mylonitic deformation of calc-alkaline magmatites in the eastern MSZ occurred under amphibolite facies conditions.

7.3 Ductile versus brittle deformation

Fault planes within marble layers display well-developed slickenlines throughout the whole length of the MSFZ. Field observations and microstructural analyses reveal that brittle deformation post-dates ductile (e.g. Naguille mylonite, Fig. 4C). Due to lack of directional data it is difficult to establish temporal correlation between brecciation of mylonites and faulting in adjacent weaker rocks. Similar structural attitudes facilitate a genetic relation between crenulation folding of competent rocks and macroscopic folding and faulting of weaker rocks. Furthermore, parallelism of brittle and ductile linear structures within the MSFZ points to similar stress orientations (Fig. 8B). An observed progression from high-strain ductile, low-strain ductile to brittle in the Laparan/Jasse de Pinet and Rialb areas furthermore corroborates temporal proximity (Figs. 4A, 5A). Relative strong microconglomerates are mylonitized, preserved in subgrain rotation and bulging recrystallization, and subsequently folded before static recrystallization obliterated the high-strain fabrics (Figs. 4H, I). Bordering marbles and phyllites are tightly folded or have developed duplex structures, indicative of semi-brittle deformation (Figs. 5C, D). Thus, mylonitization and faulting could have been part of the same deformation phase. In the western fault zone, foliation development can be related to folding of metapelites, indicated by virtual identical means of the orientation of the axial planar cleavages and the foliation (ca. 355/63, Figs. 2B, 8B).

7.4 Shift in direction of motion

Despite a wide scattering of data points, a near 60° counter-clockwise shift of the shortening direction from NNW in the western fault to 280° in the central shear zone can be inferred from the mean azimuths of ductile and brittle linear structures (Fig. 8A). Taken the E–W-trending fold axis related to the MF in the Soulcem valley into account, the total shift in shear direction amounts to 80°. This can be explained in terms of deflection of flow and strain partitioning due to approaching and a near collision of two large rigid bodies, the orthogneiss massifs of the Aston and Hospitalet domes (Figs. 9, 10). A compressive regime with a small dextral transpressive component forms southerly verging E–W-trending folds in homogeneous metapelites of the western Aston and Hospitalet domes. (Fig. 8B). Thrust direction along the MF changes from southwards in Soulcem valley to southeastwards in the Tristaina valley to ESE at the Port de Soulanet, thus turning into transpressive with a strong dextral component, as a response to the approaching of strong migmatitic paragneiss from the north. In the central region, where the hard orthogneiss cores almost come in touch with each other and are only separated by a thin metasedimentary band, dextral strike-slip with a strong vertical component is dominant. The steepening of the attitude of the foliation in the MSFZ, from an average 60° to the north in the Soulcem valley to 75° in the central high-strain zone, can also be interpreted as the result of near collision of two massive rigid bodies. If shortening is assumed to be homogeneous along the MSFZ, concentration along discrete narrow zones, such as the Laparan valley, requires a larger vertical offset along the MSFZ than in the western area, where strain is spread over several kilometres.

7.5 Displacement along the MSFZ

Most studies agree on the overall reverse dextral, top-to-the-SE motion along the MSFZ (this study, Zwart 1958, Carreras & Cirés 1986, McCaig 1986, Denèle et al. 2007, Mezger 2009). Due to the lack of cut-off markers, quantification of the offset is a major problem. Save for the western termination at Estany de Port Vell where the MFSZ dies out and offset approaches zero, the majority of the MSFZ separates two dome structures with their distinct geological evolution paths.

Soula et al. (1986) inferred a sinistral offset of 10 km, based on apparent displacement between isograds in the western Aston and Hospitalet domes. In the light of recent detailed mapping of metamorphic isograds (Mezger, 2005), this estimate is prone to error, since mineral isograds associated with the Soulcem contact aureole cannot exactly be correlated with isograds of the Hospitalet dome. McCaig (1986) proposed a southward throw of 1–8 km based on a postulated shallow northward dip and spacing of isograds observed elsewhere in the Axial Zone, as well as temperature differences between mylonitic assemblages in the Aston and Hospitalet domes of 100–150 °C. Horizontal displacement is expected to be in the same range.

Although this study recognizes the shortcomings of previous offset estimations, the tectonometamorphic evolution, in which the formation of large antiformal structures of the Aston and Hospitalet domes distorted the orientation of the early Variscan mineral isograds prior to motion along the MSFZ, precludes more precise and correct determination of offset. A recent thermobarometric study on metapelitic rocks and orthogneisses yields pressures of 3.2–4.0 kbar in the Aston and 3.2–3.6 kbar in the Hospitalet dome (Mezger, unpublished data). The strongest metamorphic contrast is observed in the northern Tristaina valley near Arcalis, Andorra, where coarse

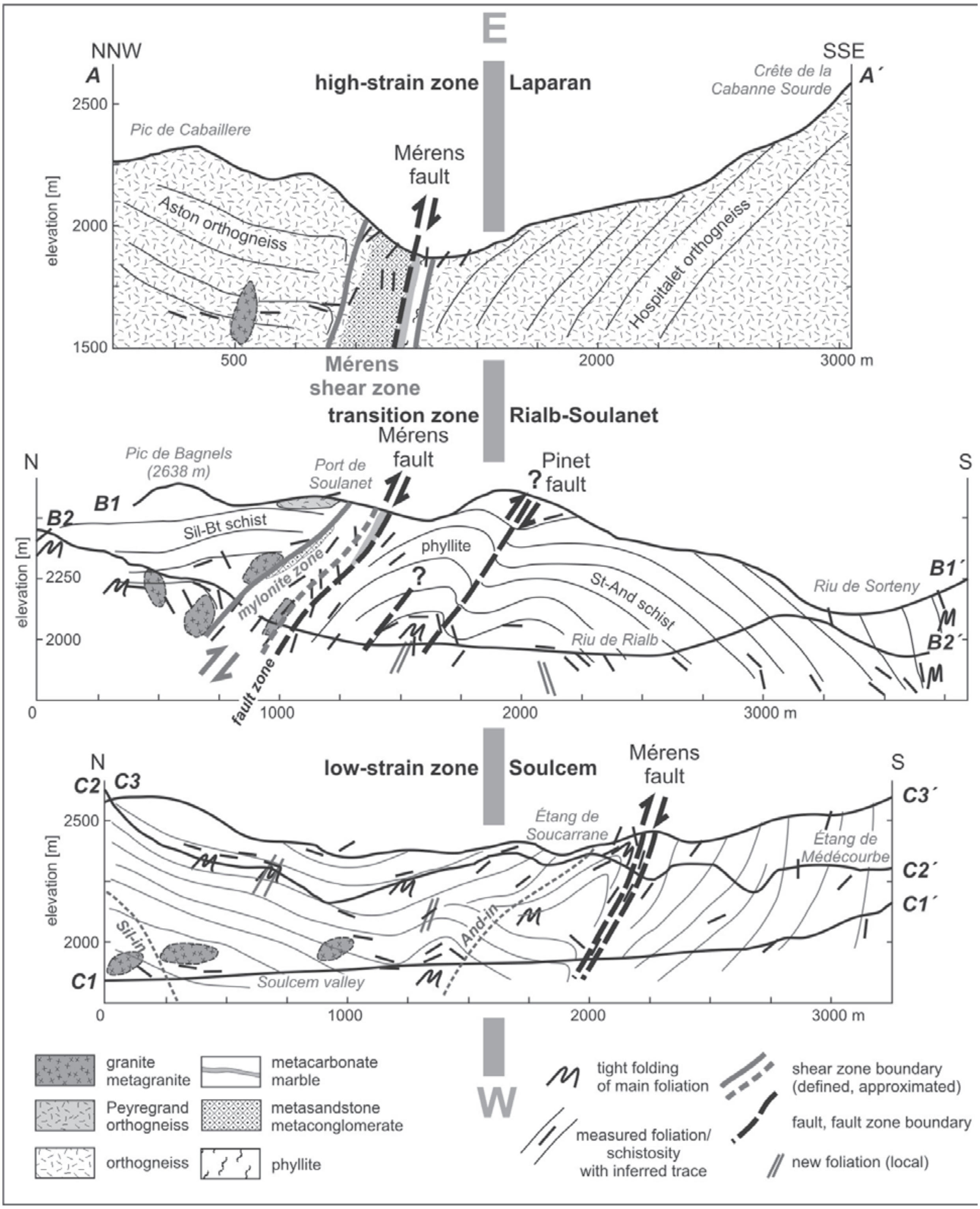


Fig. 7 Three cross sections passing through the MSFZ in the three different strain zones. The sections are compiled from different parallel section running through valley and ridges. Thus, the structural and lithological character throughout a vertical height of 500–700 m could be constructed. Location of the sections is indicated in Fig. 5 - Fig. 7.

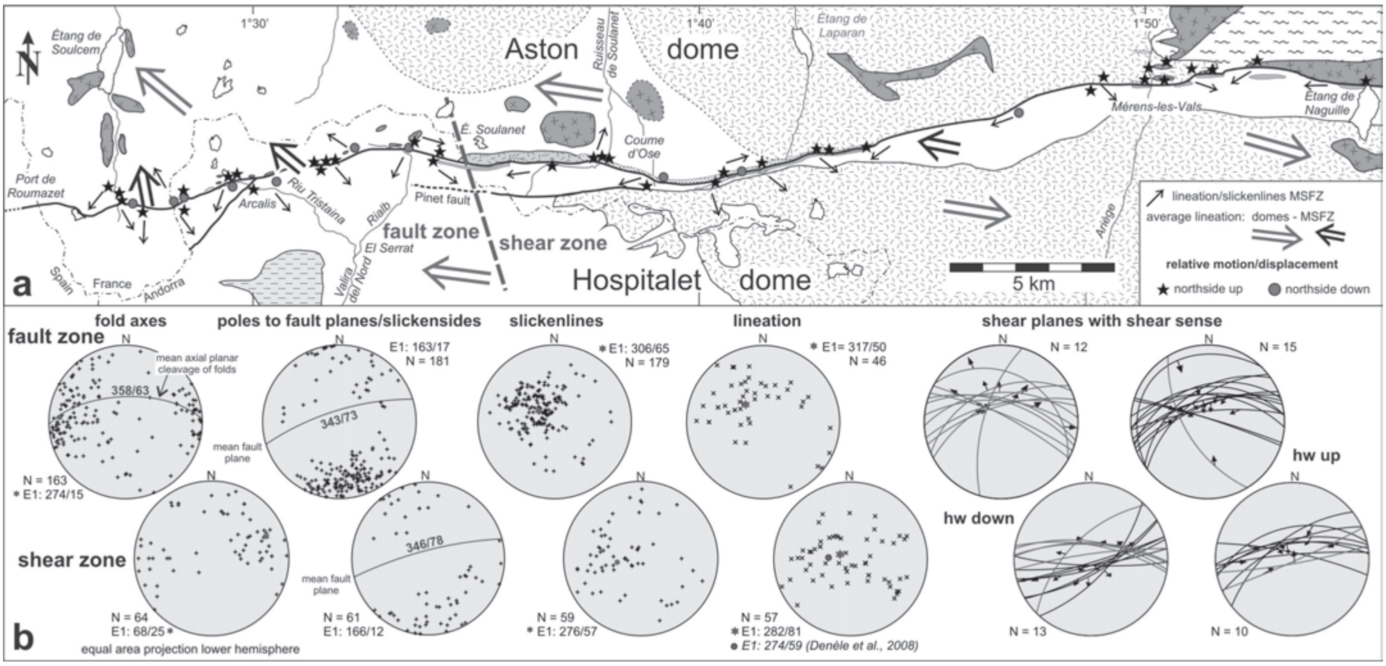


Fig. 8 Simplified geological map of the MSFZ showing relative motion of northern and southern hanging wall blocks. Note that the transition zone of Fig. 2 is assigned to the fault zone. Small arrows indicate local lineation, large arrows show average trend of lineation in Aston and Hospitalet domes (grey) and within the MSFZ (black); B: Stereographic projections of fold axes, slickenside poles (individual fault planes), slickenlines and mineral lineations within the eastern (shear zone) and western (fault zone) MSFZ. Shear sense indicators show brittle and ductile deformation combined as lineation (mineral lineation or slickenline) with arrow pointing in direction of shear on shear or fault plane. Blue planes indicate relative upward motion of northern block, while red great circles are associated with relative downward displacement of northern block. Geological signatures are the same as in Fig. 2.

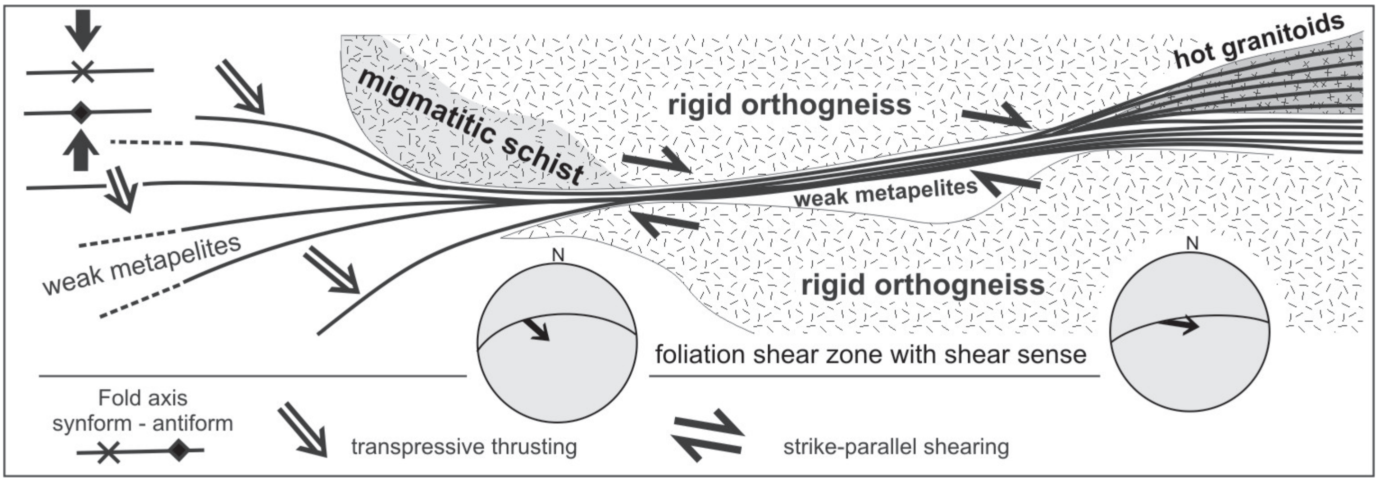


Fig. 9 Summary of main structural observations interpreted as coeval development of NNW–SSE and WNW–ESE linear fabrics. Strain magnitude in MSFZ is indicated by spacing of black lines. Stereonets display average foliation in MSFZ and mineral lineation/slickenlines from Fig. 2B and Fig. 8B.

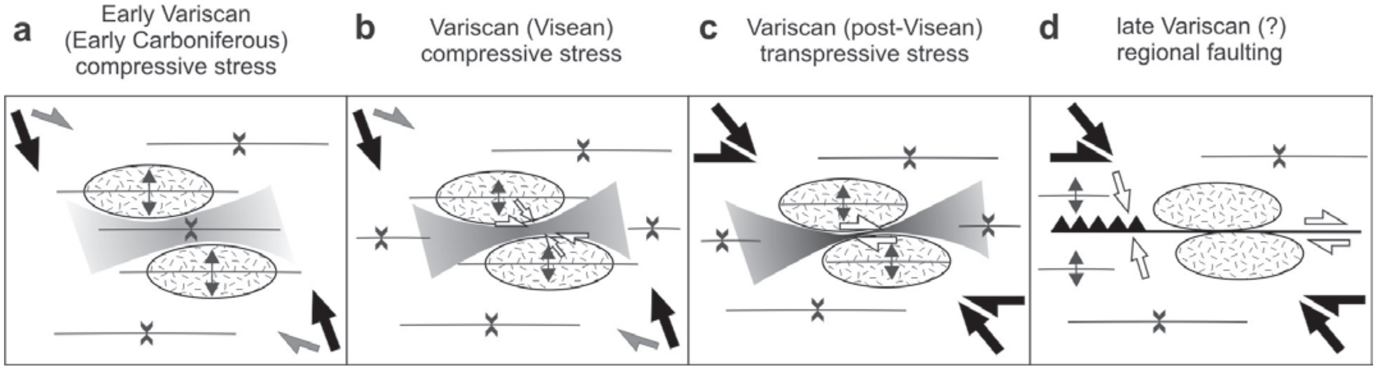


Fig. 10 Four-stage evolution model of the Mérens shear and fault zone following the initial Ordovician emplacement of the granitoid laccoliths. Darker grey shade reflects higher strain; A: Early stage of dome development during dominant compressive stress in early Carboniferous; B: High-strain zone begins to form between approaching orthogneiss cores in the Visean, local strain partitioning; C: Overall stress changes to transpressive resulting in strong strain partitioning with dextral displacement in narrow mylonite zone during the late Variscan main phase deformation; D: Late Variscan or Alpine folding and localized faulting at shallower structural levels, reactivation of older ductile fabrics.

sillimanite schists (620 °C at 3.5 kbar ≈ 11.5 km) are juxtaposed with fine-grained muscovite-chlorite phyllites (≤ 400 °C) to the south. Taken van den Eeckhout's (1986) estimation of a geothermal gradient of 72 °C/km for the western Hospitalet dome into consideration, this would amount to a vertical difference of 5–6 km. This offset is based on several assumptions, e.g. correct geothermal gradient and temperature estimation for the western Hospitalet dome, and small error in geobarometry of the western Aston schist. If the average attitude for the slickenlines and stretching lineation (280–310/50–80) had been the same during time of movement along the MSFZ, the dip-slip and strike-slip components would be in the same order of magnitude, and a dextral displacement not more than 5–6 km can be expected. These assumptions match the proposed displacement of McCaig (1986).

7.6 Dating movement along the MSFZ

McCaig & Miller's (1986) attempt to date mylonitic fabrics of the MSFZ with the ^{40}Ar-^{39}Ar method did not yield conclusive results. The postulated Alpine age (100-50 Ma) of mylonitic deformation does not exclude Variscan activities. Several studies dated micas from narrow 110°-striking mylonite zones, and correlated the obtained late Cretaceous ages with motion along the main MSFZ (Majoor 1988, Maurel 2003). Similar ^{40}Ar-^{39}Ar-ages were obtained from micas in ESE-striking shear zones of the Millas granite plutone, immediately east of the Quérigut massif (Monié et al. 1994). However, the genetic relation between these oblique mylonite zones and the MSFZ is doubtful, as they cut all Variscan structures, including the MSFZ (Soula et al. 1986, Denèle et al. 2008). More recent studies restrict Alpine activity along the MSFZ to thrust faulting (Denèle et al. 2008). Paleocene (60-55 Ma) southward thrusting is also proposed for the ENE-striking Têt fault separating the Canigó massif and the Mont-Louis pluton (Fig. 1) according to a recent thermochronological study (Maurel et al. 2008).

Unfortunately, no magmatic rocks crosscut the MSFZ, but granite sills affected by the MSFZ exist, thus providing a minimum age for plastic deformation. While none of the deformed pegmatites have been dated, a granite from the core of the Soulcem granite suite, of which the folded granitic sill in the upper Tristaina valley (Fig. 6B) is likely related to, has a Visean intrusive age (339 Ma, Mezger 2010). The same study has obtained a similar age from granitic sill near the Bossòst fault, 60 km further west, whose folding is linked to the late main-phase Variscan formation of the Bossòst dome (Mezger & Passchier 2003). Furthermore, the overall attitude of folds throughout the Aston and Hospitalet domes corresponds to that of folds close to the MSFZ, suggesting that initiation of deformation along the shear zone occurred in the later stages of Variscan dome formation (Carreras & Cirés 1986, Denèle et al. 2007). New geochronological data by Mezger (2010) suggest that activity along the MSFZ may extend back to the early Carboniferous.

8. Conclusions

Crucial for the development of the MSFZ is the progressive evolution and approach of the Aston and Hospitalet gneiss domes (Fig. 10). Metasedimentary cover rocks are wedged between strong orthogneiss cores, while their original foliation and layering has been rotated or transposed into steep northerly dips. Movement and finite strain along the MSFZ is strongly dependant on the strength and the competence contrast of rocks affected by it or within close vicinity (Fig. 10B-D). Thus, local shear directions respond to strain partitioning and are in accordance to an overall dextral transpressive regime that lasted through the main Variscan deformation phase (Fig. 10B, C). Juxtaposition of strong versus weak rocks enhanced the development of a discrete high-strain shear and fault zone with dextral, north side up thrusting (Fig. 10C, D). Where competence contrast is missing, as it is the case in the western region, deformation is spread across a wide zone without major discrete high-strain structures.

The onset of shear zone activity could have occurred early during the main Variscan deformation phase (Fig. 10D), since the youngest rocks affected by the MSFZ probably belong to the Visean Soulcem granite suite. A continuous progression from high-strain to low-strain ductile and low-strain brittle is possible and supported by preserved mylonitic orthogneisses and mica schists at the base of the hanging-wall juxtaposed with faulted phyllites in the MFZ, all displaying the same shear sense. We postulate that the majority of displacement took place during the main-phase Variscan deformation. However, until supported by chronological data, this remains an educated guess. Reactivation of brittle fabrics during Alpine orogeny is likely, though we regard its effect as minor.

The MSFZ owes its existence to the presence of two adjacent large gneiss domes, which provided a strong competence contrast and forced partitioning of strain that resulted in the development of long lasting high-strain shear zones. Regional lateral fault and shear zones in other orogenic belts may have experienced a similar tectonic evolution.

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10. References

/5/ Alsop, G.I. & Holdsworth R.E. (2004): Shear zones – an introduction and overview. –In: Alsop, G.I., Holdsworth, R.E., McCaffrey, K.J.W. & Hand., M. (eds.): Flow Processes in Faults and Shear Zones. – Geological Society of London Special Publication, 224: 1-9.

/6/ Autran, A. & Garcia-Sansegundo, J. (1996): Carte structurale de tectonique hercynienne (moitié orientale). – In: Barnolas, A. & Chiron, J. C. (eds.): Synthèse géologique et géophysique des Pyrénées: Tec H1, H2; Orléans, Madrid (BRGM - ITGE).

/7/ Besson M. (1991): Notice explicative, Carte Géologique de France (1:50.000), feuille Fontargente (1093). – 35 pp. Orléans (Bureau de Recherches Géologiques et Minières).

/8/ Bureau de Recherches Géologiques et Minières & Instituto Géologico y Minero de España (2010): Carte Géologique des Pyrénées. – Scale 1:400.000.

/9/ Carreras, J. (2001): Zooming on Northern Cap de Creus shear zones. – Journal of Structural Geology, 23: 1457-1486.

/10/ Carreras, J. & Cirés, J. (1986): The geological significance of the western termination of the Mérens fault at Port Vell, (Central Pyrenees). – Tectonophysics, 129: 99-114.

/11/ Carreras, J. & Debat, P. (1995): Tectonique Hercynienne. – In: Barnolas, A. & Chiron, J.C. (eds.): Synthèse géologique et géophysique des Pyrénées: 585-677; Orléans, Madrid (BRGM –ITGE).

/12/ Casteras, M., Destombes J.-P. & Raguin, E. (1969): Notice explicative, Carte Géologique de la France (1:50.000), feuille Vicdessos (Xxi-48). – 16 pp. Orléans (Bureau de Recherches Géologiques et Minières).

/13/ Clariana, P. & Garcia-Sansegundo J. (2009): Variscan structure in the eastern part of the Pallaresa massif, Axial Zone of the Pyrenees (NW Andorra). Tectonic implications. – Bulletin Societé Géologique de France, 180: 501-511.

/14/ Denèle, Y., Olivier, P., Gleizes, G. & Barbey, P. (2007): The Hospitalet gneiss dome (Pyrenees) revisited: lateral flow during Variscan transpression in the middle crust. – Terra Nova, 19: 445-453.

/15/ Denèle, Y., Olivier, P. & Gleizes, G. (2008): Progressive deformation of a zone of magma transfer in a transpressional regime: The Variscan Mérens shear zone (Pyrenees, France). – Journal of Structural Geology, 30: 1138-1149.

/16/ Denèle, Y., Barbey, P., Deloule, E., Pelleter, E., Olivier, P. & Gleizes, G. (2009): Middle Ordovician U-Pb age of the Aston and Hospitalet orthogneissic laccoliths: their role in the Variscan evolution of the Pyrenees. – Bulletin Societé Géologique de France, 180: 209-216.

/17/ Denèle, Y., Olivier, P., Gleizes, G. & Barbey, P. (2009): Decoupling between the middle and upper crust during transpression-related lateral flow: Variscan evolution of the Aston gneiss dome (Pyrenees, France). – Tectonophysics, 477: 244-261.

/18/ Fitzgerald, P.G., Munoz, J.A., Coney, P.J. & Baldwin, S.L. (1999): Asymmetric exhumation across the Pyrenean orogen: implications for the tectonic evolution of a collisional orogen. – Earth and Planetary Science Letters, 173: 157-170.

/19/ Institut Géographique National (1992a): Carte topographique No. 2148 OT, Vicdessos. – Scale 1:25.000.

/20/ Institut Géographique National (1992b): Carte topographique No. 2249 OT, Bourg-Madame. – Scale 1:25.000.

/21/ Majoor, F.J.M. (1988): A geochronological study of the Axial Zone of the Central Pyrenees, with emphasis on Variscan events and Alpine resetting. – Verhandeling Nr. 6 Zwo Laboratorium voor Isotopen-Geologie: 1-117; Amsterdam.

/22/ Maurel, O. (2003): L'exhumation de la Zone Axiale des Pyrénées orientales: Une approche thermo-chronologique multi-méthodes du rôle des failles. Mémoire des Géosciences de Montpellier 31: 1-218.

/23/ Maurel, O., Monié, P., Pik, R., Arnaud, N., Brunel, M. & Jolivet, M. (2008): The Meso-Cenozoic thermo-tectonic evolution of the Eastern Pyrenees: an 40Ar/39Ar fission track and (U-Th)/He thermochronological study of the Canigou and Mont-Louis massifs. – International Journal of Earth Sciences, 97: 565-584.

/24/ McCaig, A.M . (1986): Thick- and thin-skinned tectonics in the Pyrenees. – Tectonophysics, 129: 319-342.

/25/ McCaig, A.M. (1987): Deformation and fluid-rock interaction in metasomatic dilatant shear bands. – Tectonophysics, 135: 121-132.

/26/ McCaig, A.M. & Miller, J.A. (1986): ^{40}Ar-^{39}Ar age of mylonites along the Merens fault, central Pyrenees. – Tectonophysics, 129: 149-172.

/27/ McCaig, A.M., Wickham, S.M. & Taylor, Jr., H.P. (1990): Deep fluid circulation in alpine shear zones, Pyrenees, France: field and oxygen isotope studies. – Contributions to Mineralogy and Petrology, 106: 41-60.

/28/ Mezger, J.E. (2005): Comparison of the western Aston-Hospitalet and the Bossòst domes: Evidence for polymetamorphism and its implications for the Variscan tectonic evolution of the Axial Zone of the Pyrenees. – Journal of the Virtual Explorer (electronic edition), 19: paper 6.

/29/ Mezger, J.E. (2009): Transpressional tectonic setting during the main Variscan deformation: evidence from four structural levels in the Bossòst and Aston-Hospitalet mantled gneiss domes, central Axial Zone, Pyrenees. – Bulletin Societé Géologique de France, 180: 199-207.

/30/ Mezger, J.E. (2010): Cadomian, Ordovician and Visean magmatism in the Axial Zone of the Central Pyrenees revealed by LA-ICP-MS U-Pb zircon dating. – 23. Réunion des Sciences de la Terre, Bordeaux, October 25-29, 2010, Livre des Résumés: 209.

/31/ Mezger, J.E. & Passchier, C.W. (2003): Polymetamorphism and ductile deformation of staurolite-cordierite schist of the Bossòst dome: indication for Variscan extension in the Axial Zone of the central Pyrenees. – Geological Magazine, 140: 595-612.

/32/ Monié, P., Soliva, J., Brunel, M. & Maluski, H. (1994): Les cisaillements myonitiques du granite de Millas (Pyrénées, France). Age Crétacé 40Ar/39Ar et interpretation tectonique. – Bulletin Societé Géologique de France, 165: 559-571.

/33/ Oele, J.A. (1966): The structural history of the Vall Ferrera area, the transition zone between the Aston Massif and the Salat-Pallaresa anticlinorium (central Pyrenees, France, Spain). – Leidse Geologische Mededelingen, 38: 129-164.

/34/ Passchier, C.W. & Trouw, R.A.J. (2005): Microtectonics. – 366 pp.; Berlin (Springer).

/35/ Ramsay, J.G. & Huber, M.I. (1987): The Techniques of Modern Structural Geology, Volume 2: Folds and Fractures. – 391 pp.; London (Academic Press).

/36/ Rölke, C. (2009): Strukturgeologische und petrologische Kartierung der Mérens-Scher-Zone im zentralen Aston-Hospitalet-Dom, Zentralpyrenäen (Frankreich, Andorra) - Ostgebiet. – Unpublished diploma mapping thesis, Martin-Luther-Universität. – 65 pp.; Halle (Saale).

/37/ Schnapperelle, S. (2010): Gefügekundliche, mikrotektonische Untersuchung der Mérens Scher- und Störungszone im Aston- und Hospitalet-Dom, Zentralpyrenäen (Frankreich, Andorra). – Unpublished diploma thesis, Martin-Luther-Universität. – 60 pp.; Halle (Saale).

/38/ Soula, J.C., Lamouroux, C., Villard, P., Bessière, G., Debat, P. & Ferret, B. (1986): The mylonite zones in the Pyrenees and their place in the Alpine tectonic evolution. – Tectonophysics, 129: 115-147.

/39/ Twiss, R.J. & Moores, E.M. (1992): Structural Geology: 465–497; New York (Freeman).

/40/ van den Eeckhout, B. (1986): A case study of a mantled gneiss antiform, the Hospitalet Massif, Pyrenees (Andorra, France). – Geologica Ultraiectina 45: 1-193.

/41/ Verges, J., Fernández, C. & Martinez, A. (2002): The Pyrenean orogen: pre-, syn-, and post-collisional evolution. – Journal of the Virtual Explorer, 8: 57-76.

/42/ Zwart, H.J. (1958): La Faille de Mérens dans les Pyrénées ariégeoises. – Bulletin Societé Géologique de France, 8: 793-796.

/43/ Zwart, H.J. (1965): Geological map of the Paleozoic of the central Pyrenees: sheet 6, Aston, France, Andorra, Spain. – Scale 1:50.000; Leiden (Geological Institute, University of Leiden).