SWITCHING DEFORMATION MODE AND MECHANISMS DURING SUBDUCTION OF CONTINENTAL CRUST: A CASE STUDY FROM ALPINE CORSICA

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ABSTRACT

The switching in deformation mode (from distributed to localized) and mechanisms (viscous versus frictional) represent a relevant issue in the frame of crustal deformation, being also connected with the concept of the brittle-“ductile” transition and seismogenesis. In subduction environment, switching in deformation mode and mechanisms may be inferred along the subduction interface, in a transition zone between the highly coupled (seismogenic zone) and decoupled deeper aseismic domain (stable slip). On the other hand, the role of brittle precursors in nucleating crystal-plastic shear zones has received more and more consideration being now recognized as fundamental in the localization of deformation and shear zone development, thus representing a case in which switching deformation mode and mechanisms interact and relate to each other. This contribution analyzes an example of a crystal plastic shear zone localized by brittle precursor formed within a host granitic-protomylonite during deformation in subduction-related environment. The studied structures, possibly formed by transient instability associated with fluctuations of pore fluid pressure and episodic strain rate variations may be considered as a small scale example of fault behaviour associated with a cycle of interseismic creep and coseismic rupture or a new analogue for episodic tremors and slow slip structures. Our case-study represents, therefore, a fossil example of association of fault structures related with stick-slip strain accomodation during subduction of continental crust.

Key words

Mixed mode of deformation, Brittle precursors, Shear zones, HP/LT mylonites, Subducted continental crust, slow slip events, pore fluid pressure

Commentato [GV1]: Not a proper abstract. Please rewrite by introducing the main issues that you have adressed and the RESULTS of your study. Then, draw a couple of conclusions highlighting the broad implications of your own work.

As it is now, it reads more like a general sum up of fundamental concepts underpinning our current understanding of cyclicity.

Commentato [GV2]: Deformation mode is, at least to me, a generic term that does not necessarily imply distributed and/or localised strain accommodation.

Highlights

We present a new case study on the role of brittle precursors in nucleating shear zone

We described an HP/LT microscale ultramylonite developed by brittle precursors induced during deformation within an host HP/LT granitic mylonite;

We related the studied structures with transient instability associated with fluctuations of pore fluid pressure and, possibly, with episodic strain rate variations

We considered our case study a small scale example of fault behaviour associated with stick-slip strain accomodation of faults during subduction of continental crust.

ABSTRACT

The switching in deformation mode (from distributed to localized) and mechanisms (viscous versus frictional) represent a relevant issue in the frame of crustal deformation, being also connected with the concept of the brittle-“ductile” transition and seismogenesis. In subduction environment, switching in deformation mode and mechanisms may be inferred along the subduction interface, in a transition zone between the highly coupled (seismogenic zone) and decoupled deeper aseismic domain (stable slip). On the other hand, the role of brittle precursors in nucleating crystal-plastic shear zones has received more and more consideration being now recognized as fundamental in the localization of deformation and shear zone development, thus representing a case in which switching deformation mode and mechanisms interact and relate to each other. This contribution analyzes an example of a crystal plastic shear zone localized by brittle precursor formed within a host granitic-protomylonite during deformation in subduction-related environment. The studied structures, possibly formed by transient instability associated with fluctuations of pore fluid pressure and episodic strain rate variations may be considered as a small scale example of fault behaviour associated with a cycle of interseismic creep and coseismic rupture or a new analogue for episodic tremors and slow slip structures. Our case-study represents, therefore, a fossil example of association of fault structures related with stick-slip strain accomodation during subduction of continental crust.

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INTRODUCTION

The study of the deformation fabric of fault rocks has been ~~the base to~~crucial for the development of a general conceptual model for crustal~~-~~scale fault zones (Sibson, 1977; Sibson, 1983; Scholz, 1988; Scholtz, 2002; Handy et al., 2007; Cooper et al. 2010; Platt and Behr, 2011). In this model, the increasing PT conditions determine the transition from a seismogenic frictional regime, dominated by pressure-sensitive deformation and involving cataclasis and frictional sliding, to a viscous regime (Rutter, 1986; Schimd and Handy, 1991; Handy and Brun, 2004), where dominantly aseismic, mainly

Commentato [GV3]: I guess this is what you meant? What other model otherwise?

crystal-plastic and continuous shearing is localized within mylonitic shear zones. In quartzo-feldspathic rocks, this transition is primarily determined by the temperature-related quartz response to changes of~~in~~ deviatoric stress, with dislocation creep becoming the principal deformation mechanism at T>270 °C (e.g. Stipp et al., 2002 and references).

Rock deformation experiments (Rutter, 1986; Shimamoto and Logan 1986; Bos and Spiers 2002; Scholz 2002) have shown that the shear strength of simulated faults at the brittle-viscous transition may depend on normal stress (as by faulting dominated by cataclastic mechanisms) although strain is achieved through crystal plasticity and/or solution transfer. These results support the hypothesis that some mylonitic shear zones are produced by coupled frictional and viscous mechanisms under semi-brittle conditions (Shimamoto and Logan, 1986; Shimamoto, 1989; Chester, 1989; Scholz, 2002; Pec et al., 2012 and references). ~~Studies~~ The experimental approach on semi-brittle behaviour, however, still leaves several ~~open~~ questions open ~~regarding~~ for natural fault zones, such as, for example, the ~~including~~ ~~micro~~mechanisms controlling bulk-rock deformation at the microscopic scale, the degree of interdependence of active deformation mechanisms, their cyclicity and the associated bulk rock style of deformation (Sibson, 1980; White and White, 1983; Takagi et al., 2000; Handy and Brun, 2004; Pec et al., 2012). A c~~C~~omplex transitional behaviour involving mixed continuous and discontinuous, distributed vs. localized, and cyclic switching in deformation mechanisms over large variations in strain rates is inferred at the transition between frictional and viscous domains, a ~~zone~~ depth interval in the crust which ~~includes~~ contains the typical hypocentres and rupture depths of~~for~~ large earthquakes in continental crust (Sibson, 1983; Kohlstedt et al., 1995; Scholz, 2002; Handy and Brun, 2004).

In ~~convergent~~ subduction settings~~, similarly~~, thise transition zone is~~,~~ located between 10 and 35 km depth, depending on slab dip and thermal structure (i.e., between temperatures of 150 °C and 350–450 °C), and, along the subduction interface,~~s~~ is recognized as the site of megathrust earthquake nucleation and concentrated postseismic afterslip, as well as the focus site of episodic tremor and slow slip events (Hacker et al., 2003; Vannucchi et al., 2008; Meneghini et al., 2010; Angiboust et al., 2012; Andersen et al., 2014; Hayman and Lavier, 2014; Fagereng et al., 2014; Angiboust et al., 2015).

The feedback between brittle and viscous deformation modes is relevant also for the mechanisms of shear zone nucleation, ~~so that~~and fracturing has been proposed to be even a pre-requisite for the initiation of ductile shear zones in the lithosphere (Handy and Stünitz, 2002; Pennacchioni and Mancktelow, 2007; Fusseis and Handy, 2008, among others). Shear zones and style of strain accommodation in terms of distributed vs. localized deformation in granitoids have been described by different authors (Ramsay and Graham, 1970; Burg and Laurent 1978; Simpson, 1983; Gapais et

Commentato [GV4]:

Commentato [GV5R4]: I do nt understand this “as by”. Could you rephrase, please?

Commentato [GV6]: A mechanism is not macro or micro, it is only observed at different scales.

Commentato [GV7]: Why convergent and subduction? Any subduction without convergenge?

al. 1987; Goncalves et al., 2016 and references therein). In this context, the role of brittle precursors in nucleating crystal-plastic shear zones has recently received ~~more and more~~significant consideration, ~~being~~ now being widely recognized as ~~having~~ playing a fundamental role in the localization process (Segall and Simpson, 1986; Mancktelow and Pennacchioni, 2005; Pennacchioni, 2005; Pennacchioni and Mancktelow, 2007; Menegon and Pennacchioni, 2010; Pennacchioni and Zucchi 2012, among others).

~~However t~~These studies, however, ~~works~~ mostly deal with brittle precursors consisting of inherited structures, such as discontinuities already existing at the beginning of the viscous deformation history, as in the ~~the~~ case~~s~~ of cooling joints, cataclasites and veins (e.g. Guermani and Pennacchioni, 1998; Pennacchioni and Mancktelow 2007); ~~and~~ most of these studies~~hem~~ consider deformation at shallow to intermediate crustal depths (Fusseis and Handy, 2008; Mazzoli et al., 2009; Molli et al., 2011), although case studies ~~framed~~ under conditions of upper amphibolite to granulite ~~facies~~ (White, 1996; Pennacchioni and Cesare, 1997; Kisters et al., 2000; Pittarello et al., 2013; Altenberger et al., 2013) and eclogite facies (Austrheim and Boundy, 1994; Austrheim, 2013 and references therein) were also investigated.

The case of deformation of crustal units and granitoids during subduction has only recently been ~~only recently~~ analysed in terms of preserved rock records of the paleo-seismic cycle and/or slow-slip phenomena (Angiboust et al., 2015). A well-documented study from the Dent Blanche Thrust in the Western Alps, for example, is framed in the hanging wall of an ancient subduction interface zone (Angiboust et al., 2014; 2015).

~~On the contrary, o~~Our contribution integrates the existing literatue by analysing~~es~~ a microscale example of brittle precursor of a crystal plastic shear zone derived from the footwall of a subduction interface in ~~ambient~~ blueschist-facies conditions, thus ~~representing~~ investigating an hitherto ~~still not yet~~ undocumented or less ~~documented~~ thoroughly studied case~~case study~~ (Molli et al., 2005; Molli 2007).

Commentato [GV8]: Give a short description of what your study encompassed. What techniques were used, on what in detail and to do what?

By virtue of their deep origin, the analysed structures are ideal~~ly suited~~ to contribute to the ongoing discussion on the deformation style and mechanisms associated with the broad spectrum of fault-slip behaviour (from seismic slip ~~to~~ through stable aseismic creep, to episodic slow slip events and non-volcanic tremors) recorded by seismic and geodetic observations at active plate boundaries (e.g. Peng and Gomberg 2010; Beroza and Ide, 2011, and references therein). Slow slip events (i.e. fault slip events with slip rates in between coseismic rates~~slip~~ and aseismic creep, and generating equivalent seismic moments similar to large earthquakes) in subduction zones have been recorded in a depth interval that experiences temperatures between 250 and 650° C and pressures between 0.6 and 1.2 GPa (Beroza and Ide, 2011), typically in areas of high Vp/Vs ratios suggestive of local high fluid pressures. This locates slow slip events at the lower end of the seismogenic zone, at metamorphic conditions where the rheology is expected to be viscous or at the frictional-viscous transition. Accordingly, recent studies of shear zones exhumed from similar conditions along the subduction interface have suggested that coupled fracture and viscous flow, possibly associated with fluctuations of~~in~~ fluid pressure, can originate tremors and slow slips (e.g. Fagereng et al., 2014; Hayman and Lavier, 2014; Angiboust et al., 2015; Malatesta et al., 2017). Here we show that the HP brittle-viscous transition preserved in the Popolasca granitoids of northern Corsica can also be explained by ~~with~~ transient high fluid pressures triggering brittle deformation in an otherwise viscous regime, and discuss the related implications for fault-slip behaviours in subduction zones.

REGIONAL BACKGROUND AND GEOLOGICAL SETTING OF STUDIED SAMPLE

In Alpine-type orogens, the study of the meso- and microstructural record of exhumed subduction-related thrust zones and its interpretation in terms of subduction zone rheology and seismicity have received increasing attention since the end of the ~~'~~90's (Stöckhert et al.,1999; Kuster and Stöckhert, 1999). This subject has been developed and explored in particular in ~~the~~ Alpine Corsica (Austrheim and Andersen, 2004; Andersen and Austrheim, 2006; Healy et al., 2009; Andersen et al., 2014; Deseta et al., 2014a,b; Magott et al., 2016a,b) in oceanic units made up of peridotite, serpentinite, gabbro, basalt, calcareous and siliceous schist and marble exposed as remnants ~~of the lithosphere~~ of the Jurassic Piemonte- Liguria oceanic basin lithosphere and its pelagic sedimentary cover (Mattauer et al., 1981; Bezert and Caby, 1988; Jolivet et al., 1990; Molli, 2008; Vitale Brovarone et al., 2013).
The~~se~~ widespread occurrence~~s~~ and preservation~~s~~ of relicts of exhumed seismogenic ~~exhumed~~ structures are mainly ~~related~~ due to the peculiar geologic history of Alpine Corsica, which is connected with the development of the Alps/Apennine orogenic system (Molli and Malavieille, 2011; Guyedan et al., 2017; Beaudoin et al., in press). The latter did not develop ~~with the lacks of the~~ continent-continent "hard collision"- related structures and lacks the ~~th~~thermal reworking observable instead in the Alps (e.g. Polino et al., 1995; Schmid et al., 1997; Berger and Busquet, 2008; Butler, 2013; Rosenberg and Kissling, 2013; Carminati and Doglioni, 2014), thus resulting ~~and a~~in the better preservation of the early stages of subduction-related structures and fabrics.
Corsica, therefore, represents a suitable natural laboratory for the investigation of ~~the~~ subduction-related processes in oceanic and continental crust, as firstly suggested by Mattauer et al. (1981) and Gibson and Horak (1984).
In Corsica, continent~~al~~-derived units, i.e. units derived from the footwall of the ancient subduction interface zone, can be observed in three different structural positions (Fig.1), each ~~of them~~

Commentato [GV9]: Explain what you mean by Alpine Corsica for the not expert.

corresponding to ~~a~~ different peak metamorphism conditions (Tribuzio and Giacomini, 2004; Molli, 2008; Vitale Brovarone et al., 201~~3):~~

- the innermost slices (Serra di Pigno/Farinole units) interleaved with oceanic units show eclogite peak conditions at 1,5/1,8 GPa; 500 ± 50 °C;
- the intermediate ones (e.g. the Tenda and Centuri units) are instead characterized by Upper Blueschist facies peak conditions at 0,9/1,1 GPa; 450± 50 °C;
- the most external (e.g. Corte/Popolasca units) ~~show~~ contain high pressure greenschist and/or blueschist facies peak assemblages developed at T =325/370°C, P = 0.75/0.85 GPa (Malasoma et al., 2006; Di Rosa et al., 2016).

These occurrences document the progressive underthrusting of these units down to~~at~~ different depths and their in-sequence contractional exhumation within an A~~a~~lpine-age~~,~~ east-dipping subduction system (Mattauer et al., 1981; Bezert and Caby, 1988; Jolivet et al., 1990; Garfagnoli et al., 2009; Molli and Malavieille, 2011; Maggi et al., 2012; Di Rosa et al., 2016; Guyedan et al., 2017).

The studied samples come from the external continental units (Molli, 2008; Molli and Malavieille, 2011; Vitale Brovarone et al., 2013) of Alpine Corsica and, in more detail, from the Popolasca Unit (Bezert and Caby, 1988; Malasoma et al., 2006; Di Rosa et al., 2016). This unit is characterized by a pre-Mesozoic basement mainly formed by granitoids, a Permo-Mesozoic metasedimentary sequence and an Early-Eocene flysch (Fig. 1).

Blueschist assemblages in the unit have been described by regional studies (Bezert and Caby, 1988; Malasoma et al., 2006; Molli, 2008; Di Rosa et al., 2016) and can be observed in all suitable rock-types, such as~~: in~~ metapelites from cover rocks or~~, in~~ metabasic dykes and some of the granitoid suites from the basement. Peak metamorphism has been initially constrained ~~at~~ to 250/350°C, 0.4/0.55 GPa (Bezert and Caby, 1988) and~~or~~ more recently ~~better defined~~refined to ~~at~~ 325/370 °C and 0.75/0.85 GPa (Malasoma et al., 2006; Di Rosa et al., 2016). The prograde to peak pressure assemblage has been recently dated ~~in the range of~~to 45-36 Ma by Di Vincenzo et al., (2016).

Commentato [GV10]: Ar Ar? Specify, please.

The contributions of Malasoma et al., (2006) and~~,~~ Di Rosa et al., (2016) analysed the cartographic to mesoscopic-scale structural geometries of deformation in the area of the studied sample.

Superimposed foliations and fold structures are typical of the metasedimentary cover, with blueschist assemblages as relict fabrics. On the contrary, in basement rocks a main phase continuous foliation can be observed wrapping around undeformed granitoids. The main foliation is ~~either~~ associated either with blueschist or greenschist facies assemblages (Malasoma et al., 2005; Di Rosa et al., 2016; Di Vincenzo et al., 2016). Anastomosed networks of fault zones associated with subgreenschist facies assemblages overprint~~s~~ all previous structures and may be related to the activity of a major N/S trending transcurrent fault zone (Central Corsica Fault Zone) reworked by ~~normal kinematics~~an extensional phase ~~active~~ during the Oligocene-Miocene in the frame of the rototranslation of the Corsica-Sardinia microblock and upper-plate extension associated with Apenninic subduction (Faccenna et al., 2004; Molli, 2008; Molli and Malavieille, 2011; Carminati and Doglioni, 2014; Guyedan et al, 2017).

GEOMETRY OF DEFORMATION AND MICROSTRUCTURES

The analysed sample (Fig.2) is a deformed granitoid of calcalcaline affinity~~suite~~ (K-feldspar, plagioclase, quartz, biotite) well known in the Hercynian Corsica (Rossi et al., 2001). It ~~shows~~ contains a continuous foliation (Fig.2,3 a,b) mainly defined by the shape preferred orientation of quartz, feldspars and biotite grains.

Shape anisotropy of quartz in the host protomylonite shows an aspect ratio around 0,37 (mean X/Z of 2,7), which indicates a shear strain γ of c.1 assuming simple shear deformation ~~in simple shear~~. Quartz ~~shows~~ displays typical features of low temperature plasticity (Tullis et al., 2000; Stipp et al., 2002; Vernooij et al., 2006; Trepmann et al., 2007; Derez et al., 2015; Kjøll et al., 2015) (Fig 3), such as undulatory extinction, localized extinction bands (LEBs, following the terminology of Derez et al., 2015), typically forming conjugate sets, and up to 100 µm thick intracrystalline bands of recrystallized grains (5-10 µm in size). Bands of recrystallized grains occur parallel to the main foliation as well as in conjugate sets intersecting at ca. 90° and parallel to the conjugate sets of LEBs.

Feldspars show local evidence of grain size reduction by microcraking and microfaulting (Fig.3) associated with K-feldspar breakdown to albite (by bulging recrystallization; ~~) (~~Fig.3), stilpnomelane and phengite. Thin needles of Na-amphibole (Fig.4) attest to the development of this fabric in HP/LT conditions as described below.

The main foliation is cross-cut at high angle by a millimetre-thick localized zone of deformation (Fig. 2, 5). This shows a sharp boundary, truncating flattened quartz and feldspar grains, which suggests~~;~~ ~~a feature suggesting~~ its development as an unstable fracture (Schmid and Handy, 1991; Passchier and Trouw, 2011).

Three compositionally controlled domains can be recognized within the thin localized shear zone (Fig. 2, 5):

Domain 1 is feldspar-dominated and shows microstructural features typical of a cataclasite (Fig.5a,c). Feldspar clasts (and quartz grains) reveal both displacive intragranular fractures and intergranular "stable" cracks (Atkinson, 1982; Schmid and Handy, 1991).

Commentato [GV11]: Nice microstructures, which, I believe, deserve a better documentation/description. See detailed comments below.

Commentato [GV12]: Why at 90? Are they really conjugate and we are in Tresca's conditions or, maybe, they are not tru conjugates?
Btw, not clear to me where these are on the microphotograph. Please highlight them.

Commentato [GV13]: Please show on Fig. 5 and explain better, not really clear.

Commentato [GV14]: How do these quartz grains look like? It is fundamental to document them. Are they deformed qtz? Recrystallised? They should be like qtz of the protomylonite. Can you exclude that they look like qtz of the proper mylonite, thus indicating their late cataclastic age?

Commentato [GV15]: What does this mean?

Commentato [GV16]:

Commentato [GV17R16]: I think the authors need to both document this better but also explain this better. All I can see in the provided figure is a thin cataclastic band, with a brown-red ground mass and white, poorly sorted and variably angular clasts. The resolution of the image is insufficient to follow the description.

Domain 2 is “phyllosilicate”-dominated (stilpnomelane and phengite) and shows microstructural features typical of a foliated-cataclasite/phyllonite (Fig.5a,b). Asymmetric porphyroclasts and shear band systems characterize the phyllosilicate-rich parts of the shear zone.

Domain 3 is a quartz-albite rich domain showing microstructural features typical of an ultramylonite. This is characterized by very fine recrystallized albite and quartz grains (5-10 µm in size) (Fig.5a,b; Fig.6) with strong crystallographic preferred orientation (see below). A~~Q~~ quartz porphyroclast shows a mean aspect ratio around 0,09 (mean X/Z of 11) which corresponds to a calculated shear strain γ of c. 3 At this shear strain the ~~corresponding~~ angle between the foliation and the shear zone boundary in simple shear is around 15 °, which agrees with the mean orientation of the quartz porphyroclast~~s~~ shape preferred orientation defining the ultramylonitic foliation (Fig.6a,b,c). Syn to post-kinematic Na-amphibole (Fig.6a) documents shearing in HP/LT metamorphic conditions, as illustrated here below.

MINERAL CHEMISTRY AND ESTIMATE OF METAMORPHIC PRESSURE-TEMPERATURE (P-T) CONDITIONS

Chemical analyses of coexisting minerals within the metamorphic assemblage (Tab.1) were obtained using a JEOL JXA-8600 electron microprobe, equipped with four wavelength-dispersive spectrometers, at the CNR - Istituto di Geoscienze e Georisorse, Firenze, Italy. Running conditions were 15 kV accelerating voltage and 10 nA beam current on a Faraday cage. Counting time for the determined elements ranged from 10 to 60 s at both peak and background. The Bence and Albee (1968) method was employed for the correction of all data. A number of synthetic and mineral standards were used for instrumental calibration.

Amphibole. Structural formulae were calculated assuming 23 oxygens, and the classification of Leake et al. (1997) was adopted. Site assignment and ferric iron contents were calculated using the scheme proposed by Schumacher in Leake et al. (1997). Because of the small sizes of the crystals (widths *c.* 10 µm), it was not possible to make compositional traverses across individual crystals to detect intracrystalline variations in chemical composition, such as core-to-rim zonation. Thus, each analysis reported in Table 1 represents a different crystal. In the studied sample the Na-amphiboles are mostly riebeckite (Tab.1 and Fig. 4b) with low $Mg/(Mg+Fe^{2+})$ ratio (0.13-0.21) and are characterized by Si contents close to the maximum of 8.0 apfu.

Other minerals. All analysed albites, of which structural formulae were calculated assuming 8 oxygens, have composition close to the pure end-member. Stilpnomelane structural formulae were calculated assuming 24 oxygens and all Fe as divalent (Fe^{2+}). Stilpnomelanes have Fe amounts

Commentato [GV18]: Label 5°. A is not shown on the figure.

Commentato [GV19]: Again, a better photographic documentation would help.

Commentato [GV20]: Please mark more clearly om Fig. 5°. I see not dashed square for domain 3 and can only guess that it is the area to the left of square of domain 2.

Commentato [GV21]: Please show on 5 where these photos are more or less from. Not obvious.

Commentato [GV22]: Unclear whether what you call here pclast is actually the ribbon of the caption to Figure 6? Please be consistent, hard to follow otherwise.

Commentato [GV23]: Why in the microphotograph do you show elongated objects with ratios up to 15? Why don't you stick to the same value and, therefore, object? Use the one giving you the maximum strain indication.

Commentato [GV24]: Why do you refer to a, b and c? Please, outline the sz boundaries, especially in a and c.

Commentato [GV25]: Are those parallel to the short ribbon objects? I would expect so, given that you use that figure to illustrate the angular relationships between the sz boundaries and the mylonitic foliation.

ranging from 1.86 to 2.91 apfu and Mg amounts ranging from 0.24 to 0.58 apfu. For K-feldspar and biotite porphyroclasts, structural formulae were calculated assuming respectively 8 and 22 oxygens. The analysed K-feldspars have composition close to the orthoclase pure end-member, with minor amounts of Na.

The peak metamorphic mineral assemblage is characterised by Na-amphibole + phengite + quartz + albite + stilpnomelane. Na-amphibole is a typical mineral ~~related to~~of the blueschist facies and is indicative of HP/LT gradient metamorphism (e.g. Evans 1990; Schiffman & Day 1999 and references). The minimum temperature (T) and pressure (P) conditions of this mineral assemblage can be estimated using the reaction curves proposed by Schiffman & Day (1999) for the appearance of Na-amphibole (stability field of the blueschist facies); ~~meanwhile the~~ maximum temperature (T) conditions were instead constrained by the presence of stilpnomelane, as shown by the equilibrium: Stp+Phe=Bt+Chl+Qtz+W (Massonne & Szpurka, 1997). Summing up all the available thermobarometric information, the metamorphic conditions can be estimated as temperature (T) around ~~-~~320±50 °C and pressure (P) > of 0.70 GPa (Fig.4c),~~;~~ consistent with those described by Malasoma *et al.* (2006) in metagranitic rocks from the area of our studied sample.

EBSD DATA

Commentato [GV26]: As mentioned in a comment in the intro, you should really introduce this component of the study earlier on. Why did you do EBSD? On what and aiming at what? Please introduce somewhere in the text, up to you where.

EBSD analysis of quartz was conducted with a Jeol 7001 FEG-SEM equipped with a NordlysMax EBSD detector (Oxford Instruments) at the Plymouth University Electron Microscopy Centre. Working conditions during acquisition of the EBSD patterns were 20 kV, 70° sample tilt and high vacuum. Thin sections were chemically polished with colloidal silica and carbon-coated before the analysis. EBSD patterns were acquired on rectangular grids with step size of 1 μm and 2 μm. EBSD patterns were acquired and indexed with the AZtec software and processed with Channel 5 software (Oxford Instruments). Raw EBSD data were processed to reduce data noise following the procedure tested by Prior et al. (2002) and Bestmann and Prior (2003). EBSD results are shown in form of Inverse Pole Figure map, pole figures (equal angle, lower hemisphere) of crystallographic axes and planes (<0001> c-axis, <11-20> a-axis, {10-10} prism {m}, {10-11} positive rhomb {r}, {01-11} negative rhomb {z}), misorientation pofiles and plots of misorientation axis in sample coordinates.

Quartz in the host rock

We analysed a monocrystalline quartz ribbon elongated parallel to the host-rock foliation and sharply cut by the thin localized shear zone. The quartz ribbon contains two nearly orthogonal sets of intracrystalline shear bands of recrystallized grains (Fig. 7a); one set is oriented at high angle (90° ± 20, *set 1*) and one at low angle (≤ 20°, *set 2*) to the host rock foliation (vertical in Fig. 7a). The average grain size of the recrystallized grains is 5±2 μm. The host ribbon grain contains also fine localized extinction bands (up to 20 μm thick) (LEBs: Derez et al., 2015) subparallel to the bands of recrystallized grains. Low-angle boundaries are ubiquitous in the ribbon; on EBSD maps they are typically straight, poorly connected, and subparallel to the bands of recrystallized grains (Fig. 7b). Some low-angle boundaries are connected to form subgrains of approximately the same size of the recrystallized grains. Subgrains occur with a higher frequency at the intersection between two sets of recrystallized bands, and in regions sandwiched between closely spaced (≤ 100 μm) subparallel bands of recrystallized grains (Fig. 7b).

The c-axis of the ribbon is oriented near the pole to the host-rock foliation, i.e. in a position suitably oriented for the activation of the basal <a> slip system of quartz (Fig. 7c). The c-axis orientation of the recrystallized grains in the intracrystalline bands is mostly spread out along the periphery of the pole figure, although some scattered grains have their c-axis in intermediate positions between the X and Y directions of the pole figures (Fig. 7d). Such an orientation suggests that the recrystallized grains have experienced a rotation around the Y-direction of finite strain (i.e. centre of the pole figure) from the host-grain orientation (e.g. Van Daalen et al., 1999; Menegon et al., 2011).

This is confirmed by the boundary trace analysis (Prior et al., 2002; Menegon et al., 2010) of the two main sets of straight low-angle boundaries defining localized extinction bands (parallel to bands of recrystallized grains), one running ENE-WSW (*subset1*) and one running ca. N-S (*subset 2*) in Fig. 7b. The dispersion paths of crystallographic directions on the pole figures of subset 1 (Fig. 7e) identifies {m} as the rotation axis, which lies very close to the centre of the pole figure. The pole to the prismatic plane {m} is the rotation axis associated with the basal <a> and with the {a} <c> slip system in quartz (e.g. Neumann, 2000), and, accordingly, subset 1 can be interpreted as a tilt boundary plane produced by the activity of the slip system basal <a> (Fig. 7e) and containing the boundary trace of subset 1 and the rotation axis.

Subset 2 has a similar dispersion path as subset 1, with {m} as the identified rotation axis. However, in this case the boundary trace analysis is not consistent with a tilt boundary produced by the activity of the slip system basal <a>, but could indicate the activity of the {a} <c> slip system (Fig. 7f). However, activity of c-slip in quartz typically requires temperature in excess of 600° C (Kruhl, 1996; Zibra et al., 2010) and, therefore, appears unlikely in our samples. Moreover, misorientation profiles across low angle boundaries with a subset 2 orientation show abrupt misorientation jumps of up to 6º

Commentato [GV27]: Justify why you classfy these as shear bands. This is important to your model.

Commentato [GV28]: Indicate on the figure, please.

Commentato [GV29]: What ribbon? The host crystal?

Commentato [GV30]: If I may suggest it, fig. 7° and b deserve to be bigger. Some of the described features are nearly invisible at the current figure size. I suggest to split figure 7 into 2 parts.

(profiles 2 and 3 in Figs. 7b and 7g), as opposed to a gradual accumulation of misorientation towards low-angle boundaries with a subset 1 orientation (profile 1 in Figs. 7b and 7g). Thus, the low-angle boundaries with a subset 2 orientation could represent microcracks subparallel to the basal planes (e.g. Kjøll et al., 2015) that localized rigid body rotation of fragments around the Y-direction (e.g. Trepmann et al., 2007; Menegon et al., 2013). The rotation around Y of (1) the crystallographic directions of the host grain and (2) the recrystallized grains (in this case for large misorientation > 10°) is confirmed by the plots of the misorientation axis in sample coordinates (Fig. 2).

Quartz in the localized shear zone

We analysed polycrystalline ribbons of recrystallized grains from domain 3 of the localized shear zone (Fig. 8a). The c-axis orientation of the recrystallized grains defines an inclined type-I crossed girdle synthetically oriented with respect to the bulk shear sense of the shear zone (Fig. 8b). The c-axes are preferentially clustered near the pole to the shear zone boundary, i.e. in an orientation suitably oriented for the activity of the basal <a> slip system. The average grain size of the recrystallized grains in domain 3 is 6±2 μm.

ULTRAMYLONITE: PALEOPIEZOMETRY, FLOW STRESS AND STRAIN RATE

The microstructure and the crystallographic preferred orientation of quartz indicate that quartz in the protomylonite and ultramylonite deformed by dislocation creep (concomitant with microcracking in the protomylonite) and recrystallized to a fine-grained aggregate. Thus, the rheology and the flow stress in the ultramylonite can be evaluated extrapolating experimentally calibrated flow laws of quartz to the deformation conditions.

Commentato [GV31]: Ok, why concomitant? What do you consider as convincing evidence that the two processes are coeval? Not mentioned so far, nor demonstrated. Please, elaborate better.

The rheology of quartz deforming by dislocation creep is generally described in terms of a power-law equation:

$$\dot{e}= A f^{m}_{H_2O} e^{(-Q/RT)} s^{n} \qquad (1)$$

Where $\dot{\epsilon}$ is the strain rate, f_{H_2O} is the water fugacity (raised to the power of m), Q is the activation energy, R is the universal gas constant, T is the temperature, σ is the differential stress, and *n* is the stress exponent. We used the theoretical dislocation creep flow law of Hirth et al. (2001), which has derived a linear dependence of strain rate on the water fugacity (m=1) and a stress exponent of 4. A water fugacity of 172 MPa is calculated from the water fugacity coefficient reported in Tödheide

(1972) at T=350° C, P=0.8 GPa.

The paleostress (assuming steady-state flow at the time of viscous deformation) can be determined by recrystallized quartz grain size paleopiezometer (e.g. Stipp & Tullis, 2003), which have been calibrated in the form: $\Delta\sigma = B\ D^{-x}$,where $\Delta\sigma$ is the steady state differential stress ($\sigma1$ - $\sigma3$), D is the recrystallized grain size and B and X are empirical constants. Using the recrystallized grain size piezometer of quartz calibrated by Stipp and Tullis (2003), a recrystallized grain size of 5-10 μm indicates differential stress in the range of 110-190 MPa. Extrapolation of the flow law of Hirth et al. (2001) yields a strain rate in the range of $7.6 * 10^{-13} – 6.2 * 10^{-12}\ sec^{-1}$ in the ultramylonite at the estimated deformation temperature of 350° C.

DISCUSSION AND CONCLUSION

Viscous-brittle-viscous deformation under HP/LT conditions

The analysed example documents a switch in deformation mode and mechanisms within an HP/LT fault zone. On the base of overprinting relationships and microstructural features the following deformation sequence may be suggested: stage (1), associated with a distributed deformation and development of a protomylonitic foliation in the granitoid by quartz low temperature plasticity, microfracturing, and albite neo-crystallization from K-feldspar porphyroclasts (Fitzgerald and Stunitz, 1993), ~~has been~~was followed by stage (2), in which localized deformation by brittle fracturing formed a millimeter-thick cataclasite, which acted as a precursor for (3) localization of viscous deformation and ultramylonite development. The synkinematic and postkinematic growth of Na-amphibole in the host rock foliation and in localized ultramylonite indicates that the entire deformation sequence occurred at HP/LT conditions (ca. 350°C at ≥ 0.7GPa), corresponding to a depth of 23-30 km in the subduction channel.

The estimated P, T conditions are consistent with low-T plasticity regime in quartz (Stipp et al., 2002; Derez et al., 2015 and refs therein). Accordingly, the deformation microstructures of quartz produced during stage (1) in the host rock appear to be the product of the competition between dislocation activity and fracturing. Localized extinction bands at high angle to the host rock foliation (subset 1 in Fig. 7b) are consistent with the activity of the basal <a> slip system, whereas localized extinction bands subparallel to the host-rock foliation (subset 2 in Fig. 7b) are interpreted as fractures subparallel to the basal plane, as previously observed by Kjøll et al. (2015). Some localized extinction bands (especially those with a subset 2 orientation, see Fig. 7b) contain isolated small new grains that are considerably smaller than the average grain size in the recrystallized bands, and that are only slightly

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Commentato [GV32]: Very much possible, but what exactly is the evidence that low-T plasticity is coeval with microfracturing? Not necessarily obvious.

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Formattato: Evidenziato

Commentato [GV33]: Clasts therein are only of KFs? No deformed qtz aggregates? Can you exclude that cataclasis locally also postdates mylonitisation?

Formattato: Evidenziato

Commentato [GV34]: This is crucial to the story! I need to be convinced that this IS the case. Can you please list and better explain the observations and evidence that support this reconstruction?
Obviously the cataclasite postdates the protomylonitic, discordant foliation. But what is the time/geometric/dynamic relationship among all the microfracturing episodes you propose?

I miss something in your explanation.

Formattato: Evidenziato

Formattato: Evidenziato

misoriented with respect to the host grain. Moreover, the high-angle boundaries in such localized extinction bands are not always fully connected to define entire new grains. Together with the abrupt misorientation jumps (Fig. 7g), these observations further suggest that localized extinction bands with a subset 2 orientation represent fractured domains in which the fragments have rotated passively and sealed together, as proposed by e.g. Derez et al. (2015).

Despite the local fracturing of the host grain, deformation of the recrystallized grains in the conjugate intracrystalline bands involved dislocation activity, as indicated by the cluster of misorientation axis around the prism {m} for low misorientations (compare Fig. 7h with 7d). This is consistent with the local activity of the basal <a> slip system in the recrystallized bands, as also indicated by the CPO of the recrystallized grains in the bands at high angle to the host rock foliation (Fig. 7d).

Commentato [GV35]: Really?

The type-I crossed girdle c-axis CPO of quartz in the ultramylonite suggests the concomitant activity of basal <a>, rhomb <a> and prims <a> slip systems in the recrystallized ribbons (Fig. 8). The grain size of recrystallized quartz in the ultramylonites is in the same range (5-10 μm) as in the intracrystalline bands in the host rock. This suggests that 5-10 μm represents the equilibrium grain size with the flow stress (estimated in the range of 110-190 MPa with recrystallized grain size piezometry) during viscous deformation before and after the transient brittle event of stage (2).

Commentato [GV36]: True, but how representative of the bulk flow stress is the grain size of the recrystallised grains (or better, nucleated) within the intracrystalline bands? What is the role of the initial aperture of the precursor microcracks? Would that not be a limiting factor to further gain growth at least until other solution/precipitation mechanisms potentially could take over?
This aspect should be at least discussed.

During stage (2) the development of cataclasite and related dilatancy resulted in an increase of permeability and thus facilitated fluid access and fluid mobility in the shear zone. This enhanced mineral transformations as testified to by modal enrichment of stilpnomelane (by biotite breakdown) and Na-amphibole in the ultramylonite.

Therefore, the observed structures witness a change in deformation style (from distributed to localized strain in brittle precursor), and a switch in the dominant deformation mechanism (from low T plasticity in the host rock to cataclasis and back to crystal plasticity in the ultramylonite), which occurred at the footwall of the subduction interface at temperature conditions typical of the brittle/viscous transition in quartz/feldspathic rocks.

Finally, the transition from stage (2) and stage (3) is consistent with the general observation that nucleation of localized ductile shear zones requires the presence of a planar compositional or structural precursor (e.g. Pennacchioni and Zucchi, 2012, and references therein). The example from the Popolasca granite demonstrates that (i) nucleation on brittle precursors also occurs under HP/LT conditions in the subduction channel, and (ii) the brittle precursors are not necessarily inherited from an earlier deformation event, but can be the manifestation of switches in deformation mode in the footwall of the subduction interface.

Commentato [GV37]: See my comments above.

Significance of the switch in deformation mode and implications for fault-slip behaviours in subduction zones

We interpreted the inferred deformation sequence and structures as the result of transient instabilities (Sibson, 1980; White, 1996; Handy and Brun, 2004) possibly representative of mixed fault-slip behaviours at seismogenic depth in the subduction channel. We have estimated the conditions resulting in the transient brittle event during stage (2) under the following assumptions and approximations:

1. The coefficient of internal friction, μ_i, is generally between 0.5 and 1.0 in intact rocks (Sibson, 1985). We considered $\mu_i = 0.6$ in the failure envelope for the intact Popolasca granitoid.
2. We used a cohesive strength of 35 MPa as representative for granitoids (Amitrano and Schmittbuhl, 2002), with a resulting tensile strength of 17.5 MPa.
3. We assumed an~~d~~ Andersonian stress field and a thrusting regime. The effective vertical stress σ'_v corresponds to the effective minimum principal stress $\sigma'_3 = \sigma_3 - P_f$, where P_f is the pore fluid pressure. We considered a hydrostatic pore pressure ($\lambda = 0.4$, $P_f = 320$ MPa for $\sigma_3 = 800$ MPa) during the formation of the gneissic foliation in the host rock.
4. We assumed a differential stress of 110-190 MPa during viscous flow of the granitoid prior to brittle failure at $\sigma'_v = 480$ MPa, as derived from the recrystallized grain size piezometry in the protomylonite.

The results of our analysis are shown in Fig. 9a. For brittle failure, it is necessary to invoke high differential stresses (of the order of 1.1 GPa assuming a hydrostatic pore fluid pressure), or higher fluid pressure (Fig. 9a). Differential stresses in excess of 1 GPa have been associated with intermediate depth (50-300 km) earthquakes in the subduction channel (e.g. John et al., 2009), and are expected to result in extensive development of pseudotachylytes in granitoid rocks (e.g. Austrheim 2013 and references therein), which, on the contrary, have not been observed in the sample studied here.

For the differential stress range and the vertical stress considered here, brittle failure requires (sub)lithostathic fluid pressure ($0.96 < \lambda < 1$, Fig. 9). The strength envelope plotted for pore fluid pressure between 0.98 and 0.94 and for a strain rate of 10^{-12} s^{-1} (Fig. 9b) suggests that viscous deformation at the estimated depth range (23-30 km) is only possible for pore fluid pressure ≤ 0.94, otherwise brittle deformation is expected to occur. Thus, under the assumptions listed above, our analysis indicates that local fluctuations in pore fluid pressure can explain the cyclic viscous-brittle-viscous deformation switch. However, despite the synkinematic growth of hydrous minerals in the cataclasite and the high pore fluid pressure required at failure, there is no evidence of macroscopic veining or hybrid fractures in the samples. Our analysis is consistent with this observation, in that failure occurs entirely in the shear fractures field and not in the (hybrid) shear extension fractures field (Fig. 9a).

Drawing from the results shown in Figs. 9a,b and using concepts and inferences coming from modern studies of convergent subduction system (e.g. Ozowa et al., 2002; Fu and Freymuller, 2013; Bedford et al., 2013), we may suggest the following deformation path (Fig.10) for the structural features of our analysed sample. A first stage of distributed deformation (stage 1) may be associated with large scale anastomosed shear zone development and aseismic creep; subsequently~~, then,~~ following Angiboust et al. (2015), two possible deformation scenario and slip patterns may be envisaged: (a) propagation at seismic rate of microrupture followed by afterslip or (b) slow slip phenomena in an aseismically creeping crust. In the first case, a transient strain rate increase associated with brittle fracturing would represent the deep response of the transition zone to a specific stage of the seismic cycle taking place higher up along the seismogenic portion of the subduction interface (Fig.10c, scenario 1) or in less deformed domains acting as local stress raisers nearby (Fig.10c, scenario 2). Cataclastic deformation would then correlate with coseismic to postseismic deformation higher up along the interface. In that case, the mylonitization of the brittle precursor would be the record of subsequent interseismic deformation at lower strain rates, taking place in the time frame between two earthquakes.

In the second case (Fig.10d), the transient highs in strain rates ~~related~~ expressed by~~to~~ brittle fracturing and cataclasis could be the field evidence of deformation associated with slow slip events (SSEs) or other transient slips, generally localized along the subduction interface at this depth range itself (e.g., Shelly et al., 2006; Fagereng et al., 2014). We are aware that the mechanism(s) of SSEs initiation are still poorly understood, and that stress/strain rate perturbations triggered by an earthquake nearby followed by postseismic slip and interseismic creep is an equally feasible mechanism to explain the deformation sequence recorded in our sample. However, our preferred interpretation is that the transient brittle deformation recorded in our studied sample is the manifestation of a slow slip event, for the following reasons: (1) subduction interface SSEs typically occur at the downdip transition from stick-slip behaviour to aseismic creep (e.g. Wallace and Beavan, 2010), and, in granitoid rocks, this transition is expected to occur at the T range of deformation of our case-study, as witnessed by the deformation microstructures of quartz and feldspar in the host protomylonite; (2) SSEs are often related to high pore fluid pressure (e.g. Liu and Rice, 2007).

Whatever the actual process triggering brittle deformation was, we want to emphasize that detailed microstructural studies of exhumed shear zones are a valuable complement to the geodetic,

Commentato [GV38]: Where is this described? If I am not mistaken, you only described K-Fs and qtz…

Commentato [GV39]: Causing the protomylonitic texture of the sample….

seismological and experimental studies that aim to unravel the complex fault-slip behaviours at the subduction interface. In this context, our studied sample represents a new and still not yet documented case study of brittle-viscous transition zone and processes in subducted continental crust. Moreover, our study reinforces the concept that the external zones of Alpine Corsica represent an unique target to document and better understand footwall deformation structures and processes related with HP/LT continental subduction.

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FIGURES LIST AND CAPTION

Fig.1 a) Alpine Corsica within the Alps/Apennine framework. b) Tectonic map of north Corsica showing the main tectonic units and the area of studied sample. 1,2) Corsican continental crust, mainly Carboniferous-Permian granitoids, their host pre-Carboniferous basement and a Mesozoic to Eocene cover: (1a) “authochtonous” Hercynian Corsica and (1b) greenschist/lower-blueschist external continental units (Corte; Popolasca); 2) inner continental units: (2a) upper-blueschist units (Tenda Massif; Centuri) and (2b) eclogite slices (Serra di Pigno-Farinole); 3) Schistes Lustres nappe (undifferentiated); 4) Nappe Superiore, i.e. upper non-metamorphic units (Balagne, Nebbio and Macinaggio units); 6) Miocene sediments; c) Regional sketch cross-section ~~with indication of~~with location of the studied~~dy~~ sample from the Popolasca area.

Fig.2. ~~Mesoscopic view of the a~~Analyzed sample: host protomylonitic metagranite ~~host~~ with millimetre-scale ultramylonite localized after brittle precursor. All observable deformation structures were developed under HP/LT metamorphic conditions. Dashed line??
What is the difference between a) and b)? Necessary? If used, it should be commented upon.

Fig.3. ~~Microscopic views~~Microphotographs of the host-rock showing q~~Q~~uartz microstructures typical of low temperature plasticity: a) intracrystalline deformation (undulatory extinction, deformation lamellae, deformation bands, localized extinction bands~~,~~ (LEBs) associated with recrystallization along intragranular conjugate shear bands sets; b) ~~brittle fracturing in~~ domino-style brittle fracturing of K-feldspar and associated ~~with~~ K-feldspar breakdown to albite, stilpnomelane and phengite; c) quartz deformation lamellae and intergranular recrystallization by bulging; d) K-feldspar intergranular recrystallization and albite-neocrystallization, recrystallization also observable along later cracks and microfractures (horizontal to sub-horizontal).

Fig.4 a) BSE image of the metamorphic mineral assemblage in the host granitoids of the studied sample; Ab, albite; Amp, Na-amphibole; Kfs, K-feldspar; Phe, phengite; Qtz, quartz; b) composition of sodic amphiboles, using the classification of Leake et al. (1997); c) estimated peak-metamorphic pressure/temperature conditions for the studied sample (cross-hatched area) constrained by the reactions indicated.

Commentato [GV40]: Figures have to be improved to better document key microstructural relationships.

Commentato [GV41]: Kind of useless inset, really. Improve or remove.

Commentato [GV42]: Please find a way to add this text straight to the figure. Much nicer and easier to read.

Commentato [GV43]: Corte what? Popolasca what? Units? Areas?
Same below

Commentato [GV44]: Well, this is not what can see here!. Please describe what the photo shows

Commentato [GV45]: Indicate the trace of what you call conjugate bands. Far from obvious.

Commentato [GV46]: You have to indicate and label all these features and phases!

Commentato [GV47]: Where?

Commentato [GV48]: You mean from left to right in the section?

Fig.5 a) Microscopic view of millimetre-scale shear zone nucleated after brittle precursor, with indication of the structural-compositional domains within the shear zone. Domain 1 relict domain of cataclasite; Domain 2 phyllonite; Domain 3 ultramylonite; b) detail of the sharp shear-zone boundary heritage of former host-fracture contact and ultramylonite to phyllonite transition within the microscale shear zone are well observable; c) detail of domain 1, relict domain of cataclasite, mainly K-feldspar dominated. Angular fragments of variable size are indicative of brittle comminution.

Fig. 6 ~~-~~Microscopic view of ultramylonite; a) synkinematic Na-amphibole, boudinaged within the quartz and albite ultramylonite matrix; b) Quartz and albite recrystallized matrix, in the lower part the boundary of microscale shear zone to ~~with~~ an host quartz porphyroclast; c) Quartz-ribbon with a strong elongation (1:15, X/Z ratio) defining a shape preferred orientation oblique (15°) to the shear zone boundary; d) detail of quartz recrystallization in the ribbon porphyroclast and matrix.

Fig. 7. EBSD analysis of quartz in the host protomylonite. (a) Microstructure of the analysed site. The white arrows indicate localized extinction bands that have been analysed with the boundary trace analysis and misorientation profiles as shown in Figs. 7b-g. (b) Inverse Pole Figure Map with the respect to the protomylonitic foliation in the host rock (vertical in the figure). Location of subset 1 and 2 and trace of misorientation profiles 1-3 are shown. (c) Pole figure of the host grain, colour coded per the quartz inverse pole figure shown in (b). (d) Pole figure of recrystallized quartz in the intracrystalline bands. The blue and red line indicates the average orientation of the trace of the intracrystalline bands with a subset 1 and 2 orientation, respectively. (e) Boundary trace analysis of the localized extinction band of subset 1. (f) Boundary trace analysis of the localized extinction band of subset 2. (g) Point-to-point misorientation profiles. See (b) for location of the traces of the profiles. (h) Misorientation axis of the host grain and of the recrystallized grains in sample coordinates.

Fig. 8. EBSD analysis of quartz in the ultramylonite. (a) Microstructure of the recrystallized quartz from domain 3 in the ultramylonite. (b) Pole figure of recrystallized quartz grains from (a). Only grains from recrystallized ribbons in the ultramylonite were analysed.

Fig.9. a) Brittle failure analysis for our studied sample, see text for assumptions and approximations. For the differential stress range and the vertical stress considered, brittle failure requires (sub)lithostathic fluid pressure (0.96 < λ < 1) whereas viscous deformation at the estimated depth range (23-30 km) is only possible for pore fluid pressure ≤ 0.94; b) Rheological profile calculated for a fixed strain rate of 10^{-12} s^{-2} (see text for calculation details) using the quartz flow law of Hirth et al.

Commentato [GV49]: ??

Commentato [GV50]: One cannot see this, sorry.

Commentato [GV51]: Where woud domain 3 be exactly?

Commentato [GV52]: All this has to be documented and better illustrated/documented. I have some problems in linkung your words to what the image shoes.

Commentato [GV53]: Document also the quartz clasts. Any textural relationship confirming the later viscous deformation of these catalasite?? That would be crucial for the story.

Commentato [GV54]: Indicate the trace of the mylonitic foliation.

Commentato [GV55]: Show layout of clast or trace boundary

Commentato [GV56]: See my comment in the main text

Commentato [GV57]: Is there an angular discordance between the EW strike of the ribbon and the S in the upper part of the thin section?

Commentato [GV58]: Split this figure into two. Too dense. Enlarge the individual components.

Commentato [GV59]: Indicate on the figure, please.

Commentato [GV60]: Make the numbers larger.

Commentato [GV61]: Link it to a properly defined Dpmain 3 in Fig. 5.

(2001). Cycle of switching in deformation style and mechanisms is suggested for the anayzed sample in its ambient depth-range. Frictional Byerlee envelope is calculated using an average friction coefficient of 0.7 for various pore fluid pressure ratio values (Fig.9a and text). Stress estimates based on recrystallized ~~-~~quartz piezometry (grey shaded area) ~~have been~~were calculated following Stipp and Tullis, (2003).

Fig.10. ~~-~~a) Simplified ~~view~~ model of the ancient subduction of the corsican continental crust, with indication where the mechanical coupling is the highest (seismogenic zone) and lowest (stable slip). The white rectangle corresponds to the inferred zone of the studied sample. The pink to red shading ~~figured~~ plots the lower to higher blueschist- to eclogite-facies peak metamorphism in continental-derived crustal units; b,c,d Multi-stage scenario for development of the association of granitic protomylonites and ultramylonites from brittle precursors in the Popolasca area: a) General sketch showing the east-dipping ancient alpine subduction of the corsican continental crust. b): Formation of a crustal scale anastomosed network of shear zones within the host granitic crust and~~,~~ active distributed deformation with aseismic creep; c: Formation of brittle instabilities (stage 2) in the shear zone by seismic ruptures nucleated along (1) the subduction interface ~~(1)~~ or(2) the core of granites ~~(2)~~ and ~~having propagated~~propagating across and beyond the granitic mylonites, followed by post-seismic creep, with local~~t~~ization of viscous shear zone after brittle precursor (stage 3); d) Formation of brittle instabilities by seismic tremors and slow-slip events followed by post-seismic (post-tremors) creep.