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Microtectonics of low-P low-T carbonate fault rocks

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ABSTRACT

With the aim of deducing some general microtectonic processes responsible for the development of carbonate fault cores, rock samples were collected in ten of such structures, which are different in size, attitude, kinematics, displacement, and tectonic environment. Samples were thin-sectioned and analysed under an optical microscope. Microscopic evidence (i.e., at the scale of tens-to-hundreds of microns) shows that grain size reduction occurred mostly by cataclasis and, occasionally, by pressure solution. Cataclasis involved three main processes here named intragranular extension fracturing, chipping, and shear fracturing. Intragranular extension fracturing is more common in the early stages of cataclasis and produces a coarse breccia consisting of angular grains. In a few cases, pre-existing weaknesses and flaws control the fracture pattern associated with intragranular extension fracturing. Chipping is more common in the advanced stages of cataclasis and produces a gouge consisting of a few survivor rounded grains within a fine matrix. Shear fracturing seems less frequent than the other two processes and usually occurs in the advanced stages of cataclasis. By considering the microscopic and mesoscopic evidence, and the dissimilar frequency of dissolution structures in the analysed fault cores and damage zones, it is inferred that the studied fault zones probably acted as conduit–barrier permeability systems.

1. Introduction

Faults in low-pressure low-temperature carbonate rocks are known both as earthquake foci (Amato et al., 1998; Di Bucci and Mazzoli, 2003; Del Gaudio et al., 2007) and as complex permeability structures within hydrocarbon, water, and geothermal reservoirs (Eberli et al., 2004; Mancini et al., 2004; Mazzullo, 2004; Celico et al., 2006; Rossetti et al., 2007a,b). Their study, at all scales, is therefore relevant for structural geologists dealing with seismic faulting or working in the hydrocarbon, water, and geothermal industries.

Until about 1990, carbonate fault rocks were hardly studied (e.g., Turner et al., 1954; Rutter, 1974; Mimran, 1976, 1977; Friedman and Higgs, 1981) compared to fault-related crystalline and silicoclastic rocks (e.g., Engelder, 1974; Sibson, 1977; Sammis et al., 1986; Sammis and Biegel, 1989; Blenkinsop, 1991). In the last fifteen years, the study of carbonate fault rocks has significantly advanced and become systematic mostly because of its importance in the hydrocarbon industry (Burkhard, 1993; De Bresser and Spiers, 1993; Hadizadeh, 1994; Newman and Mitra, 1994; Babaie et al., 1995; Kennedy and Logan, 1997; Salvini et al., 1999; Graham et al., 2003; Kim et al., 2003; Storti et al., 2003; Llana-Funez and Rutter, 2005; Agosta and Aydin, 2006; Tondi et al., 2006; Agosta et al., 2007; Tondi, 2007). Studies of fault core permeability (Ghisetti et al., 2001; Agosta and Kirschner, 2003; Micarelli et al., 2006; Agosta et al., 2007), grain shape evolution with fault slip (Storti et al., 2007), and some earthquake indicators obtained in laboratory-simulated faults (Han et al., 2007a,b; see Billi and Di Toro, 2008 for a review) have recently improved our understanding of the mechanical and hydraulic behaviour of carbonate fault rocks. In particular, faulting simulations performed at seismic slip rates (about 1 m/s) in Carrara marble revealed very promising results and showed that the seismic (i.e., frictional) process and related indicators have to be investigated at the microscale (Han et al., 2007a,b). Unfortunately, the microtectonics of low-pressure low-temperature fault-related carbonate rocks is still poorly emphasized, and published microscopic images of these rocks are relatively rare (Wenk, 1985; Pieri et al., 2001a,b; Barnhoorn et al., 2004, 2005; Billi, 2005, 2007; Tondi et al., 2006; Tondi, 2007; Billi et al., 2008; Ferrill and Morris, 2008; Mort and Woodcock, 2008). This lack of knowledge prevents advances in the understanding of the processes responsible for the formation of carbonate fault cores and, therefore, in the understanding of the frictional and hydraulic behaviours of these structures.

The main goal of this paper is to contribute in knowing and understanding the microscopic processes that are responsible for the development of carbonate fault cores. To reach this goal,

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microscopic images from low-pressure ($<\sim 100$ MPa) low-temperature ($<\sim 100$ °C) carbonate fault rocks (i.e., fault cores) collected in Italy (Figs. 1–3) are shown and discussed. Insights into the microtectonic processes (i.e., cataclasis and pressure solution) are provided. As the main goal is to address the detailed mechanisms of fracture (or pressure solution) in carbonates, fault rocks in diverse settings were chosen in order to deduce the general processes in all such rocks (Table 1). It should be noted that observations and inferences provided in this paper are valid at the scale of analysis (i.e., tens-to-hundreds of microns; Figs. 4–6).

2. Geological setting

The analysed rocks were collected from exposures of Mesozoic shallow-water organic carbonates located in the central Apennine fold-thrust belt and in the northern Apulian foreland, central Italy (Fig. 1 and Table 1). The central Apennine fold-thrust belt mostly consists of Meso-Cenozoic carbonate thrust sheets accreted in Neogene time toward the Apulian–Adriatic foreland, in the east, during westward subduction of the foreland plate. In late Neogene time, the Tyrrhenian (i.e., western) side of the Apennine belt was extended under a backarc tectonic regime, while toward the east, tectonic accretion was still active at the front of the wedge (Malinverno and Ryan, 1986; Patacca et al., 1992; Faccenna et al., 2004). At present, reduced thickness of the lithosphere, volcanism, extensional basins, and high heat flow characterize the Tyrrhenian side of the Apennine belt and are the results of the Neogene–Quaternary backarc extensional process (Funiciello et al., 1976; Barchi et al., 1998; Jolivet et al., 1998; Billi et al., 2006; Nicolosi et al., 2006). In the central Apennines, thrust imbrication occurred mostly in a forelandward piggyback sequence, with a few out-of-sequence or backward thrusting episodes (Ghisetti and Vezzani, 1997; Cavinato and DeCelles, 1999). Post-orogenic normal faults and associated extensional basins of Miocene–Pleistocene age are widespread both in the Tyrrhenian side of the Apennines and in the axial sector of the fold-thrust belt (Keller et al., 1994; Lavecchia et al., 1994; Barchi et al., 1998, 2007; Jolivet et al., 1998; Cavinato et al., 2002). The locus of extension progressively migrated toward the east, parallel but west of the eastward-migrating locus of contractional deformation (Elter et al., 1975; Malinverno and Ryan, 1986; Carmignani and Kligfield, 1990; Patacca et al., 1992). The lag time between the onset of thrusting and initial extension at any given locality in the central Apennines is about 2–4 m.y. (Cavinato and DeCelles, 1999).

The Gargano Promontory (Fig. 1) is a structural high located in the Apulian–Adriatic foreland (Favali et al., 1993; Doglioni et al., 1994; Brankman and Aydin, 2004). The promontory consists of a thick succession of Mesozoic carbonates dissected by an active and complex fault array. Within this array, the strike-slip Mattinata Fault in the southern Gargano Promontory is the most prominent fault (Ortolani and Pagliuca, 1987; Funiciello et al., 1988; Salvini et al., 1999; Brankman and Aydin, 2004). Geological and geophysical evidence shows that the Mattinata Fault was activated during late Miocene time at the latest and is still active being the source of recent and historical earthquakes (Favali et al., 1993; Salvi et al., 1999; Patacca and Scandone, 2004; Tondi et al., 2005; Billi et al., 2007a).

The exact thermobaric regime experienced by each analysed fault is unknown; however, upper thermobaric boundaries for the central Apennines can be inferred from constraints obtained after organic matter maturity, clay mineralogy, stratigraphy, and structural geology studies. These results suggest that the investigated exposures experienced thermobaric conditions below the metamorphic regime (i.e., below a temperature of 200 °C and a pressure of 200 MPa) as also

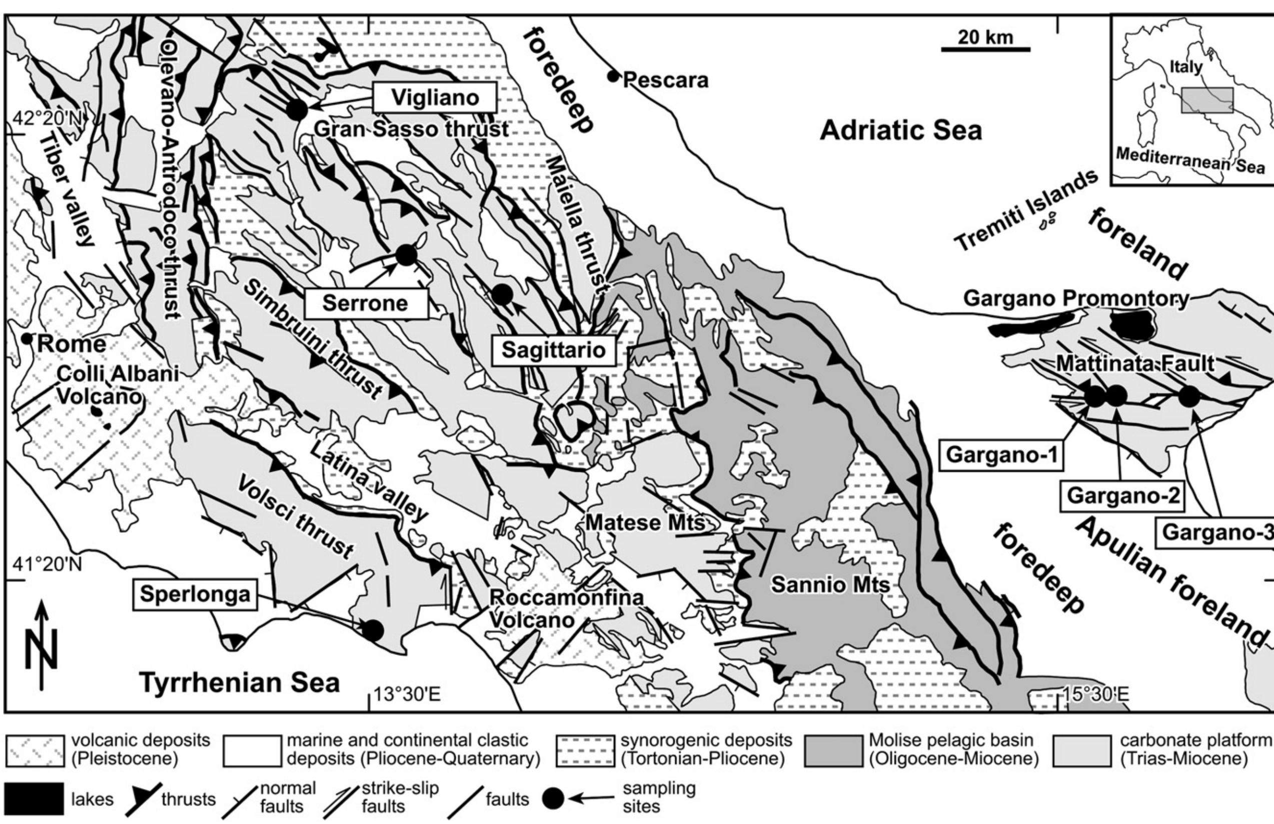

Fig. 1. Geological map of central Italy (modified after Bigi et al., 1991; Cavinato and DeCelles, 1999). Locations of the studied faults are displayed with black dots. See Table 1 for coordinates of fault locations and for fault main attributes.

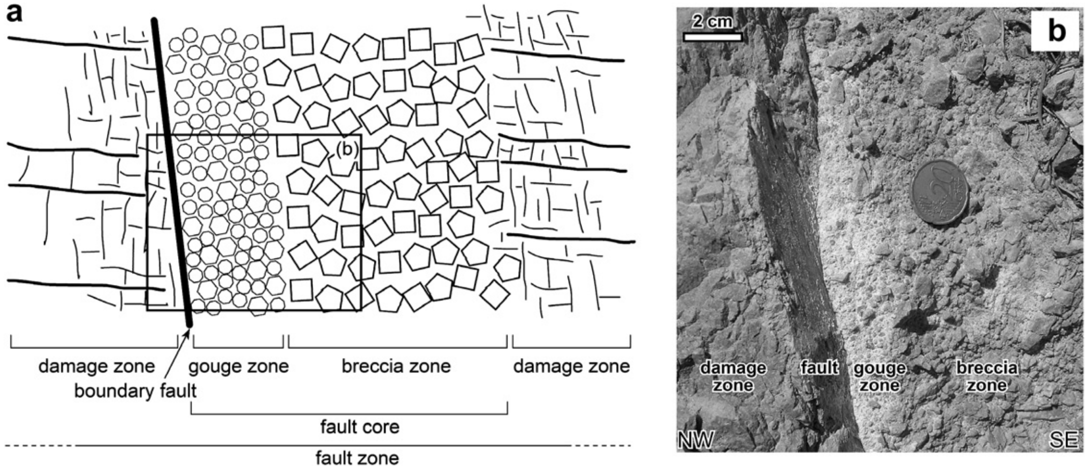

Fig. 2. (a) Conceptual cross-sectional sketch of a fault zone sectioned perpendicularly to the fault surface. The fault core includes a gouge zone and an adjacent breccia zone and it is bounded on one side by the fault surface (boundary fault). On the other side, the transition between the breccia zone and the damage zone is usually rather abrupt. The fault core is included within a damage zone. Damage zones are here considered part of the host rock, which includes also the protolith containing no fault-related deformations. (b) Photograph showing a transpressive fault core from the Sperlonga site.

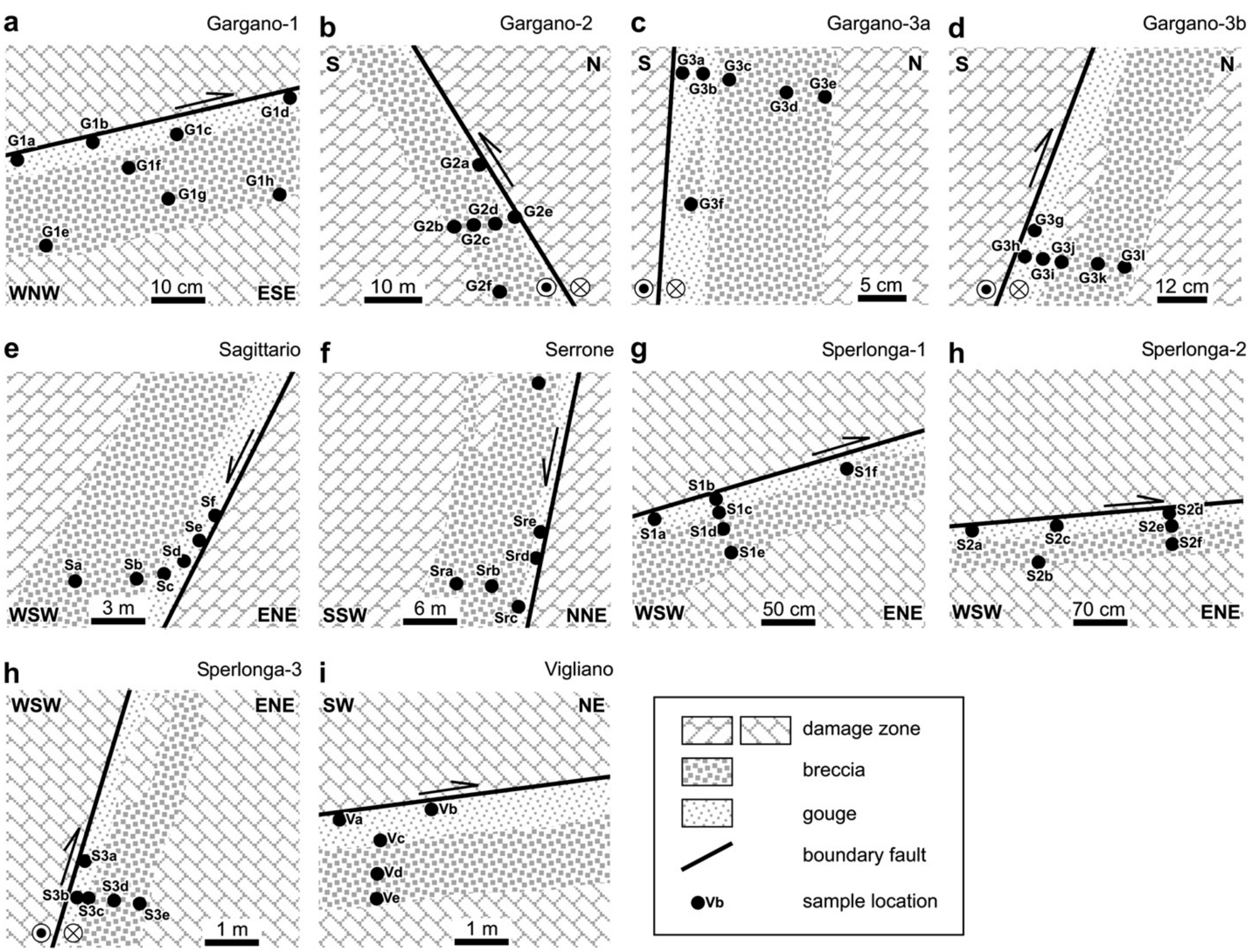

Fig. 3. Schematic cross-sectional sketches of the studied fault cores. These sketches are drawn to show the sample location.

Table 1
Sampling sites and main attributes of related fault cores.

Site	Lat. N	Long. E	Tectonic environment	Host rock type	Host rock age	Fault type	No. of analysed fault cores	Estimate of fault displacement	Samples	Thin-sections
Gargano-1	41°, 45′	15°, 36′	Foreland	Dolomitic limestone	Mesozoic (Jurassic)	Reverse	1 (Gargano-1)	ca. 10 m	G1a, G1b, G1c, G1d, G1e, G1f, G1g, G1h	Figs. 4c and 5b
Gargano-2	41°, 45′	15°, 41′	Foreland	Dolomitic limestone	Mesozoic (Jurassic)	Transpressive	1 (Gargano-2)	10–100 m	G2a, G2b, G2c, G2d, G2e, G2f	Fig. 5a
Gargano-3	41°, 45′	15°, 55′	Foreland	Limestone	Mesozoic (Jurassic)	Strike-slip and transpressive	2 (Gargano-3a, Gargano-3b)	2–2000 m	G3a, G3b, G3c, G3d, G3e, G3f, G3g, G3h, G3i, G3j, G3k, G3l	Figs. 4a and 5c, d
Sagittario	41°, 58′	13°, 56′	Post-orogenic extension	Dolomitic limestone	Mesozoic	Extensional	1 (Sagittario)	100–400 m	Sa, Sb, Sc, Sd, Se, Sf	
Serrone	44°, 04′	13°, 37′	Post-orogenic extension	Limestone	Mesozoic	Extensional	1 (Serrone)	ca. 1000 m	Sra, Srb, Src, Srd, Sre	Fig. 4b
Sperlonga	41°, 15′	13°, 26′	Thrust-fold belt	Dolostone	Mesozoic (Jurassic)	Reverse and transpressive	3 (Sperlonga-1, Sperlonga-2, Sperlonga-3)	10–100 m	S1a, S1b, S1c, S1d, S1e, S1f, S2a, S2b, S2c, S2d, S2e, S2f, S3a, S3b, S3c, S3d, S3e	Figs. 4d–f, 5e, f and 6a–f
Vigliano	42°, 24′	13°, 16′	Thrust-fold belt	Dolomitic limestone	Mesozoic	Reverse	1 (Vigliano)	ca. 100 m	Va, Vb, Vc, Vd, Ve	

suggested by the lack of exposed metamorphic rocks in the study areas and in the adjacent ones (Accordi and Carbone, 1986; Bigi et al., 1991). In particular, most rocks exposed in the central Apennines probably developed under a lithostatic maximum load imposed by a rock column of less than about 3–4 km. For carbonate rocks, such a rock column (i.e., 4 km) corresponds to a lithostatic stress of about 100 MPa. Assuming a geothermal gradient of 25 °C km^{-1}, the studied fault cores likely experienced a maximum temperature of about 100 °C (Corrado, 1995; Corrado et al., 1995, 1998). From stratigraphic evidence (Ricchetti et al., 1992), the same thermobaric limits (i.e., a lithostatic load smaller than about 100 MPa and a temperature of less than about 100 °C) are also inferred for the exposures of the Mattinata Fault in the Apulian foreland.

Sedimentary evidence (i.e., paleoenvironmental studies) and rock ages show that exhumation of the central Apennines started at the end of Tortonian time from the Tyrrhenian side, and then propagated toward the east during late Neogene time. This process was mostly accomplished by the end of Pliocene time (Cipollari and Cosentino, 1995; Cosentino et al., 2002); however, the uplift in the central Apennines is still ongoing (Ghisetti and Vezzani, 2002). With regards to the Apulian foreland, exhumation occurred as a consequence of subduction and flexure-related peripheral bulging of the foreland plate during Plio–Pleistocene times (Royden et al., 1987).

3. Sampling and fault zone structure

For this study, 59 thin-sections from 35 samples of fault-related rocks collected in 7 sites in central Italy were analysed (Fig. 1). In each site, one-to-three fault cores (Fig. 2) were sampled for a total of ten fault cores (Table 1). The analysed samples were taken from either the gouge or breccia zones (Fig. 2) forming the studied fault cores. Particular care was used in collecting and transporting these samples from the field sites to the laboratory to avoid deformation and disaggregation. In the laboratory, the samples were impregnated and stabilized with epoxy resin. Fig. 3 shows schematic sketches of the studied fault cores and the related sample location. Orientations of thin-sections with respect to the fault surface and slip direction are provided in the captions to Figs. 4–6, in which 18 microphotographs, taken from the analysed thin-sections, are displayed.

The fault cores (Fig. 3) consist of fine-grained fault rocks reflecting both shear deformation and comminution (Fig. 2). In the fault cores, any pre-existing rock fabric (e.g., bedding and compaction stylolites) was obliterated by the fault-related shear deformation. Within each fault core, a thin gouge zone (i.e., where less than 30% of the grains are not visible at the naked eye; Scholz, 1990) runs alongside the slip surface (boundary fault in Fig. 2a). On the other side, distant from the slip surface, the gouge layer is bounded by a breccia zone consisting, on average, of grains larger than those within the gouge zone (Billi et al., 2003a,b, 2008; Storti et al., 2003). Gouge zones occasionally contain lenses or pockets of breccia. The gouge and breccia zones together form the fault core (Fig. 2). No analysed fault cores are foliated or crosscut by dissolution-related structures. Veins filled by calcite or other minerals are rarely observed in the fault cores. Both gouge and breccia zones of the fault cores are generally well compacted but poorly-cemented. This is consistent with the model suggested by Sibson (1990), in which non-cohesive fault-related gouge and breccia are typical of faults located at crustal depths between ca. 0 and 4 km.

The rock flanking the fault cores is here indicated as host rock and includes both the fault damage zone, which is characterized by fault-related closely-spaced fractures and dissolution cleavages (e.g., Salvini et al., 1999; Billi and Salvini, 2001; Billi et al., 2008), and the protolith, which is the rock without fault-related deformation. The former is adjacent to the fault core, whereas the protolith may occur at distances between a few meters (2 m at least) and a few hundreds of meters from the fault core (up to a maximum of about 300 m in the case of the Mattinata Fault; Salvini et al., 1999). Lenses or pockets of fractured host rock occasionally occur within the breccia or gouge zones. Exposed fracture and cleavage surfaces in the host rock are often karstified (i.e., dissolutionally weathered; e.g., Billi et al., 2007b, 2008), whereas no evidence of karstification occurs in the fault cores, especially in the gouge zones. Host rocks are all Mesozoic shallow-water organic carbonates (Table 1) typical of central Italy. The main physical parameters of these rocks are generally as follows (e.g., Agosta and Aydin, 2006): bulk modulus (K) ≈ 70 GPa, shear modulus (μ) ≈ 28 GPa, Poisson's ratio (ν) ≈ 0.3, P-wave velocity (V_p) ≈ 6 km s^{-1}, and S-wave velocity (V_s) ≈ 3 km s^{-1}.

Displacements along the study faults are different, estimated between about 2 m and 2 km (Table 1) by considering either the fault length–displacement ratio (e.g., Bonnet et al., 2001; Xu et al., 2006, and references therein) or the fault core thickness–displacement ratio (Scholz, 1987; e.g., Micarelli et al., 2006).

4. Microscopic observations

Observations of thin-sections obtained from fault rock samples were done with a Nikon (Eclipse 50i Pol) optical microscope at various magnifications (2, 4, 10, 20, and 40×).

In the analysed thin-sections, three main types of microscopic cataclastic fabrics were recognised and named as follows:

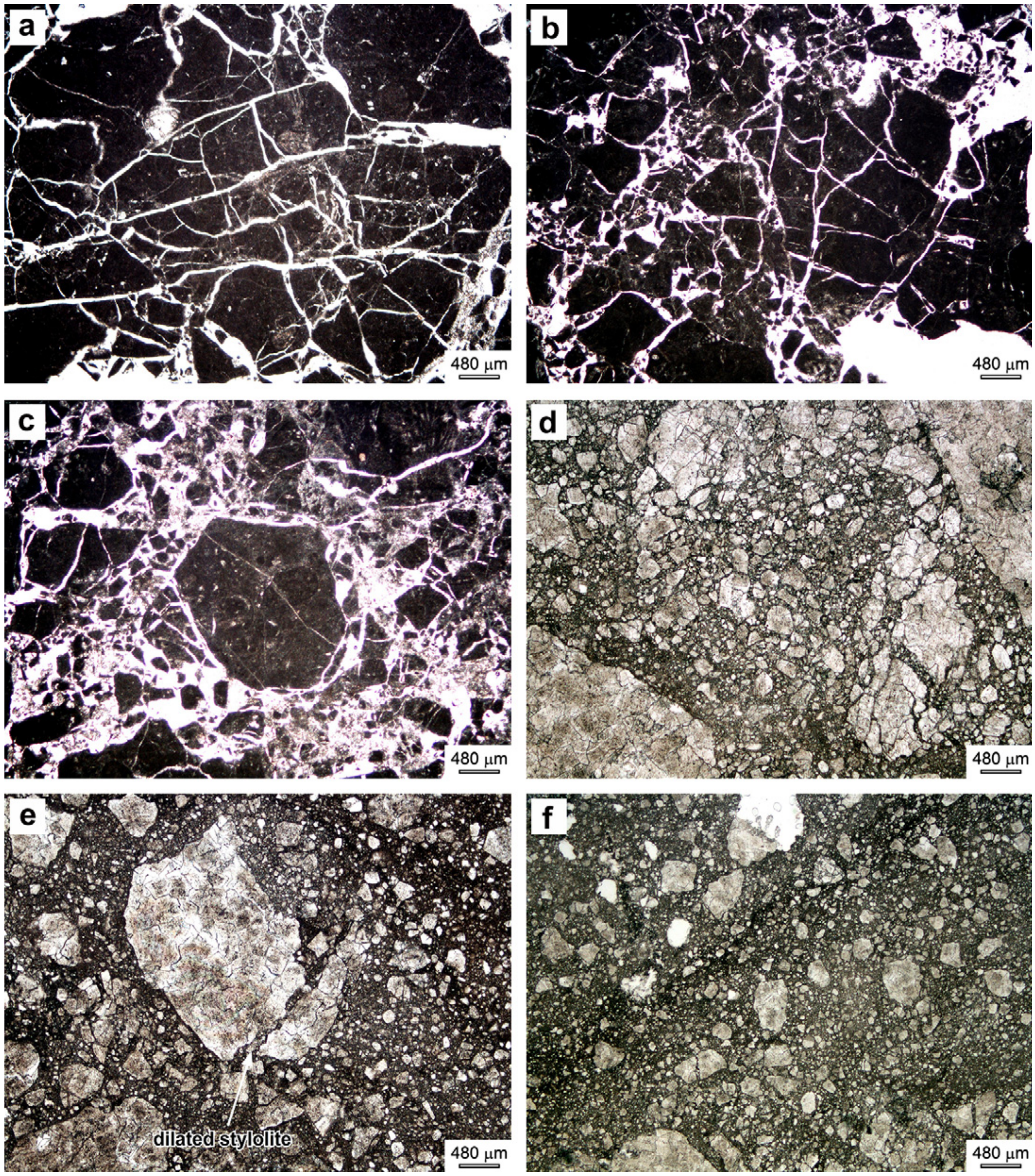

Fig. 4. Microphotographs (parallel nicols) showing the cataclastic fabrics observed in the analysed fault rocks. In (a) and (b), coarse and angular grains are in contact with one another (embryonic cataclastic fabric). Small grains forming a fine matrix are almost absent. Note the network of extensional fractures. In (c) and (d), coarse grains start to be rounded and surrounded by a fine matrix (intermediate cataclastic fabric). In (e) and (f), coarse grains are rare and rarely in contact with one another because an abundant fine matrix surrounds them (mature cataclastic fabric). Note in (e) a dilated stylolite; (a) is from sample G3l. The thin-section is normal to fault surface and to fault slip; (b) is from sample Srb. The thin-section is normal to fault surface and parallel to fault slip; (c) is from sample G1f. The thin-section is normal to fault surface and parallel to fault slip; (d) is from sample S2e. The thin-section is normal to fault surface and parallel to fault slip; (e) is from sample S3b. The thin-section is normal to fault surface and to fault slip; (f) is from sample S1a. The thin-section is parallel to fault surface.

embryonic, intermediate, and mature cataclastic fabrics (Fig. 4). These fabrics were observed in all the fault cores.

The embryonic fabric is typical of coarse breccia zones near the boundary with the damage zone and consists of an assemblage of large fractured angular grains, which are in contact with one another (Fig. 4a and b). A fine matrix is either not developed there or it is in a very incipient stage of development. The grain size distribution is generally well sorted toward the largest sizes.

The intermediate fabric is also typical of the breccia zones (i.e., the portion lying near the gouge zone). It consists of an assemblage of several large grains, which are partly in contact with one another and partly surrounded by a fine matrix (Fig. 4c and d). The large grains are less angular than those observed in the embryonic fabric (Fig. 4a and b). The grain size distribution of samples with an intermediate fabric is less sorted than that observed in samples with an embryonic fabric (e.g., Storti et al., 2003; Billi et al., 2003a; Billi, 2007).

The mature fabric is typical of gouge zones and consists of an assemblage of few large grains surrounded by a well developed and abundant fine matrix (Fig. 4e and f). Large grains are rarely in contact with one another and are generally well rounded (e.g., Heilbronner and Keulen, 2006; Storti et al., 2007). Grain sorting within samples with a mature fabric is usually poor; the grain size distribution is shifted toward the small grains as compared with that typical of the breccia zones (i.e., embryonic and intermediate fabrics) (e.g., Billi, 2007).

Microscopic evidence indicates that grain size reduction occurred (1) usually by cataclasis (i.e., by fracturing and rotation;

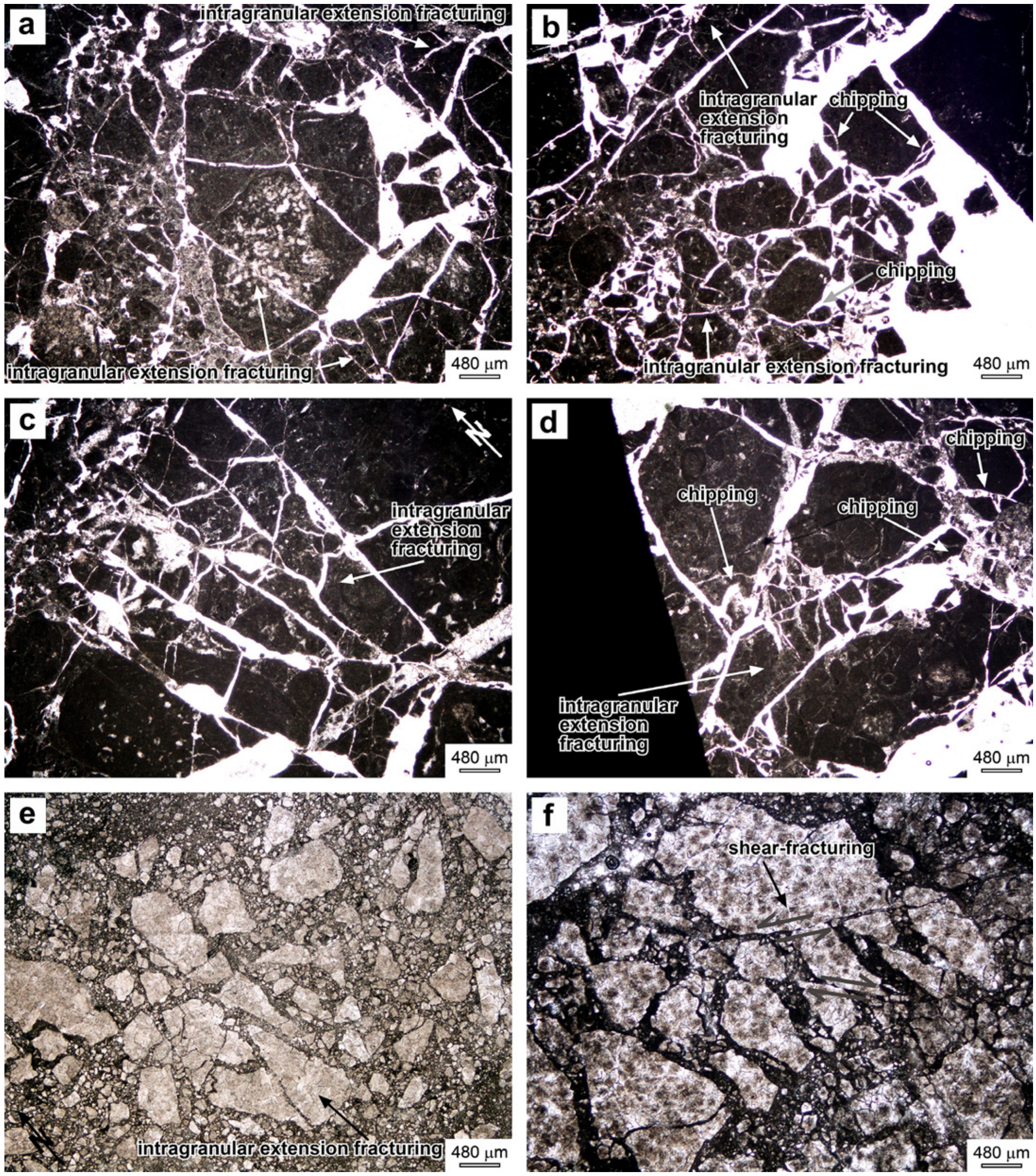

Fig. 5. Microphotographs (parallel nicols) showing examples of physical processes of grain size reduction within the analysed fault rocks. (a) Intragranular extension fracturing of a coarse grain occurring in the photograph centre. Note that some grains are affected by non-systematic fractures and some others by subparallel ones. (b) Intragranular extension fracturing and chipping contributed to form the shown sample. Note the poorly-systematic network of extensional fractures. (c) Intragranular extension fracturing determined the formation of N-S-elongated slivers (orientation is fictitious), which were successively affected by orthogonal fractures. (d) Examples of grain chipping. (e) Examples of intragranular extension fracturing in a sample characterized by an intermediate cataclastic fabric consisting in some coarse grains partly surrounded by an incipient fine matrix. (f) Intragranular extension fracturing and development of shear fractures; (a) is from sample G2b. The thin section is normal to fault surface and parallel to fault slip; (b) is from sample G1h. The thin-section is normal to fault surface and parallel to fault slip; (c) is from sample G3l. The thin-section is normal to fault surface and parallel to fault slip; (d) is from sample G3e. The thin-section is normal to fault surface and to fault slip; (e) is from sample S1c. The thin-section is normal to fault surface and to fault slip; (f) is from sample S2c. The thin-section is normal to fault surface and parallel to fault slip.

e.g., Engelder, 1974) and (2) occasionally by pressure solution (i.e., dissolution of grain boundaries caused by increasing stress in the presence of water-rich fluids; e.g., Gratier and Gueydan, 2007).

(1) In this study, cataclasis includes at least three main physical processes of grain size reduction, named as follows: intragranular extension fracturing (e.g., Gallagher et al., 1974; Hadizadeh and Rutter, 1982), chipping, and shear fracturing (Figs. 4–6). Microscopic evidence of these processes was detected in all the studied fault cores.

By intragranular extension fracturing, the original structure of grains is fully destroyed by fracturing (i.e., extensional fractures). The size of the produced grains generally ranges in about the same order of magnitude as the original grain. Very small grains are rarely the product of intragranular extension fracturing, which is typical of breccia zones (Figs. 4a, b and 5a–c). Two types of intragranular extension fracturing were recognised on the basis of the fracture systematics. In the first type, several non- or poorly-systematic fractures (as observed in two-dimensional views) separate the original grain into a few angular and irregular grains (Figs. 4a, b and 5a, b, and e). A particular case of this type of intragranular extension fracturing was observed in the fault-related dolomitic rocks collected in the Sperlonga site (Fig. 6; Billi et al., 2008). These rocks are characterized by

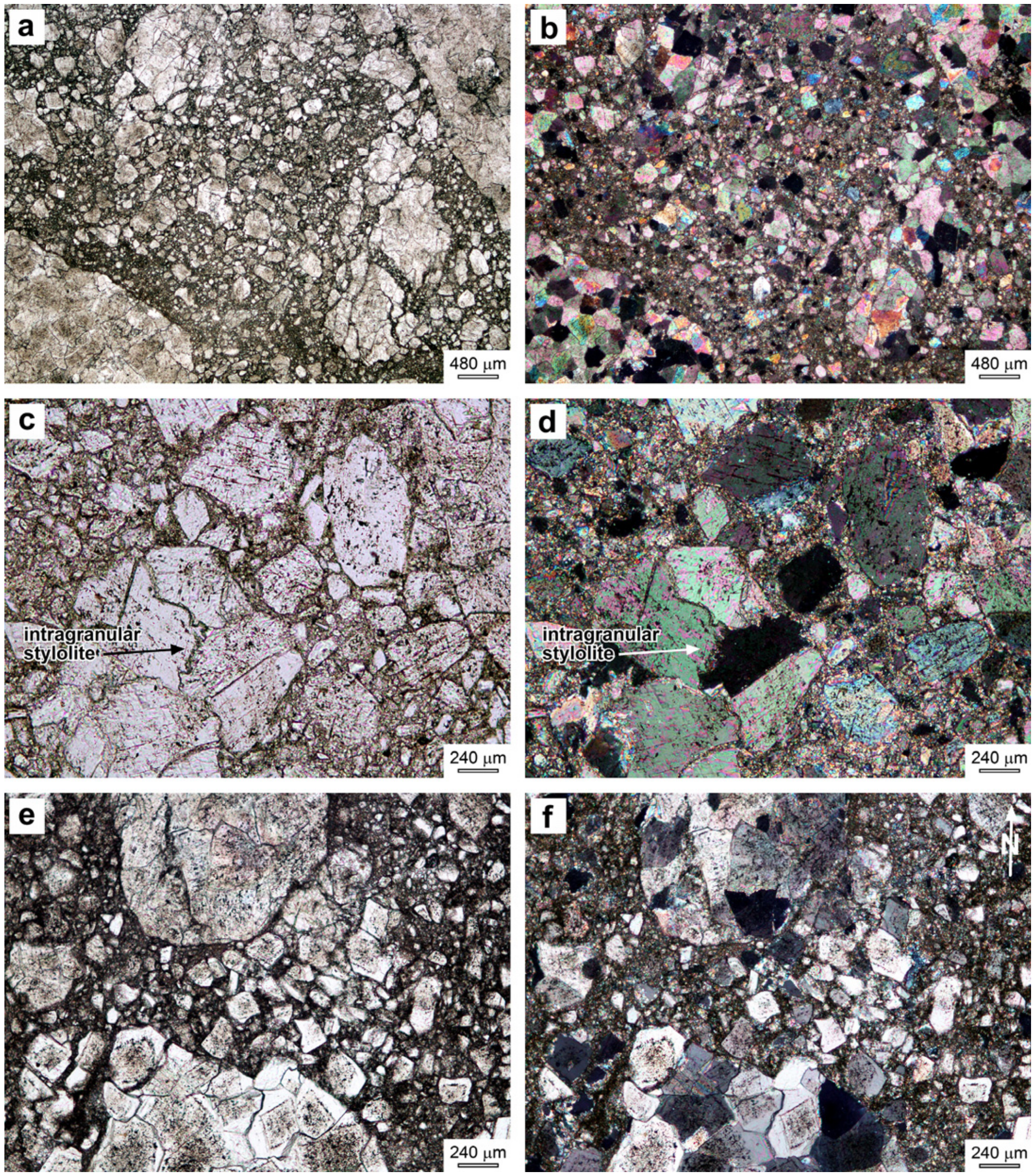

Fig. 6. Microphotographs showing the relationship between crystal and fracture patterns in dolomitic fault rocks from the Sperlonga site; (a), (c), and (e) show microphotographs under parallel nicols, whereas (b), (d), and (f) show the same views but taken under crossed nicols. Note in all samples the recrystallized fabric. Note also that most fractures occur right along the crystal boundaries as inferred from the comparison between parallel and crossed nicols views. It follows that most small grains are monocrystalline (i.e., at the microscope resolution). Note in (c) and (d), an intragranular stylolite; (a) and (b) are from sample S3d. The thin-section is normal to fault surface and parallel to fault slip; (c) and (d) are from sample S1e. The thin-section is normal to fault surface and parallel to fault slip; (e) and (f) are from sample S3d. The thin-section is normal to fault surface and parallel to fault slip.

a recrystallized fabric connected with the secondary dolomitization of an organic limestone. The dolomitization is not connected with faulting because it involves the entire host rock (i.e., also the rock distant from the faults). In these rocks, each large grain includes a few dolomite crystals easily discernible under crossed nicols (Fig. 6b, d, and f). The boundaries between these crystals are sharp mechanical discontinuities (i.e., incipient intragranular extensional fractures) as inferred from the comparison between the same images taken under parallel and crossed nicols (Fig. 6). Intragranular extension fracturing of these grains occurs by fracturing taking place mostly along the crystal boundaries. As a consequence, the small grains derived from the comminution of the large ones (i.e., the small grains surrounding the large ones; e.g., Fig. 6b and f) are mostly monocrystalline (i.e., at the microscope resolution).

In the second type of intragranular extension fracturing, original grains are fragmented into long slender pieces by roughly systematic fractures (i.e., subparallel fractures as observed in two-dimensional views; Figs. 4a and 5c, d). Because of their elongated shape, these long pieces are then easily broken into near-cubic (i.e., near-square in two-dimensional views) or slightly elongated fragments by fractures perpendicular to the long geometrical axis (e.g., Fig. 5c and d; see also Billi et al., 2003a). The fragments produced during this type of intragranular extension fracturing are angular and characterized by dimensions on the same order of magnitude as the original grain. Very small grains are rarely the product of this process.

Small fragments are produced from the abrasion of corners and edges of angular grains by chipping (Figs. 4c, e and 5b, d).

Both bulk structure and size of original grains are only slightly changed by chipping processes and the original grains tend to become rounded (e.g., Heilbronner and Keulen, 2006; Storti et al., 2007). Grain chipping produces several fragments, which are from one to few orders of magnitude smaller than the original grains and form the fine-grained matrix. Chipping is typical of fault rocks characterized by either an intermediate or a mature cataclastic fabric.

An original grain is cut by one or more shear fractures into two or several angular grains by shear fracturing. The size of the resulting grains depends both on the number of shear fractures and on the location where these fractures cut across the original grain. Due to their elongated shape, grains produced by shear fracturing are often affected by perpendicular extensional fractures that produce near-cubic (i.e., near-square in two-dimensional views) or slightly elongated fragments (Fig. 5f). Shear fracturing is only occasionally observed in the samples collected from cataclastic rocks with a mature fabric and results less frequent in the fault rocks characterized by embryonic or intermediate cataclastic fabrics than in the rocks characterized by a mature fabric (e.g., Sammis et al., 1987).

(2) In this study, microscopic evidence of pressure solution includes intragranular and intergranular structures. Intragranular structures, such as stylolites, are common in several grains and generally constitute weakness surfaces along which the grain breaks apart when undergoing cataclasis. These structures do not propagate through the surrounding matrix (Figs. 4e and 6c, d) and are therefore inherited from the previous sedimentary and tectonic history. Intergranular structures, such as matrix stylolites or indented grain-to-grain boundaries are very rare. Few of these structures were found in only 7 out of 59 analysed thin-sections. Stylolites affecting sample matrix are usually unfilled structures possibly dilated because of cataclasis or because of sampling and handling (Fig. 6a–e). In one case, very narrow selvages of insoluble residue were observed along the stylolite walls (Fig. 6d). In another case, a dilated stylolite is filled by secondary calcite (Fig. 6f). No further evidence of intergranular veins (including veins and dilated calcite-filled stylolites) was observed.

5. Discussion

As already stated, the observed microscopic fabrics are typical of specific fault core components. The embryonic fabric is typical of the breccia zone, the intermediate fabric is typical of the portion of the breccia zone lying near the gouge zone, and the mature fabric is typical of the gouge zone. Occurrences of lenses and pockets of breccia within the gouge zones (e.g., Billi et al., 2008), as well as the results of several laboratory simulations of faults (e.g., Mandl et al., 1977; Marone and Scholz, 1989), show that gouge zones usually develop from pre-existing (i.e., early) breccia zones by grain size reduction (e.g., Engelder, 1974; Billi and Storti, 2004; Sammis and King, 2007). It follows that the observed embryonic, intermediate, and mature cataclastic fabrics probably represent three subsequent evolutionary stages of developing carbonate fault cores. Using this assumption, the microscopic observations presented in this paper are synthesized in the following temporal model for the evolution of carbonate fault rocks (Fig. 7).

In the early deformational stages (Fig. 7b–d), large angular grains are in contact with one another and therefore cannot, or can barely, rotate. Lithostatic and tectonic stresses are transmitted through a limited number of contact points among large grains. This grain-to-grain contact configuration leads to development of indentation stresses, which, combined with the presence of possible pre-existing weak surfaces or flaws (e.g., stylolites, organic structures, and crystal boundaries; Figs. 4e and 6c, d), cause the fragmentation of the grain mainly by intragranular extension fracturing and, occasionally, by shear fracturing. In the case of intragranular extension fracturing by subparallel fractures, fractures orthogonal to the long axis of the elongated fragments (Fig. 7b and c) are probably induced by the marked shape anisotropy. The elongated fragments, in fact, may easily be flexed during shear deformation and, subsequently, be fractured perpendicularly to the long axis and to the flexure-related tensile fibre stress (e.g., Turcotte and Schubert, 1982; Engelder, 1987; Billi et al., 2003a). This process significantly increases the fracture connectivity and, therefore, the fluid transmissibility through the rock.

With the disappearance of several large grains by intragranular extension fracturing, and with the development of an embryonic fine matrix (Fig. 7d and e), the degree of spatial freedom of the large grains increases. In particular, the absence of large neighbouring grains allows the large survivor grains to rotate and roll under the effect of the fault slip (Fig. 7e). Grain rotation and the rolling, together with grain sliding, are probably the main causes of chipping and subsequent rounding (i.e., by abrasive wear) of large grains (Fig. 7f) (e.g., Heilbronner and Keulen, 2006; Storti et al., 2007). Shear fractures seem more frequent when a fine matrix develops (i.e., mature cataclastic fabric). This may be connected with a different state of stress undergone by the grains and with the fact that the development of a fine matrix allows the tectonic and lithostatic stresses to be transmitted through a greater number of contact points among grains. Further speculations upon this subject are not possible because the number of observations is scarce.

The above-discussed model of fault core development is somewhat supported by recent numerical models of fault gouge evolution (Guo and Morgan, 2007, 2008). In these models, the evolution of gouge zones with slip is split into two subsequent stages. In the early and shorter stage, shear strain is accommodated primarily by extensional fractures and related grain comminution (intragranular extension fracturing?). During this stage, the shear zone attains a peak friction value and porosity varies greatly (i.e., either increasing or decreasing) as a function of normal stress. In the subsequent and longer stage, grain rolling and sliding become more significant and fault gouge accumulates (chipping?). During this stage, the sliding friction, porosity, and average grain size tend to decrease with slip that is mostly accommodated in the gouge zone.

The general paucity of dissolution structures in the observed samples suggests that pressure solution through the analysed fault cores has been absent or minimal, regardless of lithology, fault type and displacement, and tectonic environment. This inference is supported by the substantial non-cohesive status of the fault core rocks (Ngwenya et al., 2000), and also by the almost total absence of veins and karst features in the fault cores. The cause for the general lack of dissolution structures is probably due to the low porosity and permeability of the fault cores, which inhibit mass transfer and, hence, rock dissolution, whereas dissolution structures (e.g., solution cleavage and karst features) are observed in the surrounding damage zones (Salvini et al., 1999; Billi and Salvini, 2001; Graham et al., 2003; Billi et al., 2008). This evidence suggests that the analysed fault zones mainly acted as conduit–barrier permeability systems (Caine et al., 1996; Main et al., 2000). In such systems, a strong hydraulic anisotropy between a permeable damage zone (and also a rather permeable breccia zone) and an impermeable gouge zone can be the origin of fault-parallel preferential pathways for fluids, which migrate through the damage and breccia zones, and of a fault-perpendicular barrier to fluids, whose flow across the fault core is impeded by the gouge zone.

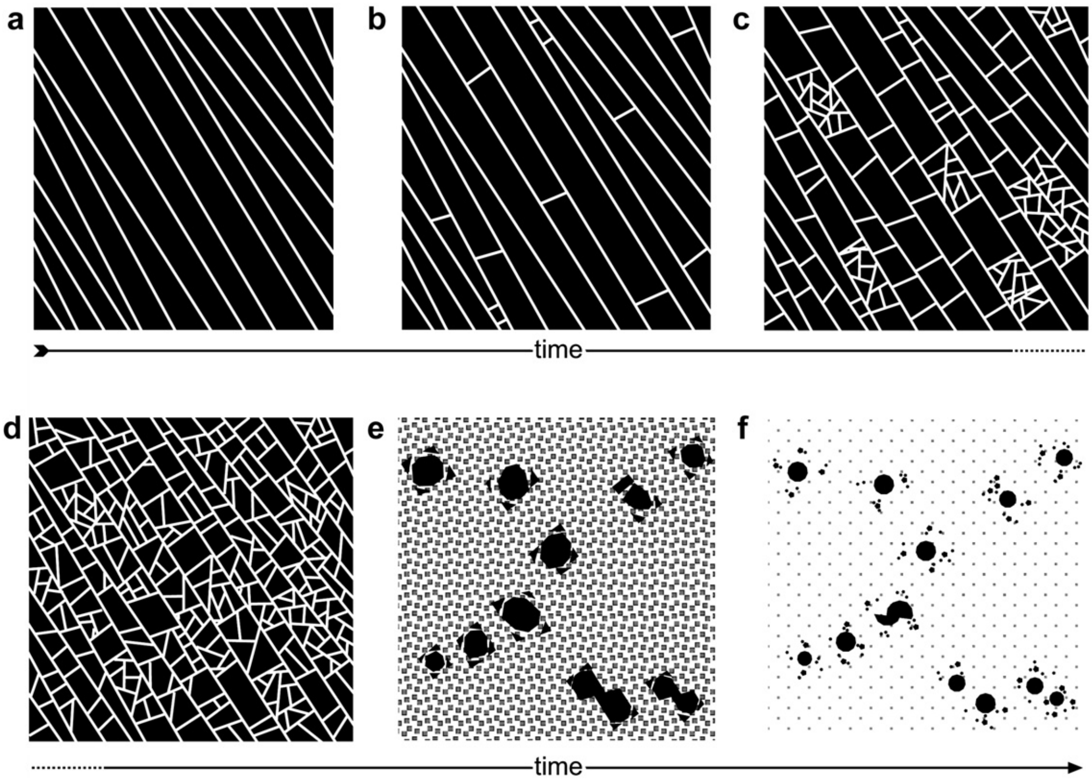

Fig. 7. Cartoon showing an evolutionary model of comminution for the studied carbonate fault rocks. (a) Damage zone affected by subparallel deformations such as joints or solution cleavage surfaces. (b)–(d) Progressive cataclastic deformation mainly accomplished by intragranular extension fracturing. (e) and (f) Progressive cataclastic deformation mainly accomplished by chipping and occasionally by shear fracturing.

6. Conclusions

(1) Regardless of the different size, attitude, kinematics, displacement, host rock, and tectonic environment, the microscopic evidence shows that all the studied fault cores, which experienced a low thermobaric regime (i.e., $P < 100$ MPa and $T < 100\,^{\circ}$C), developed mainly by cataclasis and, subordinately, by pressure solution.

(2) Intragranular extension fracturing is recognised as the main cataclastic process operating during the early stages of deformation. This process leads to embryonic and intermediate cataclastic fabrics, which are typical of breccia zones. Chipping and, subordinately, shear fracturing are recognised as the main cataclastic processes operating during advanced stages of deformation, which leads to the mature cataclastic fabric typical of gouge zones.

(3) The distribution of dissolution structures in the fault cores and damage zones, as well as the inferred cataclastic processes, suggests that the analysed fault zones acted as conduit–barrier permeability systems, in which damage and breccia zones constituted fault-parallel conduits for fluid flow, whereas gouge zones constituted fault-perpendicular barriers.

(4) Limitations to the above-mentioned conclusions are mainly connected with the representativeness of the analysed samples and with the resolution of the optical microscope. Additional microscopic data are therefore necessary in the future to better address the cataclastic processes dealt within this paper.

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