

## Geological-structural mapping and geochronology of shear zones: A methodological proposal

### Mapeo geológico-estructural y geocronología de zonas de cizalla: Una propuesta metodológica

Forero Ortega, Anny Julieth; López Isaza, Julián Andrés; López Herrera, Nelson Ricardo; Cuéllar Cárdenas, Mario Andrés; Cetina Tarazona, Lina Mara; Aguirre Hoyos, Luis Miguel

**Anny Julieth Forero Ortega**

ajforte056@gmail.com

Servicio Geológico Colombiano, Colombia

**Julián Andrés López Isaza**

Servicio Geológico Colombiano, Colombia

**Nelson Ricardo López Herrera**

Servicio Geológico Colombiano, Colombia

**Mario Andrés Cuéllar Cárdenas**

Servicio Geológico Colombiano, Colombia

**Lina Mara Cetina Tarazona**

Servicio Geológico Colombiano, Colombia

**Luis Miguel Aguirre Hoyos**

Servicio Geológico Colombiano, Colombia

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**Abstract:** The deformation registered in rocks in the field can be characterized based on the structures preserved in outcrops, which can be related to wide discontinuity zones named faults and shear zones. The geological-structural mapping and the geochronology of these tectonic structures are a topic of great interest not only for tectonic modeling but also for reconstruction of the geological evolution of the national territory. The methodology suggests for the analysis of faults and shear zones is based on eight steps, including: 1) definition of the geological context in which the structure was developed; 2) photointerpretation, image geoprocessing, and geological-structural mapping of the structural and lithological characteristics of the faults and shear zones; 3) petrographic analysis of field-oriented samples; 4) quantification of strain orientation and geometry through 3D finite strain analyses and quantification of non-coaxiality of deformation through vorticity analyses; 5) SEM-TEM-EBSD microanalysis; 6) quantification of the P-T conditions of deformation through phase-equilibria modeling or conventional geothermobarometry; 7) dating of syn-kinematic mineral phases and mylonitic rocks through Ar-Ar analyses, in order to determine the reactivation and deformation ages of the structure, respectively, as well as the implementation of the U-Pb technique in syn-kinematic calcite crystals developed in the fault planes; and 8) dating of geological elements adjacent to the structure, such as syn-kinematic intrusive bodies associated with the deformation event using zircon U-Pb dating, rocks hydrothermally altered through Ar-Ar method, and zircon and apatite fission-tracks dating of the blocks adjacent to the faults for determining exhumation ages.

**Keywords:** Shear zones, fault rocks, brittle regime, brittle-ductile transition, ductile regime, mapping, geochronology.

**Resumen:** La deformación registrada en las unidades geológicas en campo se puede caracterizar a partir de las estructuras encontradas en afloramientos, las cuales se relacionan, generalmente, con amplias zonas de discontinuidad denominadas fallas y zonas de cizalla. El mapeo geológico-estructural y la geocronología de estas estructuras tectónicas

es un tema de gran interés, no solo para el modelamiento tectónico, sino para la reconstrucción de la evolución geológica de un territorio. Metodológicamente, para el mapeo geológico-estructural y la geocronología de fallas y zonas de cizalla se sugieren ocho fases, que incluyen: 1) definición del contexto geológico en el cual se desarrolló la estructura; 2) fotointerpretación, geoprocésamiento de imágenes y mapeo geológico-estructural de las características estructurales y litológicas de las fallas y zonas de cizalla; 3) análisis petrográfico de muestras orientadas en campo; 4) cuantificación de la orientación y geometría de deformación mediante análisis de deformación finita 3D y cuantificación de deformación no-coaxial por medio de análisis de vorticidad; 5) análisis submicroscópico SEM-TEM-EBSD; 6) cuantificación de condiciones de deformación P-T por medio de modelamiento de equilibrio de fases o geotermobarometría convencional; 7) datación de minerales sincinemáticos y rocas miloníticas por el método Ar-Ar, con el fin de determinar las edades de reactivación y deformación de la estructura, respectivamente, así como, la implementación de la técnica U-Pb en cristales sincinemáticos de calcita desarrollados en los planos de falla y en cristales de apatito con la finalidad de entender la evolución temporal de las zonas de cizalla; y finalmente, 8) fechamiento de elementos geológicos adyacentes a la estructura, tales como: intrusivos sincinemáticos asociados al evento deformacional por el método U-Pb en circón, rocas alteradas hidrotermalmente mediante el método Ar-Ar, y datación por trazas de fisión en circón y apatito de los bloques adyacentes a las fallas para determinar edades de exhumación.

**Palabras clave:** Zonas de cizalla, rocas de falla, régimen frágil, transición frágil-dúctil, régimen dúctil, mapeo, geocronología.

## 1. INTRODUCTION

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The Geological-structural data acquisition for mapping of faults and shear zones is important as basic technical input in the formulation of engineering projects and territorial and national development plans, for the design and layout of road infrastructure with large civil works such as dams, viaducts and tunnels, among others, exploration of water, geothermal, mineral and hydrocarbon resources, for the reconstruction of the deformational evolution of an area, and for studies of regional tectonics (cf. Cox et al., 2001; Sausgruber and Brandner, 2001; Vega Granillo et al., 2009; Mejía et al., 2012; Jiang et al., 2020).

In this regard, it is necessary to develop a unified methodology facilitating fractal qualitative analysis of the pattern of the trace of the shear zone and segments of associated faults (cf. Turcotte, 1989, 1997; Cello, 1997; Park et al., 2010; Barão et al., 2018), from analysis on different scales by photointerpretation, mapping of outcroppings, hand samples and microscopy. This allows establishment of the structural level of the crust represented by the structures, the types of related fault rocks and their distribution, the amplitude over which the movement is distributed, geometry and its respective kinematics, the establishment of structural patterns, and the timing of the deformation.

A shear zone is defined as a high-deformation planar or curvilinear surface, along which the movement is distributed, it presents definite limits characterized by the presence of structures and fault rocks included

in centimetric to kilometric areas, according to the scales of deformation (Ramsay, 1980 a,b;Rutter, 1986; Ramsay and Huber, 1987; Jiang and White, 1995; Fossen and Cavalcante, 2017). The shear zones focus the deformation heterogeneously (both coaxial and non-coaxial), arranged as subparallel or conjugate sets (anastomosed) in response to the rheology of rocks and minerals that they affect, and usually are surrounded by lithologies showing low deformation (Ramsay, 1980 a, b;Rutter, 1986; Ramsay and Huber, 1987; Jiang and White, 1995; Fossen and Cavalcante, 2017).

Considering that the deformation is heterogeneously distributed throughout the rock body (Ramsay, 1980 a, b;Ramsay and Huber, 1987;Fossen and Tikoff, 1997; Carreras et al., 2013), it is essential that interpretation of the timing of any deformative event and the analysis of the tectonic evolution of a shear zone is based not only on evaluation of the isotopic dating of the fault rocks (cf. Schneider et al., 2013), but also on the integration of cartography, microstructural analysis, ages of intrusion igneous bodies related to the shear zone, and ages of uplift and exhumation of blocks adjacent to the structure are integrated (cf. Oriolo et al., 2018).

This review presents a methodology for the identification, mapping and analysis of the nature of faults and shear zones. For this, it is suggested to characterize the structures and fault rocks, define the structural level observed, determine the movement's distribution and kinematics, and assign the relative and absolute ages of deformation. In order to propose some minimum parameters necessary to perform geological-structural studies for acquiring basic information that aids understanding of the tectonic evolution.

## 2. PHASES FOR GEOLOGICAL-STRUCTURAL MAPPING OF FAULTS AND SHEAR ZONES

The characterization of geological-structural faults and shear zones, as first to measure, requires correct application of concepts and principles related to the study of faults and shear zones. Furthermore, as a second measure, one should be clear about the objective of the analysis, as well as the “for what?” the structural data are collected, which means being clear about what one will do with the data and how the subsequent analysis will be conducted. Finally, as a third measure, there must be broad and detailed knowledge of the methods of mapping, compilation of structural data, recognition of structures, field classification of fault rocks, collection of structural data with statistical representativeness, and determination of the distribution, kinematics (for which it is necessary to know the kinematic indicators) and geometry.

As a starting point, it is necessary to execute photointerpretation and geoprocessing analysis of satellite images with band compositions that highlight the structures. Such analysis should be complemented by geophysical inputs (e.g., maps of total magnetic intensity, total-field magnetic anomaly, analytical signal of the total-field magnetic anomaly, and complete Bouguer anomaly, among others) and an adequate map base (with drainage networks and contour lines). These allow identification and definition of the preliminary outline of the structure (cf. Gunn et al., 1997; Betts et al., 2003;Grauch and Hudson, 2007;Aitken and Betts, 2009; Stewart et al., 2009; Stewart and Betts, 2010;Kadima et al., 2011; Blaikie et al., 2014; Armit et al., 2014;Blaikie et al., 2017). This implies the identification of geomorphological features that define the fault or shear zone and sites of interest for subsequent campaigns for field checking.

For mapping structures, it is recommended to perform a multiscale analysis of nature and the spatial distribution of structural elements in the central and marginal areas of the shear zone (Figure 1), in which one will observe the fractal distributions, considering similar arrangements of geological characteristics, independent of the scale of observation (Turcotte, 1989; 1997;Cello, 1997;Park et al., 2010;Fossen, 2013; Baron et al., 2018). This allows identification of the timing of the deformation events basen mainly on cross-cutting relationships. For this, it is suggested to do transects perpendicular to the strike of the structures that limit and constitute the shear zone (cf. Chetty, 2014;Choi et al., 2016), recording morphotectonic, geometric and kinematic characteristics, and acquiring structural data on the fabric developed in the main

structure, as well as satellite faults, branches and other secondary structures (e.g., folds, tension gashes and foliations, among others), emphasizing the hierarchical structure of the structures.

Another relevant aspect in the mapping of this type of structure is a description of the associated fault rocks. These are a group of rocks developed in shear zones at different structural levels of the crust, which originate by heterogeneous deformation from cataclastic processes, intracrystalline plasticity or flow, depending on physical conditions of the deformation and the type of affected lithology (Sibson, 1977; 1983; Rutter, 1986; Ramsay and Huber, 1987). These are subdivided into non-cohesive rocks, mainly fault gouge and breccia, and cohesive rocks such as cataclasites and mylonite, which are the result of episodes of intense deformation, which reduce the grain size of the rocks, either by processes of abrasion or intracrystalline plasticity (Sibson, 1977; 1983; Killick, 2003), imprinting particular mechanical characteristics on the rocks (Sibson, 1977; Wise et al., 1984; Spray, 1995; Woodcock and Mort, 2008; Magloughlin, 2010).

The analysis of structural data acquired along fault planes and shear zones starts from the register of the orientation and direction of movement of the fault planes (cf. Petit, 1987; Hancock, 1985; Doblas, 1998). Moreover, it includes the categorization of structural data according to their quality determined in the field (supposed, probable or possible, and true) (Hardcastle, 1989; López-Isaza et al., 2021) and the cross-cutting relationship between the structures and geological units this affects (cf. Angelier, 1994; Sippel et al., 2009; Sperner et al., 1993; Tranos, 2009, 2011; Sperner and Zweigel, 2010). With the fault data, different inversion methods can be used to help determine the principal axes of deformation and stress, from kinematic and dynamic analyses, respectively (cf. Twiss and Unruh, 1998; Žalohar, 2014; Thakur et al., 2020).

The kinematic analysis involves plotting data on different diagrams, like the tangent-lineation (Twiss and Unruh, 1998) or H $\ddot{o}$ eppener (1955), in order to display fault planes as poles, including the relative movement of blocks adjacent to the fault (Sperner and Zweigel, 2010). It is also recommended to perform a kinematic compatibility analysis, which allows the establishment of compatible structures kinematically generated or activated under the same deformation event (e.g., Santamaría-Díaz et al., 2008). Furthermore, if possible, one should calculate the principal axes of elongation  $\epsilon_1$ ,  $\epsilon_2$  and  $\epsilon_3$  (Sperner and Zweigel, 2010), and it is suggested to use the Linked Bingham method of distribution, which solves the axes of shortening and extension of a set of faults (Marrett and Allmendiger, 1990).

The dynamic analysis is performed based on classification of the subsets established in the kinematic analysis. With each subset, it is recommended to test the mechanical coherence of the data from their distribution on Mohr's circle (cf. Velandia, 2017), followed by use of the inversion method of slickensides (Sperner and Ratschbacher 1994), to obtain the paleostress tensors. This method assumes that movement of the fault is controlled by a stress tensor and that the associated failures and slickensides were developed under the same deformation event. Therefore, the direction of the stress field can be determined from the preferential slips of the fault planes (Twiss and Unruh, 1998; Lacombe, 2012). This analysis is based on the following assumptions: 1) the direction of movement is parallel to the maximum shear stress resolved on the plane (Wallace, 1951; Bott, 1959); 2) the rock behaves as a rheologically linear material (Srivastava et al., 1995); 3) the displacements on the fault planes are independent and small with respect to their lengths, and in them there is no rotation of the planes, obeying the Wallace-Bott criterion; 4) the stress field is homogeneous (Twiss and Unruh, 1998). The inversion method is calculated by graphical (right dihedral method) or mathematical solutions (numerical dynamic analysis), usually included in software of structural data processing (e.g. Wintensor 5.8, TectonicsFP 1.7.7).

On the other hand, considering that the data associated with slip faults may be heterogeneous, it is proposed to apply a genetic algorithm method (MGA, for its acronym in English). This allows direct estimation the states of paleostress from this type of data and does not require the separation of subsets of data (Thakur et al., 2020).

The right dihedral method developed by Angelier and Mecheler (1977) allows one to obtain an optimum orientation of stress fields from the directions and slip sense. It is based on graphing an auxiliary plane

perpendicular to the slickensides, which divides the area around each fault plane into four quadrants that define compression and tension zones (Ortner et al., 2002). From the sum of the areas of solutions for each structural data is obtained the dihedrals resulting from distension and compression corresponding to the dataset, in which the common orientations correspond to the optimal positions of axes  $\sigma_3$  and  $\sigma_1$  in the dihedrals of extension and compression, respectively (Angelier, 1979, 1994; Sainz et al., 1990; Casas et al., 1990; Delvaux and Sperner, 2003; Delvaux, 2012; Delvaux et al., 2012).

## 2.1 Conceptualization and reference geological context

To do the geological-structural mapping it is necessary to have knowledge of the tectonic and geological context in which the structure developed. Moreover, one should previously consider the concept of shear zones, what characterizes them, their architecture, what their regimes are, and how they relate to depth. This is in order to have a theoretical approach to the depth position in a crustal cross-section or, in other words, the observed structural level of the crust.

On the surface, shear zones are characterized by morphotectonic features (Bull, 2007) that show the landscape forms produced specifically by tectonic processes, rather than by processes of sedimentation and erosion (Keller and Pinter, 2002; McCalpin, 2009; Burbank and Anderson, 2012). These types of features have been extensively documented in neotectonics studies. Among these features can be found mountain fronts with triangular facet, fault escarp, flexural escarp, sag pond, fault saddle, positive (pressure ridge) and negative (pull-apart basin) flower structures, shoulders, shutter ridge, and forebergs, in addition to the development of interference drains, which are very sensitive to changes in the surface, among others.

The identification of morphological and structural features associated with the development of shear zones have been widely characterized by aerial photograph analysis, satellite and radar images, during phases prior to field work, in order to identify the location and distribution of the landscape features, as well as the elevation of the terrain and its composition (Smith and Pain, 2009). These are characteristics that may not be directly recognizable in the field because of their scale.

In this sense, the photointerpretation is a fundamental part of any geological study that involves the recognition of morphological and structural features. Traditionally, aerial photographs have been used, which facilitate a perception of the terrain in 3D through stereoscopy, with vertical exaggeration that makes it possible to estimate the slopes, stratification, and terrain configuration. However, given its optical character, its usefulness is restricted by cloudiness and thick vegetation cover, which prevent the assessment of the land (Mendivelso, 2008; Rampal, 1999).

During recent decades, great progress has been made in satellite and remote sensing technologies (Rao, 2002). The variety of available satellite images, with differing spatial, radiometric and temporal resolutions, allows their use in recognizing the composition of the earth's surface, its cover and temporal evolution. One can distinguish two major categories of satellite images: those from passive sensors (e.g., Landsat, Sentinel 1, Aster, SPOT, and others), and those from active sensors, commonly known as radar images (e.g., Sentinel 2, ALOS-PALSAR, GeoSAR, RADARSAT-1, and others). The latter have the advantage of allowing observation of the earth surface even in overcast conditions and at night. Additionally, some of them in the L band go beyond the vegetation cover and penetrate layers of soil, snow or ice (Paine and Kiser, 2003). They are therefore the most useful in geological and geomorphological studies.

Lidar (acronym for light detection and ranging) is another technology of the active type, like radar, which facilitates topographic data of high accuracy and density, so it supports the construction of digital elevation models with high spatial resolution. This technology can be implemented using unmanned aerial vehicles (UAV- drones), which enables the recognition of vast areas. Compared with traditional photogrammetry, Lidar provides data on terrain covered by vegetation (Paine and Kiser, 2003), as shown by the detection and characterization of active faults in wooded areas (e.g., Chen et al., 2015).



To facilitate the detection of fault zones and associated geomorphology features, various algorithms and processing techniques have been created, such as the application of spatial filters that highlight and automatically extract information from satellite imagery (e.g., Gannouni and Gabtni, 2015; Mallast et al., 2011). However, despite the quality and utility of remote sensing tools, field testing is always necessary (Marchionni and Cavayas, 2014) to fully recognize and characterize faults and shear zones.

The architecture, both longitudinal and transverse to the trace of the shear zones, is distributed in three main elements (Caine et al., 1996; Ganerød et al., 2008): core, transition and damage zones (Figure 1). The core corresponds to a zone parallel or semiparallel to the direction of the main fault and is a product of high strain and focused shear, accommodating displacement through one or more surfaces. It is made up of shear zones, sets of conjugate fractures with different orientations, geometry, and morphology (Caine et al., 1991; Caine et al., 1996; Gabrielsen and Braathen, 2014). It also includes lenses or areas of sheared rocks (e.g., foliate cataclasites and mylonites), locally fractured or crushed (e.g., ultracataclasites, breccia and fault gouge), with evidence of processes of precipitation and geochemical alteration (Sibson, 1977; Anderson et al., 1993; Chester and Logan, 1986; Childs et al., 1996; Wibberley et al., 2008; Gabrielsen et al., 2008; Bastesen and Braathen, 2010). Its width can vary from a few centimeters to hundreds of meters (Gabrielsen and Braathen, 2014).

The transitional zone is mainly composed of elongated fragments of protocataclasites and/or protobreccia (Lindanger et al., 2007; Gabrielsen and Braathen, 2014). It also includes fracture corridors, shear and gouge zones of minor faults, arranged in a subangular manner with respect to the general trend of the main fault plane, cross-cutting locally previously formed structures (Gabrielsen and Braathen, 2014). This zone also has of greater cohesion and is generally less affected by processes of hydrothermal alteration, as compared to the fault core (Berg and Skar, 2005; Faulkner et al., 2010; Gabrielsen and Braathen, 2014; Choi et al., 2016).

The damage zone, which surrounds the previous zones, corresponds to the volume of rock that is around the fault core and transitional zone, and preserves the original lithological properties of the deformed rocks (Chester et al., 1993; McGrath and Davinson, 1995; Beach et al., 1999; Storti et al., 2003; Billi et al., 2003; Chester et al., 2004; Berg and Skar, 2005; Johansen and Fossen, 2008; Mitchell and Faulkner, 2009; Gudmundsson et al., 2010; Riley et al., 2010; Hausegger and Kurz, 2013; Lin and Yamashita, 2013; Choi et al., 2016). The damage zone is characterized by thin bodies with rhombuses-shapes, with low deformation intensity (Gabrielsen and Braathen, 2014), as well as by structures of second order, synthetic and antithetic fractures, joints, veins and folds. These are related to the kinematics of the structure, being less representative in relation to increasing distance to the fault core (Chester and Logan, 1986; Smith et al., 1990; Berg and Skar, 2005; Faulkner et al., 2010; Gabrielsen and Braathen, 2014).

Inside the crust, the shear zones present structural levels characterized by the deformation regimes and structures produced in response to the distribution of the deformation, called brittle, brittle-ductile and ductile (Ramsay, 1980 a,b; Rutter, 1986; Ramsay and Huber, 1987; Fossen and Cavalcante, 2017). In conditions of average geothermal gradient (between 25 °C and 30 °C per kilometer), these structural levels are demarcated by limits depending on the rheological characteristics of quartz and the feldspar (Figure 1), which respond to fracturing deformation by abrasive processes (cataclastic) or flowing by accommodation processes (frictional sliding), respectively (Hills, 1972; Sibson, 1979; 1983; Knipe, 1989; Hatcher, 1995; Snoke et al., 1998; Blenkinsop, 2002).

In this regard, the brittle or elastic-frictional regime (Sibson, 1977; 1983; Ramsay, 1980 a, b; Rutter, 1986; Ramsay and Huber, 1987; Hanmer, 1988; Fossen and Cavalcante, 2017) is characterized by fault zones (Sibson, 1979; Chester et al., 1993; Chester and Chester, 1998; Kim et al., 2003; Chester et al., 2004; López et al., 2008; Hausegger and Kurz, 2013; Choi et al., 2016; Ostermeijer et al., 2020) and include interconnected and closely spaced subparallel structures that from fracturing, abrasion, grain-boundaries Sliding, pressure-dissolution and grain size decrease processes favor the development of breccia or gouge fault zones between them (Higgins, 1971; McClay, 1987; Davis and Reynolds, 1996). The fault zones include domains with

different deformation intensity, distributed in the fault core and damage zone (Figure 1), characterized by structural elements defined by the variation in fracturing intensity. Furthermore, the development of fault rocks (fault gouge, breccia and cataclasites) and kinematic indicators (major fault -Y, -P shear, riedel shear R, conjugate riedel shear R', tensile fracture -T, crescentic marks, slickenside, congruous and incongruous steps, trailed material, asymmetric protruding grains, carrot-shape, trains of inclined planar structures, asymmetric elevations, mineralogical/crystallographic orientation, among others.) allows definition of the kinematics and type of structure, plus the associated stress field (cf. Hancock, 1985; Petit, 1987; C  lerier, 1988; Doblas et al., 1997; Doblas, 1998; Fry, 2001).

Under conditions equivalent to the brittle-ductile regime or elastic-frictional/quasi-plastic transition, in a normal geothermal

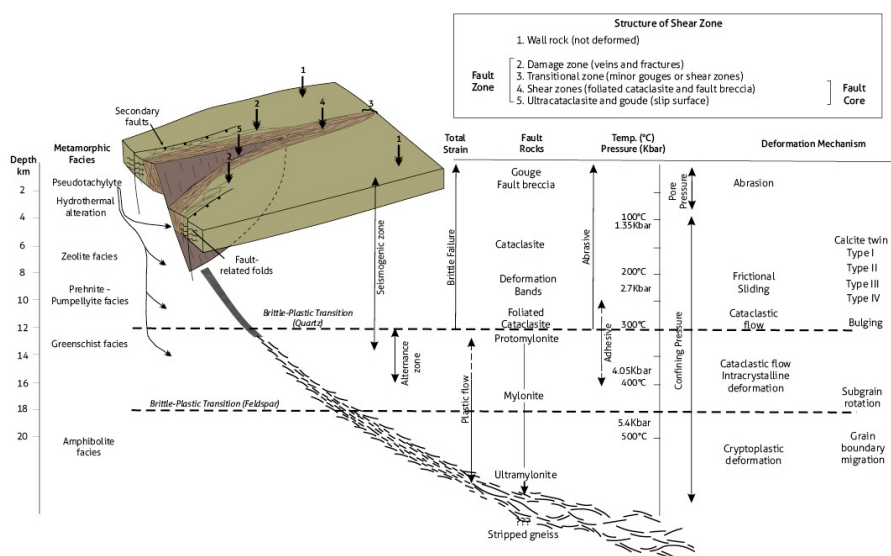


FIGURE 1

3D architecture of shear zone, which includes transition from brittle, brittle-ductile and ductile deformation to depth, and their relationship with deformation mechanisms and associated fault rocks

Source: Constructed using information from Sibson (1979; 1983); Wise et al. (1984); Scholz (1988, 2006); Swanson (1992); Passchier and Trouw (2005); Chester et al. (2004); L  pez et al. (2008), Fossen (2013), and Choi et al. (2016).

gradient (Sibson, 1977; 1983; Ramsay, 1980 a, b; Rutter, 1986; Ramsay and Huber, 1987; Hanmer, 1988; Scholz, 1988; Fossen and Cavalcante, 2017), rocks undergo a transition that is registered in the behavior of quartz, which goes from a brittle fracture mode to plastic flow (Rutter, 1986; Duba, 1990). Shear zones developed under this regime show a complex behavior related to processes of granular flow, cataclastic mechanisms and intracrystalline plastic deformations, which result in heterogeneous deformation of the rock with partition in stronger and weaker domains. Under these conditions, the behavior depends on temperature, pressure, rheology, previous anisotropies, and angular and linear geometric relationships in which the structure was generated (Sibson, 1977; Blenkinsop and Rutter, 1986; Scholz, 1988; Schmid and Handy, 1991; Babaie et al., 1991; Hadizadeh and Tullis, 1992; Passchier and Trouw, 2005; Balsamo et al., 2010; Fossen, 2013; Moreira and Dias, 2018; Kjenes, 2018; Nicchio et al., 2018; Bar  o et al., 2020). Moreover, it includes the development of fault rocks such as cataclasites S-C, phyllonites and protomylonites (Chester et al., 1985; Lin, 1999a; Passchier and Trouw, 2005; Passchier and Coelho, 2006; Trouw et al., 2010; Torgersen and Viola, 2014).

Another term commonly used for this deformation region is semi-brittle shear zones. These are areas of brittle shear influenced by mechanisms of plastic deformation, a situation in which the elements in the structure are intrinsically brittle but mechanically resistant to stop the propagation of fractures (Gumbsch,

2001;Fossen, 2013). The fault rocks in semi-brittle shear zones show mineral fragments embedded in a fine matrix of crushed and fractured minerals, veins with mineral precipitation, and evidence of cataclastic flow product of hydrothermal processes, tension gashes, intergranular and transgranular microfractures domino type, and shear bands, the latter contribute to the development of intracrystalline plastic processes (Hodgson, 1989; McClay, 1991;Hadizadeh and Tullis, 1992;Babaie and La Tour, 1993;Gumbsch, 2001;Pec et al., 2012; Prigent et al., 2017;Papeschi et al., 2018;Reber and Pec, 2018;Barão et al., 2020).

In the ductile or quasi-plastic regime (Figure 1) with a normal geothermal gradient (Sibson, 1977; 1983;Ramsay, 1980 a, b;Rutter, 1986; Ramsay and Huber, 1987;Hanmer, 1988;Fossen and Cavalcante, 2017), shear zones are dominated by the mechanical response of feldspar and are generated at deeper levels in the crust, where metamorphic conditions prevail (Sibson, 1983;Riedmüller et al., 2001; Lopez et al., 2008). In these zones, the processes of plastic flow are distributed homogeneously and involving reduction of grain size. These zones are also characterized by the development of protomylonites, mylonites, ultramylonites, blastomylonites, ribbon mylonites, stripped gneiss, and mylonite gneiss (Sibson, 1980;White et al., 1980;Snoko et al., 1998;Killick, 2003;Passchier and Trouw, 2005;Trouw et al., 2010;Fossen, 2013).

### 2.1.1 Fault rocks

Initially, fault rocks were classified according to the level of primary cohesion and planar fabrics developed (Higgins, 1971;Hills, 1972; Sibson, 1977;Marshak and Mitra, 1988). However, this classification does not consider that under certain conditions cataclastic rocks can present foliation and that the deformation mechanisms that affect fault rocks can be different (cf. Bell and Etheridge, 1973;Ramsay and Huber, 1987;Lin, 1999a).

For the description of fault rocks, it is recommended to consider guidelines of the Subcomission on the Systematics of Metamorphic Rock (SCMR) of the International Union of Geological Sciences – IUGS (Brodie et al., 2007), supplemented by recent reviews that describe the characteristics of fault rocks and classify some of their types (cf. Woodcock and Mort, 2008;Magloughlin, 2010; Fossen and Cavalcante, 2017). The following describes fault rocks according to the structural level they generate (cf. Figure 1), starting with the brittle regime and ending with the ductile regime. It should be noted that the type of fault rock generated inside the crust also depends on the prevailing geothermal gradient at the time of their formation, which is directly related to depth.

### 2.1.2 Fault breccia

This is a product of the brittle shear zones of the upper crust, and they are developed from the surface to a few kilometers deep in the crust (~0-6 km; cf. Sibson, 1977) and vary between non-cohesive and cohesive, foliated and non-foliated (Figure 2a)(Higgins, 1971;Sibson, 1977, 1986; Woodcock and Mort, 2008). Based on their cohesion, matrix or cement concentration, and clast size (> 2 mm), they are classified as: a) crackle breccia (75% to 100% of large clasts), with clasts with little rotation, separated by thin cement or matrix sutures; b) mosaic breccia (60% to 75% of large clasts) with adjacent clasts that fit together, but with more separation and rotation; c) chaotic breccia (30% to 60% of large clasts), with strongly rotated clasts and loss of geometric fit to each other (Woodcock and Mort, 2008;Magloughlin, 2010).

### 2.1.3 Fault gouge

This is a non-cohesive rock with less than 30% large clasts (> 2 mm), formed in the first kilometers of the upper crust (~0-4 km; cf. Sibson, 1977) from cataclastic processes dominated by fracturing and rotation



of the rigid body of grains and fragments. It is restricted to shear fractures and slip planes in fault zones (Figure 2b) and it is composed by a material rich in clay, which can present foliation and plastic response in the presence of moisture (Higgins, 1971; Engelder, 1974; Wu, 1978; Chester et al., 1985; Scholz, 1987; 1988; Snoke et al., 1998; Killick, 2003; Vannucchi et al., 2003; Woodcock and Mort, 2008; Magloughlin, 2010; Haines et al., 2013; Vrolijk et al., 2018).

#### 2.1.4 Cataclasite

This denomination classifies a cohesive rock with or without foliation (figure 2c), developed in the upper crust at depths shallower than those of the ductile-brittle transition (~4-10 km) (Figure 1). It is characterized by angular porphyroclasts in a fine-grained matrix of similar composition, which are deformed by mechanisms associated with processes of cataclastic and granular flow, fracturing, rotation and frictional sliding of particles (Higgins, 1971; Sibson, 1977; 1979; Scholz, 1988; Blenkinsop and Rutter, 1986; Babaie et al., 1991; Hadizadeh and Tullis, 1992; Snoke et al., 1998; Woodcock and Mort, 2008; Magloughlin, 2010; Balsamo et al., 2010; Moreira and Dias, 2018; Kjenes, 2018; Nicchio et al., 2018; Barão et al., 2020). According to the degree of proportion of the matrix (cf. Sibson, 1977; 1979; Snoke et al., 1998; Woodcock and Mort, 2008; Magloughlin, 2010), it is classified as: a) protoclasite, in the which the matrix makes up less than 50% of the rock volume; b) mesocataclasite, for which the matrix comprises more than 50% and less than 90 % of the rock volume; c) ultracataclasite, in which the rock volume consists of more than 90% matrix and is restricted to fault cores where the shear concentrates (Figure 1); d) S-C cataclasites, which are characterized by a shape orientation of some mineral phases (S planes) and the generation of micro-shear planes or shear bands (C planes), with quartz and feldspar crystals developing brittle deformative microstructures with no dynamically recrystallized grains in conditions of temperatures between 150 °C and 250 °C (Chester et al., 1985; Lin, 1999a).

#### 2.1.5 Phyllonite

Cohesive rock with phyllitic appearance, formed by processes of aseismic creep (drag) (Higgins, 1971; Sibson, 1977; Passchier and Trouw, 2005; Trouw et al., 2010; Torgersen and Viola, 2014) or by syndeformational fluid circulation and flow during its generation (cf. Spruzeniece and Piazzolo, 2015). It may be related to cataclasites and mylonites and develops transitional contacts with protomylonite in some shear zones, locally with complete lateral variation of the structure, or hosting quartz tension cracks (cf. Hippertt and Massucatto, 1998).

#### 2.1.6 Mylonite

This designation includes cohesive rocks with mineral-stretching lineations (Figure 2d), strongly deformed in a ductile shear zone, surrounded by less deformed rocks, developed at depths equivalent to the middle-lower crust (Figure 1), under a normal geothermal gradient. It is characterized by two types of well-defined foliation, called S-C structures (Figure 2e), which result from a reduction in grain size from plastic processes. They are also constituted by lithic fragments or porphyroclasts of type  $\sigma$ ,  $\delta$ , fish or sigmoidal, in addition to porphyroclastic systems with the development of mantle-core structures, quarter structures, reaction rings, and deformation shadows that, generally, have a composition similar to that of a fine-grained matrix (Bell and Etheridge, 1973; Lister and Snoke, 1984; Hooper and Hatcher, 1988; Hanmer, 1989; Ten Grotenhuis et al., 2003; Passchier and Trouw, 2005; Brodie et al., 2007; Trouw et al., 2010; Barão et al., 2020). Classification is based on the proportion of its original grains (size) and the recrystallized matrix (Sibson, 1977, 1979; Scholz,

1988; Snoke et al., 1998; Brodie et al., 2007; Woodcock and Mort, 2008; Magloughlin, 2010; Fossen and Cavalcante, 2017), and includes: a) protomylonite, in which less than 50% of the rock has been subjected to grain-size reduction processes and has greater development on C planes; b) (meso)mylonite, in which more than 50% and less than 90 % of the rock shows grain-size reduction processes; c) ultramylonite (Figure 2f), in which more than 90 % of the rock shows grain-size reduction; d) blastomylonite, corresponding to rocks that once deformation has ceased, increase the grain size due to processes associated with static recrystallization. However, this term is sometimes used to describe mylonites with a coarse-grained recrystallized matrix (Passchier and Trouw, 2005).

Besides the above descriptions of mylonites, other terms are also used, such as augen mylonite, used to describe rocks generated from a combination of cataclastic and crystalloblastic processes that show crystals or lithic fragments, generally of lenticular shape (larger than 0.5 mm), embedded in a



FIGURE 2.

General appearance of a fault rock outcrop

- a) Cohesive and non-foliated fault breccia; b) fault plane with fault gouge; c) foliated cataclasis; d) Mylonite with mineral orientation defined by the presence of milky quartz porphyroclasts; e) S-C structure characteristic of mylonites; f) ultramylonite with porphyroclasts of mafic minerals and oxides; g) pseudotachylite injection veins.

fine-grained matrix (recrystallized or neoformed), which shows deformation in solid state (Passchier and Trouw, 2005; Mukherjee, 2014). These rocks generally form symmetric or asymmetric anastomosed

planar structures, consisting mainly of felsic minerals (Higgins, 1971; Brodie et al., 2007). The term ribbon mylonite is also used to describe strongly foliated rocks, mainly constituted by parallel monomineralic lenses (Passchier and Trouw, 2005; Trouw et al., 2010), common in high-grade shear zones that are transitional deformed to stripped gneisses. The stripped gneisses correspond to rocks interpreted as mylonites with gneissic structure defined by planar compositional, formed under high-grade metamorphism (Hippertt, Rocha et al., 2001; Passchier and Trouw, 2005; Trouw et al., 2010).

### 2.1.7 Pseudotachylites

These originate from (paleo)seismic movements producing melting during rapid frictional sliding and are mainly characterized by evidence of a precursor melt phase (Magloughlin, 2010; Kirkpatrick and Rowe, 2013). The pseudotachylites occur as phases included within veinlets and veins with thicknesses from millimetric to centimetric (Figure 2g). They may consist of glass, devitrified glass, oxides, carbonates, chlorite and even epidote, as veins of the main fault plane, generation veins or injection veins, with sharp contacts, locally corroded, and cooling margins (cf. Lin, 2008; López et al., 2008; Magloughlin, 2010; Kirkpatrick and Rowe, 2013; Altenberger and López, 2014). They are constituted by fragments of broken crystals from host rock embedded in a matrix ranging from

TABLE 1.  
Recommended synoptic table for the textural classification of fault rocks

Pronounced grain growth	Incohesive	Non-foliated <span style="float: right;">→ Foliated</span>		Fault breccia Percentage of large clasts >2 mm		crackle breccia 75 - 100%		Foliated fault breccia	
		Nature of the matrix and deformation				mosaic breccia 60 - 75%			
Tectonic reduction in the size of grain and grain growth by recrystallization and neomineralization				chaotic breccia 30 - 60%					
		Cataclasite series Fracturing prevails, clasts <0.1 mm		Fault gouge Visible fragments < 30%		Foliated fault gouge Visible fragments < 30%			
Pseudotachylite Present glass injections in its sidewalls								Phyllite appearance, mylonite rich in fine-grained micas	
		Glass or devitrified glass							
Nature of the matrix and deformation				Protocataclasite / Foliated protocataclasite		Mesomylonite			
		Tectonic reduction in the size of grain and grain growth by recrystallization and neomineralization				Mesocataclasite / Foliated mesocataclasite		50%	
Cataclasite series Fracturing prevails, clasts <0.1 mm				Ultrapseudotachylite / Foliated ultrapseudotachylite				90%	
		Glass or devitrified glass				Foliated Pseudotachylite		Ultramylonite	
Nature of the matrix and deformation				Blastomylonite				Lenticular eye-shaped mineral in planar fabric	
		Tectonic reduction in the size of grain and grain growth by recrystallization and neomineralization				Lenses monophasic parallel Rock strongly shear			
Cataclasite series Fracturing prevails, clasts <0.1 mm				Stripped gneiss					
		Glass or devitrified glass							

Source: Prepared with information from Sibson (1977;1979); Scholz (1988), Snoke et al. (1998), Killick (2003), Woodcock and Mort (2008), and Magloughlin (2010).



cryptocrystalline to aphanitic, which develop structures due to flow, compositional banding, and spherulites caused by rapid cooling and devitrification (Philpotts, 1964; Sibson, 1975; Snoke et al., 1998; Lin, 1999b; 2008; Wenk et al., 2000; Winter, 2001; Passchier and Trouw, 2005; López et al., 2008; Magloughlin, 2010; Fossen, 2013; Kirkpatrick and Rowe, 2013; Altenberger and López, 2014).

### 2.1.8 Classification of fault rocks in the field

For fault rock classification in the field, it is suggested to use a tabular summary (Table 1) that synthesizes the previously described features, which includes the minimum classification parameters based on descriptions of the lithologic, physical, mineralogical, structural and textural characteristics of the rocks to be classified.

## 2.2 PETROGRAPHY AND MICROTTECTONICS

Before performing the petrographic and microtectonic analyses, it is suggested to have control over the orientation of the sample collected in the field (Figure 3), indicating in the sample basic information related to the attitude of the plane, which should be orientated by specifying the strike of the reference surface according to the right-hand rule. It is also necessary to mark geographic North, the base (lower plane) and the top (upper plane), and the free faces of the sample (cf. Turner and Weiss, 1963; McClay, 1987; Passchier et al., 1990; Hopgood, 1999; Passchier and Trouw, 2005).

In the preparation of thin sections, one should consider that the tectonically deformed rocks can develop planar and linear fabrics, which are associated geometrically with the deformation ellipsoid, in which the greater mineral lineation is

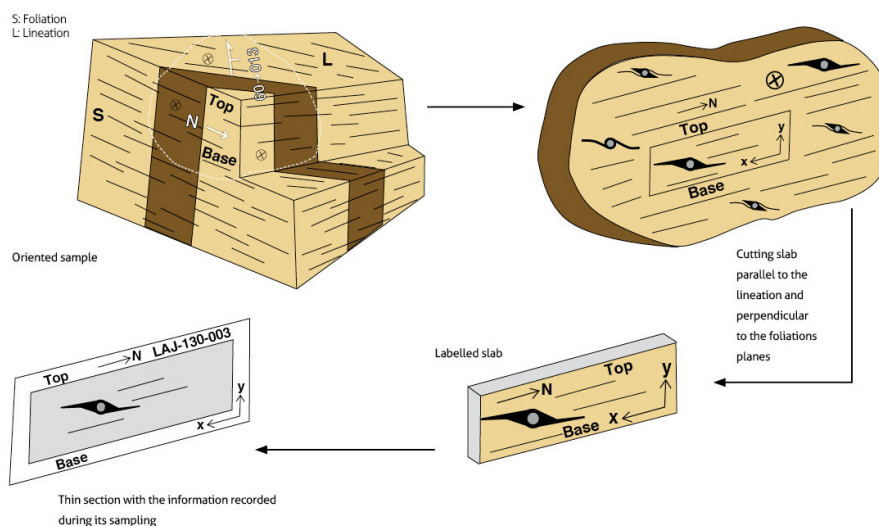


FIGURE 3.

#### Schematic illustrating the orientation of the sample in the field

The direction of geographic North ( $\rightarrow N$ ), structural data, free face, and top and base are indicated. The cut oriented to obtain the labelled slab must be parallel to the mineral lineation and perpendicular to the foliation planes.

Finally, on the thin section for petrographic and microtectonic analysis, the information recorded in the field is indicated as a reference for interpretation of the kinematics in relation to the sample orientation.

Source: Prepared with information from Passchier and Trouw (2005), and García (2011).

generated along the X direction and the foliation correspond to the XY planes of the ellipsoid (Flinn, 1979; Ramsay and Huber, 1983; Passchier and Trouw, 2005; Fossen, 2013). To make a cut that reveals the greatest deformation, and additionally, the geometry and direction of movement registered by the kinematic



indicators is observed, a section normal to the axis of deformation symmetry must be defined (XZ section of the deformation ellipsoid). This corresponds to the direction perpendicular to the foliation and parallel to the mineral lineation or direction of stretching of the linear fabric elements of interest (Figure 3) (Lister and Hobbs, 1980; Schmid and Casey, 1986; Law, 1990; Hippertt, Lana et al., 2001; Passchier and Trouw, 2005; Trouw et al., 2010; Rutter et al., 2011; Parsons et al., 2016; Goswami et al., 2018). A thin section normal to the mineral direction and parallel to the planes of foliation do not reflect the geometry of the kinematic indicators, because in this cut will be observed  $\Phi$  and  $\theta$ -type porphyroclasts and porphyroblasts. These do not exhibit deformation shadows, and therefore they do not allow the establishment of a kinematic sense (Passchier and Trouw, 2005; Trouw et al., 2010; Davis et al., 2011). It is important also that the thin section replicates the information registered for the rock during sampling, i.e., which show the coordinate axes, geographic north and indications of top and base, among other aspects.

According to the mechanical response of the minerals, the conditions of pressure, temperature, differential stresses and deformation rate under the deformation occurs it is possible microscopically observe the following deformation mechanisms (Blenkinsop, 2002; Passchier and Trouw, 2005).

1. Brittle fracturing – a cataclastic flow, in which fragmentation, rotation and frictional sliding occurs between grain boundaries through inter and intragranular or transgranular microfractures, whether those microfractures are between the grains, affect one grain, or pass through several grains, respectively (Passchier and Trouw, 2005). This mechanism occurs mainly in brittle conditions of low temperature ( $\sim 150$  through  $\sim 400$  °C) (Blenkinsop and Rutter, 1986; Hadizadeh and Tullis, 1992; Picazo et al., 2013; Baron et al., 2020) and high deformation rates.

2. Intragranular deformation, assisted by dissolution-precipitation processes, responsible for accommodating deformation along the contact surface between the grains or minerals, under semi-brittle regime and brittle-ductile transition conditions.

3. Intracrystalline deformation, which orders point defects and linear distortions along the lattice planes of the crystal, in conditions of brittle-ductile transition, developing deformation lamellae, twinning, undulatory extinction, and microkinks.

4. Processes of recovery and recrystallization, mechanism generating displacement and organization of dislocations, to be accumulated in the form of planes subsequently generate subgrain limits along the edges of larger crystals. One can identify three deformation mechanisms in dynamic recrystallization: bulging (BLG) (Figure 4a), subgrain rotation (SGR) (Figure 4b) and grain boundary migration (GBM) (Figure 4c), developed in ranges of deformation temperatures  $\sim 300$ - $410$  °C,  $410$ - $520$  °C and  $> 520$  °C, respectively (Zulauf, 2001; Stipp et al. 2002; Faleiros et al., 2010). At low temperatures ( $< 400$  °C), the calcite is deformed by crystalloplastic processes along the twinning planes, called “e planes” (Despaigne and Cáceres, 2009). According to the geometry in which the twinning planes develop, it is possible to estimate the deformation temperature conditions of the calcite crystals. These are classified as: a) type I, developed at temperatures of  $170$  °C (Burkhard, 1993; Passchier and Trouw, 2005; Despaigne and Cáceres, 2009); b) type II, which occur at  $150$ - $300$  °C (Ferrill, 1991; Ferrill et al., 2004; Despaigne and Cáceres, 2009); c) type III, which develop at temperatures  $> 200$  °C (Burkhard, 1993; Despaigne and Cáceres, 2009), d) type IV, developed at temperatures  $> 250$  °C (Burkhard, 1993; Ferrill et al., 2004; Despaigne and Cáceres, 2009).

5. Solid-state diffusion creep, which occurs mainly at high temperatures, and is related to the transfer of atoms along grain boundaries. It generates curved contacts and chemical zonations.

### 2.3 3D FINITE DEFORMATION AND KINEMATIC VORTICITY ANALYSIS

The 3D finite deformation analysis is based on the calculation of the deformation ellipsoid, which characterizes the planar and linear fabrics of the deformed rocks (corresponding to the XY planes and X axis of the finite deformation ellipsoid, respectively) (Passchier and Trouw, 2005) and quantifies its scalar

parameters of magnitude and shape (T parameter, Jelinek, 1981; Flinn parameter or K value, Ramsay and Huber, 1983; degree of anisotropy, Ramsay and Huber, 1983; Borradaile and Werner, 1994; Lagroix and Borradaile, 2000; Nakamura

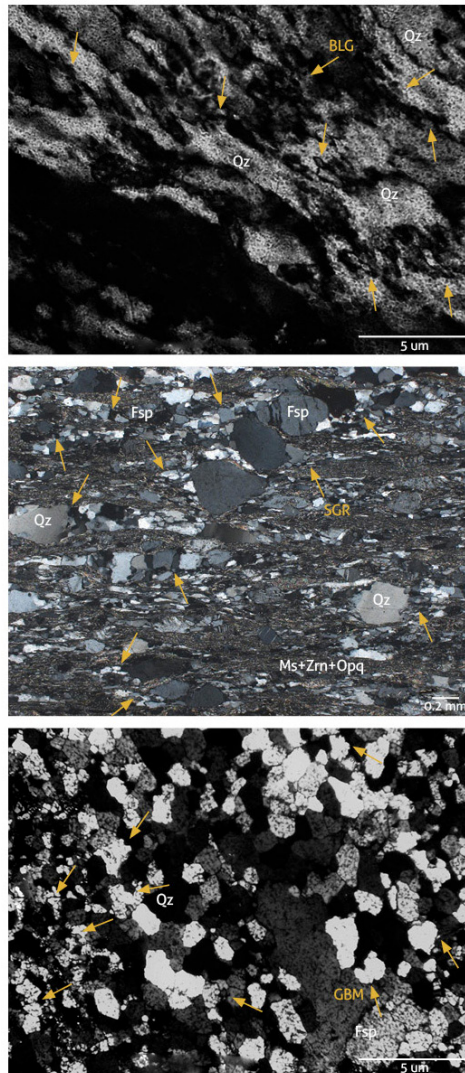


FIGURE 4.

Microphotographs of samples showing intercrystalline plasticity processes evidenced by deformation mechanisms

a) Bulging; b) Subgrain rotation; c) Grain boundary migration. Yellow arrows indicate sites where one observes the dynamic recrystallization previously described (see text).

and Borradaile, 2004). This analysis usually implements the inertial tensor (cf. Launeau and Cruden, 1998) and quadratic tensor (cf. Robin, 2002; Launeau and Robin, 2005) methods. It assumes that the deformation was homogeneous and that the crystals initially had equidimensional shapes. Therefore, the quantification of the geometric properties of the deformation ellipsoid and orientations of its main axes reflect the planar and linear structures developed in the rock body (Launeau and Robin, 2005; Passchier and Trouw, 2005). From the shape geometry and the orientation of the ellipsoid obtained, it is also possible to determine the nature of the deformation (coaxial vs. non-coaxial shear, and constrictional vs. flattening deformation) and the stress regime under which the shear zone developed (transtension, transpression, shearing, tension and compression) (cf. Sanderson and Marchini, 1984; Fossen and Cavalcante, 2017; Riberio et al., 2019; Forero-Ortega et al., 2020).

The vorticity analysis determines and quantifies the kinematics of the flow in the shear zones (Xypolias, 2010). It focuses on the interpretation of structural fabrics in terms of the degree of noncoaxiality, i.e., evaluating the relationship between rotation and flow stretching components via numerical modeling (cf. McKenzie, 1979; Means et al., 1980; Lister and Williams, 1983), which quantifies the values of kinematic vorticity ( $W_k$ ) (Truesdell, 1953), sectional kinematic vorticity ( $W_n$ ) (Passchier, 1997), and mean kinematic vorticity ( $W_m$ ). This analysis implements 2D methods on surfaces parallel to the XZ plane of the finite deformation ellipsoid, in order to estimate the spatial contribution of the simple and pure shear components in deformation zones in terms of  $W_m$  or  $W_n$  values, such as the distribution of linear materials (dikes or sets of deformed veins) (cf. Wallis, 1992; Kumerics et al., 2005; Short and Johnson, 2006), rotation of rigid objects (cf. Johnson et al., 2009; Langille et al., 2010; Thigpen et al., 2010; Forero-Ortega et al., 2020), inclusion patterns in porphyroblasts (cf. Beam and Fisher, 1999; Iacopini et al., 2007), RXZ/ $\beta$  method (cf. Sarkarinejad et al., 2010; Law, 2010; Xypolias et al., 2010),  $\delta/\beta$  method (cf. Xypolias, 2009; Ribeiro et al., 2019) and others (cf. Xypolias, 2010).

## 2.4 SUBMICROSCOPIC SEM-TEM-EBSD ANALYSIS

The Scanning electron microscopy (SEM) and transmission electron microscopy (TEM) analyses with electron backscattering diffraction (EBSD) are performed to evaluate textures and compositions (number of phases), based on quantitative microstructural analysis of crystallographic disorientation angles (characteristics of the boundary mineral) and the average shape and size distribution. These analyses also characterize crystallographic textures, intracrystalline deformation, dynamic recrystallization mechanism, and preferential crystallographic orientation (cf. Behrmann and Mainprice, 1987; Prior et al., 1999; Maitland and Sitzman, 2007; Mariani et al., 2008; Chen and Thomson, 2010; Kilian et al., 2011; Oliot et al., 2014; Platt, 2015; Desbois et al., 2016; Vrolijk et al., 2018; Cavalcante et al., 2018; Goswami et al., 2018; Scheiber et al., 2019; Bella et al., 2019; Ribeiro et al., 2019; Ault, 2020; Conte et al., 2020).

These analyses are carried out on fine-grained minerals in mylonites (mainly quartz and feldspar crystals), which define the transitions of structural levels and rheological conditions of the crust, as well as the location of deformation and development of the shear zone, depending on the prevailing geothermal gradient. These analyses are also performed on clay minerals in fault gouge and iron oxide precipitates developed along the sliding planes, for which it is not possible to easily observe the deformation mechanisms generated, the geometry, and the direction of movement registered by the kinematic indicators.

## 2.5 GEOTHERMOBAROMETRY AND THERMODYNAMIC MODELING

Geothermobarometry has wide applications in the study of pressure and temperature conditions of the various phases and associations of mineral phases. These include establishing the emplacement depth of plutonic bodies (cf. Hammarstrom and Zen, 1986; Bohlen and Lindsley, 1987; Schmidt, 1992; Moazzen and Drop, 2005; Bernet et al., 2019; Cetina et al., 2020), determination of the metamorphic peak of metamorphic rocks, and definition of pressure-temperature-time path (cf. Bohlen and Lindsley, 1987; Essene, 1989; Spear, 1993). The objective of geothermobarometry is to infer the pressure and temperature conditions under which a sample has reached equilibrium (cf. Spear, 1993). The description of geothermometers and geobarometers, as well as the analytical methods used, is beyond the scope of this review, therefore, the interested reader can consult Essene (1989), Spear (1993), and Powell and Holland (1994, 2008) for the basis, and Moecher and Brearley (2004), Grujic et al. (2011), Cross et al. (2015) and Cao et al. (2017), among others, for specific applications to fault rocks.

Starting from petrographic and microtectonic analyses, the aspects that must be considered when conducting geothermobarometry studies in fault rocks involve: 1) recognition of the main deformation structures and kinematic indicators; 2) determination of the paragenesis and equilibrium mineral assemblage present in the rock related to the structures and kinematic indicators; 3) definition of minerals susceptible to microanalysis, in order to obtain compositional data for geothermometry (e.g., chlorite, quartz and mullite) and geobarometry (e.g., white mica or fengite, plagioclase + garnet); 4) calculation of pressures and temperatures.

It has been considered that the deformation in shear zones may affect the rates and mechanisms of chemical equilibrium in different mineral phases (cf. Steffen and Selverstone, 2006; Richard et al., 2014). The combination of metamorphism and deformation causes transient changes in the speed of the reactions and kinetic pathways that generate mineral associations in chemical equilibrium. These processes, therefore, affect the accuracy and uncertainty of results obtained from geothermobarometers analysis, thus evidencing the presence of heterogeneities or chemical imbalances (cf. Steffen and Selverstone, 2006).

To minimize uncertainties caused by chemical imbalances, another method used to obtain pressures and temperatures is thermodynamic modeling. This method allows the prediction of mineral phases in equilibrium within certain pressure and temperature ranges, based on a given chemical composition, and aims to understand the evolution of mineral associations (Spear, 1993). The diagrams that include mineral associations distributed in stability fields are called pseudosections, and correspond to maps or sections of phase diagrams that show the mineral equilibrium that can be used to predict the pressure-temperature conditions in which the major mineral associations are generated (Tinkham, 2007; Vernon and Clarke, 2008), from specific compositions of total rock (Powell et al., 1998; Zuluaga et al., 2006; Tinkham, 2007).

Starting from petrographic analysis, the aspects that must be considered to do thermodynamic modeling of fault rocks include: 1) defining the mineral phases to be modeled; 2) establishing the chemical components of the system; 3) determining the chemical system that defines the phase equilibrium (e.g., KFMASH, which corresponds to K<sub>2</sub>O, FeO, MgO, Al<sub>2</sub>O<sub>3</sub>, SiO<sub>2</sub>, H<sub>2</sub>O); 4) finding the intensive variables or degrees of freedom of the system to be modeled by application of the phase rule; 5) running the Gibbs free energy minimization routine (e.g., Perple\_X); 6) constructing the P-T (pressure-temperature) pseudosection. Additionally, if geothermobarometry data are available, the P-T trajectories can be reconstructed.

Construction of a pseudosection is done using software, among which Thermocalc (Powell and Holland, 2001 and 2005; Zuluaga et al., 2006; Tinkham, 2007; <https://hpxeosandthermocalc.org>), Theriak Domino (De Capitani and Petrakakis, 2010; <https://titan.minpet.unibas.ch/minpet/theriak/theruser.html>), and Perple\_X (Connolly, 2005; 2009; <http://www.perplex.ethz.ch>) stand out. The software description is beyond the scope of this review, so the reader is referred to the links cited above. Pseudosections are used in the modeling of faulting controlled by fluid infiltration in mylonites at different structural levels (cf. Selverstone et al., 2012; Précigout et al., 2017), of seismically coupled and uncoupled segments in subduction interfaces (cf. Angiboust et al., 2015) and the metamorphic rocks that host the shear zones (cf. Forero-Ortega et al., 2020). The pseudosections also demarcate deformation histories evidenced in most cases by prograde and retrograde segments of P-T paths owing to shear (cf. Steffen and Selverstone, 2006).

### 3. ESTIMATION OF THE DEFORMATION AGE OF FAULTS AND SHEAR ZONES

Fault and shear zones control the distribution of deformation throughout the lithosphere and play a fundamental role in the emplacement of magmas and circulation of fluids (Ramsay, 1980b; Sibson, 1990; Brown and Solar, 1998; Micklethwaite et al., 2010; Vauchez et al., 2012; Clerc et al., 2015; Smeraglia et al., 2016; Précigout et al., 2017). Therefore, understanding of the timing of activity in the fault and shear zone is essential to comprehend the tectonic evolution in a specific geological context. For this, it is necessary to integrate the mapping and analysis of structures at macro and micro-scales, geochronological and



thermochronological data that help determine directly and indirectly the thermal history of a shear zone over time, related to processes of exhumation, uplift and sedimentation associated with evolution of the structure (Bossi and Campal, 1992; Oyhantçabal et al., 1993; Oyhantçabal et al., 2001; Zwingmann and Mancktelow, 2004; van der Pluijm et al., 2006; Löbens et al., 2011; Davids et al., 2013; Hnat and van der Pluijm, 2014; Viola et al., 2016; Oriolo et al., 2016 a,b; Ring et al., 2017; Süssenberger et al., 2017).

### 3.1 Geochronology of fault rocks

In order to delimit the deformation age in shear zones, determined in a relative way from field observations such as cross-cutting relationships between the structures and the geological units, generation of zones or surfaces erosion and discontinuities, among others, the use of geochronological and thermochronological methods is recommended for fault rocks or lithological units that constitute blocks adjacent to the shear zones (Table 2). The most relevant isotopic systems and dating methods for the study of shear zones are highlighted in certain reviews (cf. Ohtani et al., 2004; Oriolo et al., 2018; Hueck et al., 2020; and references cited therein). In brittle deformation conditions, methods such as K-Ar and Ar-Ar in neoformed phyllosilicates, such as chlorite, white mica, sericite, muscovite and illite have been implemented (cf. Crone, 1996; Chan et al., 2000; Jefferies et al., 2006; McWilliams et al., 2007; Vinasco, 2001; Löbens et al., 2011; Oyhantçabal et al., 2012; Wang et al., 2016), in fault gouge (cf. Vrolijk and van der Pluijm, 1999; van der Pluijm et al., 2001; Zwingmann and Mancktelow, 2004; Haines and van der Pluijm, 2008; Duvall et al., 2011; Fitz-Díaz and van der Pluijm, 2013; Viola et al., 2016; Scheiber et al., 2019) and in the pseudotachylite matrix (cf. Crone and Omdahl, 1987; Sherlock and Hetzel, 2001; Magloughlin et al., 2001; Streepey et al., 2002; Cosca et al., 2005; Reiners and Brandon, 2006; Whitmeyer, 2008; Cassata et al., 2009; Sherlock et al., 2009; Wolff et al., 2012; Di Vincenzo et al., 2013; Bense et al., 2014; Viola et al., 2016; Süssenberger et al., 2017; Vrolijk et al., 2018). Likewise, U-Pb and U-Th analyses in veins and fibres of calcite and opal precipitated on the fault planes have been performed (cf. Verhaert et al., 2003; Uysal et al., 2007; Watanabe et al., 2008; Nuriel et al., 2011, 2017, 2019; Roberts and Walker, 2016; Pagel et al., 2018), in order to establish the age of reactivation of the shear zones (Kelley, 2002).

TABLE 2.  
Synthesis of materials and methods proposed to establish the age under different thermal conditions present in the shear zones, and the probable interpretation of these data

Deformation conditions	Material for dating	Proposed dating methods	Probable interpretation
Brittle	Quartz grains that constitute the fault gouge	Electron spin resonance (ESR)	Reactivation age of fault zone
Brittle	Hematite, magnetite and goethite precipitate	(U-Th)/He and fission track	Reactivation age of fault zone
Brittle	Veins and fibers of calcite in fault planes	U-Pb	Reactivation age of fault zone or syntectonic precipitation
Brittle	Neoformed micas	Ar-Ar/K-Ar	Reactivation age of fault zone
Brittle/Brittle-ductile	Neoformed micas, fault gouge, and pseudotachylite matrix	Ar-Ar/K-Ar	Reactivation age of fault zone
Ductile	Neoformed micas in mylonites	Ar-Ar	Reactivation age, cooling or isotopic resetting of shear zone
Ductile	Pre and synkinematic minerals in mylonites	U-Th-Pb	Deformation age in shear zone

Watanabe et al., 2008; Nuriel et al., 2011, 2017, 2019; Roberts and Walker, 2016; Pagel et al., 2018), in order to establish the age of reactivation of the shear zones (Kelley, 2002).

The U-Pb geochronology in calcite is proposed because calcite can register a rapid deformation event at low temperature in the upper crust, process that would be difficult to dated radiometrically with other minerals, since they do not have the time necessary to crystallize and record deformation under these conditions. Therefore, calcite constitutes an effective chronometer that relates deformation processes (development of faults and folds) and hydrothermal mineralization (cf. Roberts and Walker, 2016; Goodfellow et al., 2017; Mottram et al., 2018, Beaudoin et al., 2018; Roberts et al., 2020a). The main objective is the U-Pb dating of calcite crystals in mineralized fault planes that show textures filled by fracturing and sealing (crack-

seal) (Nuriel et al., 2011; Robert and Walker, 2016). Likewise, the material to be dated must be present in syntectonic structures upon the brittle deformational event, either as tension gashes or fault-gouge injection veins, and displaying some of the following characteristics (Nuriel et al., 2011; Dressel et al., 2018; Miranda et al., 2020; Roberts et al., 2020b): 1) crack seal textures with sliding in micro pull-apart structures; 2) crack seal textures with growth fiber lineation; 3) crack seal textures within implosion breccia structures; 4) sigmoidal, in echelon, and pocket-type shapes with normal kinematics, the latter associated with discontinuous veins, massive internal textures, brechoid or sheared.

To guarantee the simultaneity of the calcite and deformational event that gave rise to its precipitation, i.e., the temporal relationship between the fault, fracturing, fluid circulation, and the process of precipitation and growth of calcite (Ramsay, 1980; Barker et al., 2006; Nuriel et al., 2011), geochemical, microstructural and petrographic analyses should be integrated into the U-Pb geochronology of calcite minerals (cf. Nuriel et al., 2011; Miranda et al., 2020).

Low-temperature thermochronometers ((U-Th)/He and fission track) (Table 2) have been used in fault zones that show fluid circulation along the damage zones and in the core of the structure, these generally synkinematic upon the deformational event and characterized by the precipitation of hematite, magnetite and goethite in veins, breccias, and as patina along the sliding planes (cf. Tagami and Murakami, 2007; Wolfer et al., 2010; Gorynski et al., 2014; Hickey et al., 2014; Reiner et al., 2014; MacNamee and Stockli, 2015; Ault et al., 2016; Louis et al., 2019; Tagami, 2019). Under these conditions, the occurrence of iron oxides is the deformation dating target, these also record the temperatures in which the displacement of the structure was induced (cf. Wernicke and Lippolt, 1997; Shuster et al., 2005; Blackburn et al., 2007; Farley and Flowers, 2012; Evenson et al., 2014; Monteiro et al., 2014; Farley and McKeon, 2015; Ault et al., 2015, 2016; Cooperdock and Stockli, 2016; Miller et al., 2017; McDermott et al., 2017; Moser et al., 2017; García et al., 2018; Calzolari et al., 2018; Ault, 2020).

Other techniques applied to fault zones that present fluid circulation and that host mineralizations are dating of U-Th-Pb in hydrothermal monazite (cf. Bergemann et al., 2017, 2018; Ricchi, Bergemann et al., 2020; Ricchi, Gnos et al., 2020; Prent et al., 2020), Rb-Sr in feldspar crystals (cf. Tillberg et al., 2017; Tillberg et al., 2020), and Re-Os in hydrothermal sulfides (Vernon et al., 2014; Lawley et al., 2013; Dichiarante et al., 2016).

For fault gouge, the method of dating by electron spin resonance (ESR) has been used for quartz grains that make up this non-cohesive rock (Table 2) (cf. Ikeya et al. 1982; Ikeya, 1993; Lee and Schwarcz, 1994; Lee and Yang, 2007). The technique is based on the ability to analyze damage caused by natural radiation in geological and biological materials such as shells, corals, bones and fossilized teeth, by detecting alpha, beta and gamma radioactivity from the sample and its environment (Duval et al., 2011; Duval, 2018). This radiation produces defects in the crystal lattice that can trap electrical charges that eventually form a “paramagnetic center” that produces a detectable signal in ESR spectrometry. The height (intensity) of the spectrum obtained is proportional to the number of electrons trapped in each paramagnetic center, the radiation dose and its duration (Ikeya et al., 1982; Ikeya, 1993; Grün, 1989; Lee and Schwarcz, 1994; Lee and Yang, 2007; Qju et al., 2018; Duval, 2018).

In shear zones that exhibit deformations of the brittle-ductile transition, which include associations of cataclases, phyllonites, protomylonites and pseudotaquilites (Sibson, 1983; Simpson, 1986), as well as mineral reactions and fluids, the Ar-Ar isotopic system in micas and the pseudotaquilite matrix is the most widely used tool for absolute dating of deformational activity (Crone, 1996; Vinasco, 2001; Oyhantçabal et al., 2012; Bense et al., 2014; Süssenberger et al., 2017). This is because of substantial retentiveness of Ar in the structure of neofomed minerals (closed isotopic system) (Purdy and Jäger, 1976; Reddy and Potts, 1999; Harrison et al., 2009) in intervals of low-to-medium temperature, conditions in which the reactivation of structures normally occurs (Grégoire et al., 2009; Webb et al., 2010; Bonamici et al., 2004; Oriolo et al., 2016a). When pseudotaquilites are processed, it is recommended to exclude clasts present in the matrix,

since they can contribute foreign Ar derived from preexistent rock, which can yield ages without geological sense (Di Vincenzo et al., 2004).

The deformation age under middle crust conditions, mainly affected by the ductile regime, is traditionally determined by fine muscovites and neoformed biotites (cf. Freeman et al., 1998; Fossen and Dunlap; 1998; Müller et al., 2000; Vinasco, 2001; Whitmeyer, 2008, Vinasco and Cordani, 2012), and recrystallized minerals (cf. Kligfield et al., 1986; Oriolo et al., 2015, 2016b) (Table 2), which provide information on tectonic activation from cooling ages or isotopic resetting, given the low closure temperature of these isotopic systems (cf. van der Pluijm et al., 1994).

Nevertheless, there are debates about the meaning of the geochronological data of these isotopic methods when diffusion, dissolution and precipitation processes by creep processes and fluid-assisted recrystallization have developed in the fault rocks. This is because the age intervals obtained would be related to cooling processes of the system and not be directly related to the deformation timing (cf. Harrison et al., 1985; Jenkin, 1997; Carreras et al., 2013; Eberlei et al., 2015; Vissers et al., 2016; Druguet et al., 2018). For these conditions, in order to constrain the thermal history in the shear zone, in-situ Rb-Sr and Ar-Ar micas and U-Th-Pb zircon, monazite, titanite, rutile and apatite have been integrated. The latter are resistant to physical and chemical alterations of the system, and in which the reported ages are related to crystallographic deformation process, since the migration and formation of crystalline faces are subject to development without incorporation of radiogenic Pb in the crystal lattice (Dahl, 1997; Lee et al., 1997; Cherniak and Watson, 2000; Santos et al., 2003; Cherniak et al., 2004; Cocherie et al., 2005; Dahl et al., 2005; Harley et al., 2007; Reddy et al., 2007; Timms et al., 2011; Oyhantçabal et al., 2012; Bonamici et al., 2015; Erickson et al., 2015; Oriolo et al., 2016a; Kirkland et al., 2018 a, b; Giraldo et al., 2019; Odlum and Stockli, 2020; Ribeiro et al., 2020 a,b; Van Daele et al., 2020).

### 3.1.1 Considerations for mineral dating of fault rocks

For analysis and application of different isotopic systems is necessary to consider the closure temperatures of the minerals. This is because once the isotopic system has reached a specific temperature range, the host mineral does not exchange isotopes with the system (a closed isotopic system), so negligible diffusion rates are obtained, as therefore precise chronological information on the deformation event (cf. Dodson, 1973; Villa, 2002; Braun et al., 2006). Closure temperatures depend on physical factors such as grain size, mineral composition, cooling rate, and pressure, among others (cf. Dodson, 1973; Hames and Bowring, 1994; Grove and Harrison, 1996; Cherniak, 1995; Farley, 2000; Jenkin et al., 2001; Braun et al., 2006; Harrison et al., 2009). For this reason, to date fault rocks, it is necessary to rely on the reconstruction of geothermobarometric diagrams (P-T-t), geochemical analysis, thermodynamic modeling, and microstructural studies (cf. van der Pluijm et al., 1994; Stünitz, 1998; Steffen and Selverstone, 2006; Villa et al., 2014; Odlum and Stockli, 2020; Ribeiro et al., 2020 a,b).

The temperature reached during the deformation process must be less than or equal to the closure temperature of the isotopic system of the geochronological method used. Otherwise, the age obtained will be a cooling age of the isotopic system and not of the deformation (Vinasco, 2001). This is because dynamic recrystallization, neo-crystallization, metamorphic reactions and fluid circulation (dissolution-precipitation processes) take place during deformation, which modify the physical properties (reset) of the isotopic system of minerals in fault rocks (cf. Dunlap, 1997; Villa, 2002; Mulch et al., 2002; Mulch and Cosca, 2004; Harley et al., 2007; Cosca et al., 2011; Tagami, 2012; Harlov, 2015; Oriolo et al., 2016a).

In the case of U-Pb dating of carbonates (e.g., calcite), once the event of interest has been identified through field observations, petrography, cathodoluminescence and electron microscopy, one should confirm that: 1) the carbonate is contemporary or synchronous, or very close, to the geological event of interest; 2) the carbonate formed rapidly and under conditions that allowed enrichment in U and exclusion of

Pb; 3) there is no evidence of dissolution, recrystallization, or alteration (e.g., diagenetic); 4) there are no different generations of carbonate; 5) there are optimal concentrations of U in the mineral to ensure accurate dating, which can be determined by compositional maps (cf. Rasbury and Cole, 2009; Roberts and Walker, 2016; Roberts, 2019; Roberts et al., 2020a).

### 3.2 Radiometric dating of geological features adjacent to shear zones

Given the different geological processes (deformation, metamorphism, and fluid-rock interaction) related to the evolution of a shear zone, it is necessary to integrate geochronological and thermochronological studies of syntectonic elements of the structure, in order to constrain the age of deformation in shear zones (Fossen and Tikoff, 1997; Xypolias and Kokkalas, 2006; Vitale and Mazzoli, 2008; Carreras et al., 2013; Pennacchioni and Zucchi, 2013; Ribeiro et al., 2020b). These studies include:

1. Intrusive synkinematic (Table 3), which correspond to igneous bodies emplaced during reactivation episodes of fault systems, in which magmatic units are transported through the crust by shear zones that behave as channels connecting magma generation areas with the upper crust (D'Lemos et al., 1992; Reavy, 1989; Clemens and Mawer, 1992; Rosenberg, 2004). The emplacement of granitoids thus marks a direct relationship between tectonic deformation and magmatic bodies (D'Lemos et al., 1992; Druguet and Hutton, 1998; Brown and Solar, 1999; Grocott and Taylor, 2002; Vinasco and Cordani, 2012; Ávila et al., 2019) and delimits strongly deformed areas in which igneous bodies fill open spaces within brittle-ductile transition zones, mainly associated with transtensional deformations and limited by the traces of regional faults. To date these intrusives, it is necessary to perform a priori detailed analysis of magmatic fabrics (cf. Romn-Berdiel et al., 1997; Blumenfeld and Bouchez, 1988; Paterson et al., 1989; Bouchez et al., 1992; Paterson and Vernon, 1995; Paterson et al., 1998), in order to discriminate between pre-, syn- and postkinematic igneous bodies (Steenken et al., 2000; Rosenberg, 2004; Siegesmund et al., 2004; Wang et al., 2009; Oyhantçabal et al., 2009). Once determined from the magmatic fabric that the igneous body is synkinematic, the igneous units are dated by U-Pb dating of zircon, in order to obtain the age of emplacement of the igneous body, information directly related to reactivation periods of the shear zone.

2. Hydrothermally altered rocks related to shear zone activity (Table 3), for which the Ar-Ar dating method is mainly used for neoformed white micas, which provide information on tectonic reactivations, because they are related to fluids generated during deformational events in the shear zones (Vinasco, 2001; Vinasco and Cordani, 2012).

3. Blocks adjacent to the structure (Table 3) in which the exhumation ages are related to periods of deformation or reactivation of the fault zone, because low-temperature thermochronology dates the cooling ages, which in shear zone contexts are widely related to timing of the fault slip (cf. van der Pluijm et al., 1994; Stockli et al., 2002; Wells et al., 2000; Echler and Farley, 2003; Colgan et al., 2008; Bidgoli et al., 2015; Curry et al., 2016; Oriolo et al., 2016b; Abbey and Niemi, 2018; Collett et al., 2019; Heineke et al., 2019; Amaya-Ferreira et al., 2020). Exhumation ages are determined using low-temperature thermochronology methods (surface conditions of the lithosphere at ~10-km depth, for a normal geothermal gradient), including thermochronometers such as apatite fission track (AFT) and zircon fission track (ZFT), which give closing temperatures between ~110-120 °C and ~230-240 °C, respectively, under conditions of constant cooling and relatively rapid exhumation (Zaun and Wagner, 1985; Hurford, 1986; Laslett et al., 1987; Brandon and Vance, 1992; Ketcham et al., 1999; Bernet et al., 2002; Bernet et al., 2019).



#### 4. CONCLUSIONS AND RECOMMENDATIONS

For geological-structural mapping of shear zones, it is suggested to perform an analysis of photointerpretation and geoprocessing of images in which the trace of the structure is preliminarily delimited. Once defined, field transects must be made perpendicular to the strike of the deformed area in order to collect structural data with statistical representativeness and determine the distribution, kinematics and geometry of the central and marginal areas of the principal structure in the shear zone, as well as of satellite faults, branches and secondary structures, in order to define the timing of deformation events and rank the developed structures.

To classify fault rocks in the field, it is recommended to adopt the modified proposal of Sibson (1977;1979), Scholz (1988), Snoke et al. (1998), Killick (2003), Woodcock and Mort (2008) and Magloughlin (2010), which differentiated between cohesive and non-cohesive rocks, categorizes the rocks

TABLE 3.  
Synthesis of the material and proposed method for dating geological elements adjacent to the shear zone, as well as the information obtained thereby

Material for dating	Proposed methods for dating	Information obtained
Zircons of synkinematic intrusive	U-Pb	Age of emplacement
Neoformed micas of hydrothermally altered rocks	Ar-Ar	Age of reactivation of shear zone, cooling or isotopic resetting
Fission tracks in apatite and zircon of blocks adjacent to the structure	AFT and ZFT	Exhumation ages

according to the percentage of original and matrix grains, and describes the characteristics and deformation mechanisms developed during the deformation event.

The proposed methodology to determine and constrain the period of deformation in shear zones uses technical differences according to the deformational regime in which the fault rocks were generated, implements materials and methods of geological elements adjacent to the shear zone, and it is constituted by the following activities: 1) submicroscopic analysis of minerals; 2) dating of neoformed micas, fault gouge, and pseudotakylite matrix by the Ar-Ar method, and of calcite fibers and iron precipitates by the U-Pb and (U-Th)/He systems, respectively, under conditions of brittle/brittle-ductile deformation; 3) in-situ K-Ar dating of micas and U-Th-Pb of pre- and synkinematic minerals in mylonite rocks; 4) dating of synkinematic intrusives, adjacent blocks, and hydrothermally altered rocks by the U-Pb, AFT/ZFT and Ar-Ar methods, respectively.

The minerals that make up fault rocks are formed in the presence of fluids and conditions of varying pressure and temperature, and can be isotopic resetting during the evolutionary history of shear zones. Considering that the dating of fault rocks depends on the deformation temperature, it is suggested to establish the temperature from geothermometers and microscopic analysis of the microstructures. However, the presence of retrograde, recrystallization or dissolution-precipitation processes must be determined, since these processes substantially affect the closure temperature of the minerals used in the geochronological and thermochronological methods implemented and do not guarantee a closed isotopic system with negligible diffusion rates.

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