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Are fluid inclusion planes useful in structural geology?

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Abstract

Quantification of deformation or stresses in structural geology is often difficult due to the uncertainity of the paleodepth or fluid pressure. The best records of fluid percolation are paleofluids trapped as fluid inclusions in healed microcracks of the rock-forming minerals. Fluid inclusion planes (FIP) are mode I cracks which form in sets with a predominant orientation perpendicular to the least principle stress σ_3. However, the repeated microfracturing and healing of the rock-forming minerals yield complex superimposed patterns of healed microcracks. Such patterns are often difficult to interpret due to the lack of suitable chronological criteria. These problems have been recently documented and solved by coupling deformation studies, detailed examination at all scales of the relationships between trapped fluids and their host structures, and studies of fluid inclusions. This paper summarizes recent advancements in FIP analysis and suggests further ways for research. © 1999 Elsevier Science Ltd. All rights reserved.

1. Introduction

During the last few years, much work has been done on microcracks in rocks. Nevertheless a better understanding of microcrack creation, propagation mechanisms and spatial distribution is necessary, as microcracks affect many physical properties of rocks, such as their strength, seismic wave velocity and permeability. Compared to the large quantity of experimental data on crack initiation and propagation, only a few papers discuss the relationship between the preferred orientations of cracks and the regional framework of deformation.

Much recent work has been carried out on the determination of paleostress by using fault slip data and assumptions concerning the paleodepth and the fluid pressure magnitude. Stress magnitudes are generally calculated for dry conditions when the fluid pressure (P_f) is unknown, yielding a maximum value of effective stresses. Fluid pressure can be estimated in different ways, including estimation using fluid inclusions. Fluid migration in rocks is favored by fissure permeability which forms during brittle deformation. Evidence of paleofluid migration through fractured rock may be scarce, whatever the observation scale, when little change occurs in the mineral assemblages resulting from fluid–rock interactions (dissolution, alteration, new crystallization). The best records of fluid percolation are paleofluids trapped as fluid inclusions in healed microcracks (fluid inclusion planes, FIP). However, the repeated microfracturing and healing of the rock-forming minerals yield complex superimposed patterns of healed microcracks. Such patterns are often difficult to interpret due to the lack of suitable chronological criteria. These problems have been recently documented and solved by coupling deformation studies, detailed examination at all scales of the relationships between trapped fluids and their host structures, and studies on fluid inclusions.

2. What are fluid inclusion planes?

Fluid inclusion planes result from the healing of former open cracks and appear to be fossilized fluid pathways (review in Roedder, 1984; Fig. 1). Microcracks

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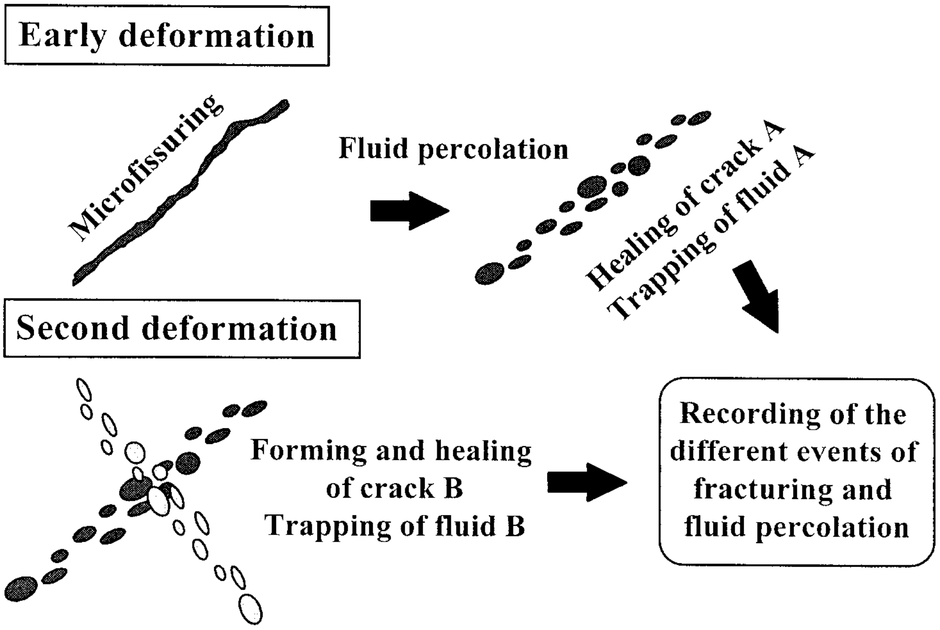

Fig. 1. Fluid inclusion planes are mode I cracks which form in sets with a predominant orientation perpendicular to the least principle stress σ_3. The repeated microfracturing and healing of the rock-forming minerals yield complex superimposed patterns of healed microcracks. FIP, which are good records of successive episodes of crack initiation and fluid migration, permit the elaboration of a chronology.

should provide valuable information about the local stress in rocks and can be assumed to be $(\sigma_1 - \sigma_2)$ planes (Tuttle, 1949; Wise, 1964; Lespinasse and Pecher, 1986). The FIP are mode I cracks that occur in sets with a predominant orientation perpendicular to the least principal compressive stress axis, σ_3. These mode I cracks propagate in the direction which favors the maximum decrease in the total energy of the system (Gueguen and Palciauskas, 1992). They do not disrupt the mechanical continuity of mineral grains and do not exhibit evidence of shear displacement like mode II and III cracks. The FIP are usually observed and characterized in minerals which crack according to the regional stress field, independently of their crystallographic properties (as demonstrated for quartz by Lespinasse and Cathelineau, 1990), and may easily trap fluids as fluid inclusions when healing. In some minerals (carbonates, feldspars), the fluids are not always preserved due to alteration or dissolution and cracks display more complex patterns resulting from the presence of cleavages, subgrain boundaries or twin planes. The rate of healing is rapid in quartz (compared to geological times) as shown by Smith and Evans (1984) and Brantley (1992).

Frequently, FIP form well defined networks which allow the determination of a chronology (Fig. 1). After a first generation of FIP, a second crack family can be formed with the trapping of different fluid. This second generation of FIP generally cross-cuts the first one. Thus, FIP are good records of successive episodes of crack initiation and fluid migration (Pecher et al., 1985). For each FIP, one can determine their dip direction, length, thickness and the microthermometric properties of the fluid inclusion trapped in the cracks (Fig. 2).

3. How can fluid inclusion planes be utilized?

3.1. Witness of stress orientation

The determination of paleostress orientations is a significant problem in understanding the tectonic history of any region. The analysis of the preferred orientations of FIP versus the successive average regional paleostress states has been documented (Kowallis et al., 1987, 1995; Laubach, 1989; Ploegsma, 1989; Ren et al., 1989). The analysis of the relationships between paleostress field and the geometry of FIP has been realized in the Le Bernardan open pit in the example of the La Marche granite (NW French Massif Central) (Fig. 3; Lespinasse and Pecher, 1986). The FIP exhibit several distinct preferred orientations on the scale of a grain, which may be observed in many samples. The orientations of the FIP are similar to those of micro- and mesoscale fractures in the granite. The dominant FIP direction is parallel to the main direction of regional shortening (NNE–SSW). Thus, FIP can be used as microstructural markers of paleostress fields like tension gashes.

More complex patterns are usually found within folded metamorphic rocks where the mechanical dis-

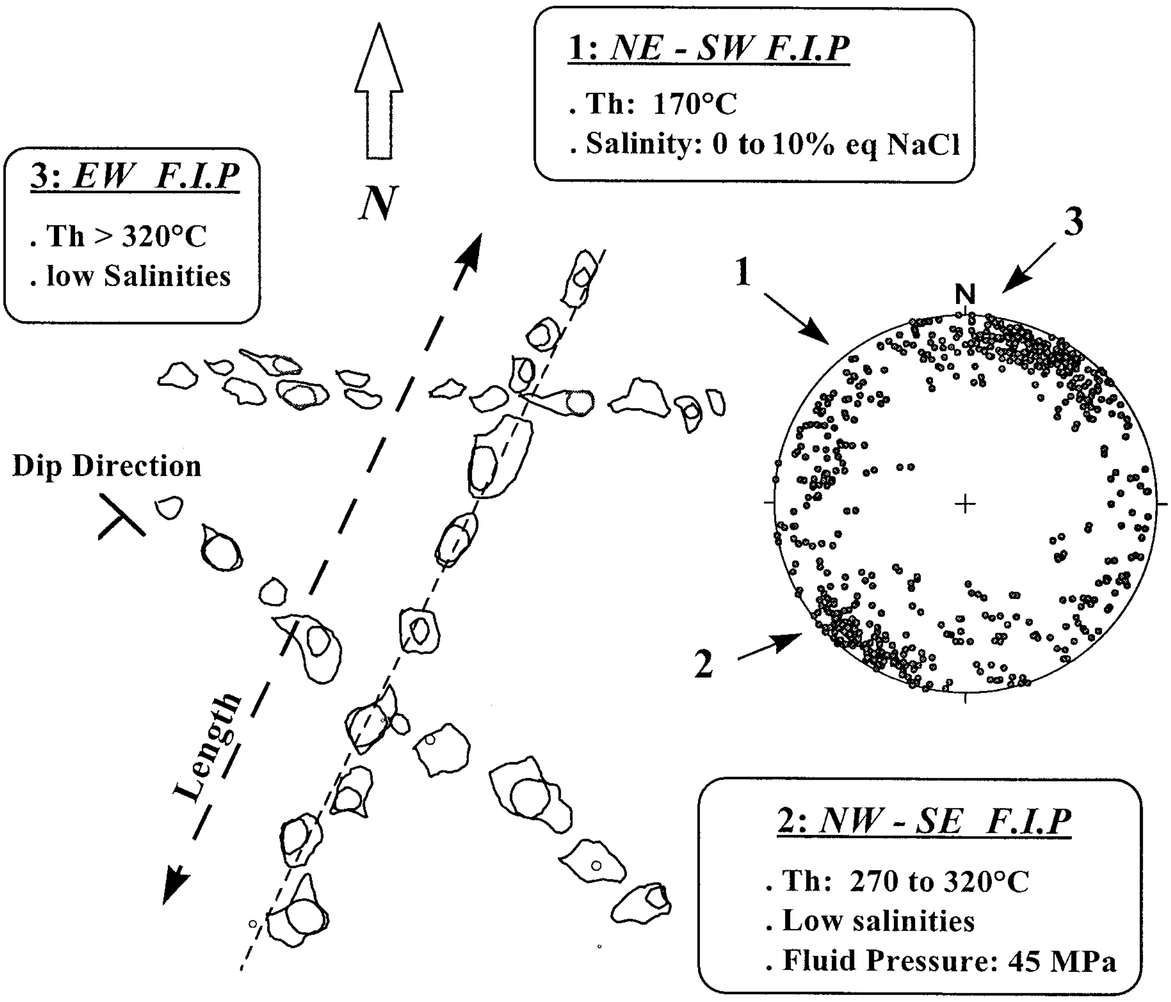

Fig. 2. Presentation of the different parameters collected for each FIP family (noted 1–3) in the vicinity of a major NW–SE strike-slip fault (Lespinasse and Cathelineau, 1995). Geometrical parameters (length, thickness, dip direction) are collected by using an image analyzer and presented on stereographic projections. Microthermometric properties of each fluid inclusion can be correlated to a specific crack. The poles of each FIP are plotted on lower hemisphere, equal area stereographic projection.

continuities (e.g. folds) induce local stress reorientations (Cathelineau et al., 1990). In addition, early microfissuring in metamorphic quartz is in part erased by quartz recrystallization, as hidden by late microfissuring associated with retrograde metamorphism (Alvarenga et al., 1990).

3.2. Fluid pressure in fault systems and paleostress quantification

One can use the physico-chemical differences among the inclusion fluids to separate different sets of FIP; however, it should be possible to use the FIP geometry to relate the different stages of fluid percolation to a regional succession of deformational events. Applications are important in the reconstruction of paleofluid pressure and stress quantification for a tectonic event.

In the case of brittle deformation, paleostress magnitudes can be estimated by using fault slip data, and rupture and friction laws for dry conditions (Angelier, 1989). However, estimation of stress magnitude is difficult if fluids are present during deformation. In that case, the lithostatic load and the fluid pressure are usually unknown. The quantitative estimation of the lithostatic load and the fluid pressure value during a tectonic event can be derived from paleofluid analysis in FIP (Lespinasse and Cathelineau, 1995; Meere, 1995). Since FIP are healed mode I cracks with a consistent orientation with respect to regional or local structures, stress and fluid features may be obtained for a given deformation event (Cathelineau et al., 1993; Pecher et al., 1985). This approach has been applied to a fault system which affects an Hercynian granite of the NW French Massif Central (Lespinasse and Cathelineau, 1995).

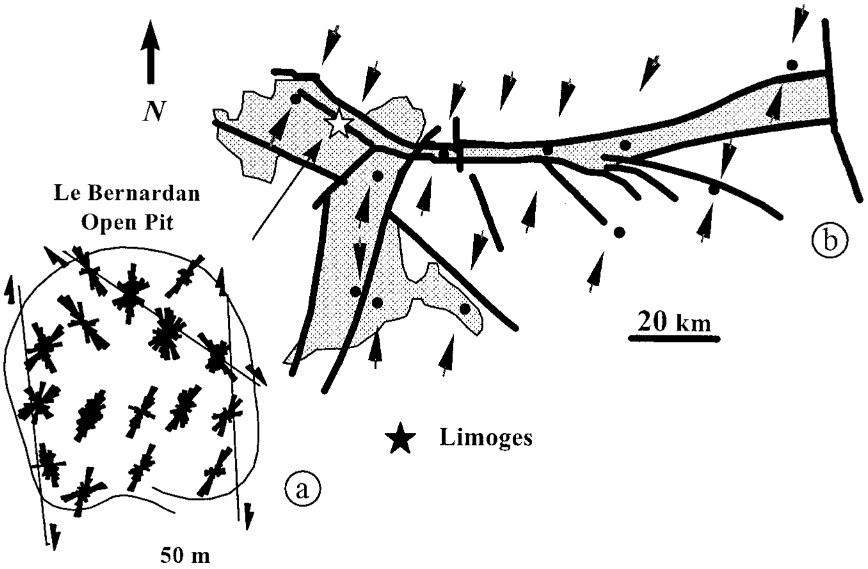

Fig. 3. Relationships between FIP orientations and regional paleostress field. (a) Sketch map of FIP orientations (rose diagrams) in the Le Bernardan open pit. The main orientation of the FIP is NNE–SSW. (b) Regional NNE–SSW compression in the La Marche zone (NW French Massif Central). Arrows show the direction of the maximum principal stress σ_1 at each sampling site. σ_1 was determined by striated fault plane analysis (Lespinasse and Pecher, 1986).

The method consists of linking the P, V, T, x properties of the fluid inclusions trapped in each crack family to the paleostress tensors. Since fluid composition and minimum trapping temperatures are known, the fluid density can be determined, and representative isochores for the studied fluid can be drawn. It provides estimates of the most probable range of pressures during fluid migration. In a fluid-saturated rock, the effective stress is given by $\sigma_n' = (\sigma_n - P_f)$ (Hubbert and Rubey, 1959). The fluid pressure at a depth z in a rock mass of average density ρ can be defined in relation to the overburden pressure (vertical stress) σ_v by means of the pore fluid ratio: $\lambda_v = P_f/\sigma_v$ (Sibson, 1981, 1989). Thus, the effective overburden σ_v' can be written:

$$\sigma_v' = (\sigma_v - P_f) = \rho g z(1 - \lambda_v) \quad (1)$$

where g is the acceleration due to gravity and z is the depth. When pore spaces are interconnected to the surface, $P_f = \rho g z$, and a state of hydrostatic fluid pressure prevails with $\lambda_v = 0.4$. However, if P_f is lithostatic, λ_v approaches 1 and the vertical effective stress (σ_v') is zero ($\sigma_v' = 0$). In an 'Andersonian stress state' (Anderson, 1951), during a strike-slip regime of faulting, $\sigma_2 = \sigma_v$ and is considered as vertical. The relationships between stress axes can be expressed as follows (Yin and Ranalli, 1992):

$$\sigma_1 - \sigma_3 = [2\mu\rho g z(1 - \lambda) + 2C]/[(\mu^2 + 1)^{1/2} + \mu(2R - 1)] \quad (2)$$

where R is a stress ratio $R = [(\sigma_2 - \sigma_3)/(\sigma_1 - \sigma_3)]$, μ is the static coefficient of rock friction and C is the cohesion of the rock. These relationships show that stress intensities inferred from this type of failure analysis depend strongly on the fluid pressure. The vertical stress magnitude can thus be determined, by the analysis of paleofluids trapped as fluid inclusions in minerals during the brittle deformation event. In spite of the uncertainities in the values of parameters such as P_f, σ_v, R and μ, a quantification of stress magnitudes has been attempted. Relations between the fluid pressure and the stress differences ($\sigma_1 - \sigma_3$) are based on Eq. (2). Uncertainities in the values of R and μ yield to the uncertainity domain included within two extreme lines.

In the case of the fault system which affects a Hercynian granite of the NW French Massif Central (Lespinasse and Cathelineau, 1995), a NW–SE compression has been defined from a population of 51 faults characterized by orientations around N60°E to N110°E for dextral strike-slips and N135°E to N175°E for sinistral movements. The stress ratio, R, has been determined with fault slip data to be around 0.52 ± 0.08 (Etchecopar et al., 1981). The dominant FIP trend is NW–SE, vertical or dipping toward the SW (Fig. 2). The fluid inclusions from NW–SE FIP are characterized by homogenization temperatures with a mode around 300°C and ice melting temperatures with a mode around −1.0°C. The possible P–T pairs of the trapping conditions were estimated at about 50 ± 10 MPa. Considering that fluids are trapped in faults under hydrostatic conditions (water density of 1000 kg/m^3, neglecting the density changes

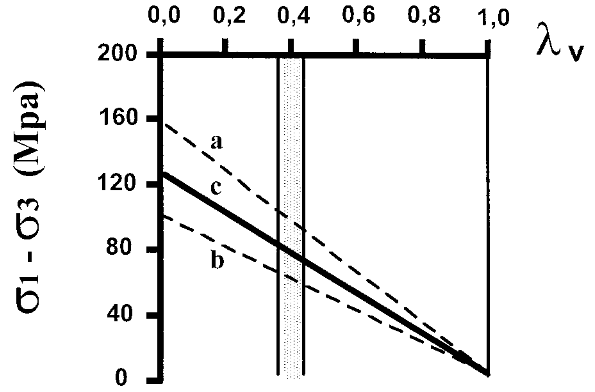

Fig. 4. Determination of the stress magnitude differences ($\sigma_1 - \sigma_3$) as a function of the pore pressure factor (λ_v) in the case of a Hercynian fault zone. The differential stress ($\sigma_1 - \sigma_3$) is determined in the case of strike-slip faults by using the Yin and Ranalli (1992) relations. Each line (a, b, c) is calculated with different values of R and μ. (Line a) $R = 0.52$ and $\mu = 0.58$. (Line b) $R = 0.46$ and $\mu = 0.66$. (Line c) $R = 0.64$ and $\mu = 0.50$. The hydrostatic domain corresponds to λ_v within the 0.37–0.43 range, assuming that rock density in the upper crust may vary from 2300 to 2700 kg/m^3.

with increasing temperature), the trapping depth of the major fluid migration (also present in NW–SE cracks) can be estimated at about 5 km. Thus, assuming a vertical column of granite (density of 2700 kg/m^3), $\sigma_v = 132 \pm 10$ MPa. The presence of fluid in rocks during deformation can drastically change the conditions of rupture in fault systems and also the stress magnitude. Schematic representation of the relations (Fig. 4) suggests the following: (1) for a lithostatic fluid pressure ($\lambda_v = 1$), the effective stresses are equal to zero; (2) for $\lambda_v = 0$ (dry conditions), the values of $\sigma_1 - \sigma_3$ are in the range of the results obtained from fault slip inversion; and (3) for hydrostatic conditions ($\lambda_v = 0.4$), $\sigma_1 - \sigma_3$ values are in the 70–105 MPa range for the studied case.

4. Fluid inclusion planes: what future?

Quartz-rich rocks supply fruitful objects to study FIP in a brittle context of deformation. FIP should provide (in absence of mesostructures) the best way to compare the evolution of fluids and the chronological sequence of fracture opening. Relevant results strongly suggest that FIP are useful in solving problems in structural geology (paleostress orientation and magnitude determination, and quantification of fluid pressure contemporaneous with a tectonic event).

Physical properties of rocks (such as seismic velocity, rheology, density, conductivity,...) and pressure dissolution reactions depend strongly on the presence of fluids in rocks. Discontinuities such as joints or cracks and fissures are potential sites for fluid circulation and have important implications for the hydraulic properties of the rock. It is therefore important to quantify fluid flow in these discontinuities, in order to characterize and understand fluid transfers. A major problem facing theoretical modeling of fluid flow is that the fracture porosity of a rock is often poorly known. The problem is more complex in considering paleofluid circulation in rocks. The only witnesses to paleofluid flow through a dense set of cracks in rocks are the FIP. Fissure permeability quantification depends strongly on the precise description of the fractured medium and on the theoretical approach used [geometrical models, based on the Snow approach (Snow, 1965) or statistical approaches (Long et al., 1985; Gueguen and Dienes, 1989; Gueguen and Palciauskas, 1992)]. Statistical approaches consider that the fractures are finite. The fissure permeability is directly related to the fracture lengths and therefore to their connectivity. Several works based on percolation theory (Broadbent and Hammersley, 1957; Stauffer, 1985) lead to a possible estimation of the permeability tensor (Canals and Ayt Ougougdal, 1997). Its determination needs a complete description of the three-dimensional geometry of the microcrack network, including definitions of crack family orientations, average lengths, apertures, and volumetric densities. Detailed analysis of the FIP geometry allows a quantification of these parameters and then, a quantification of the paleofissure permeability. Such approaches seem to be very useful in the case of economic geology or structural geology.

The FIP tool is at this time not widely used even though many other pertinent and potential uses exist. Other possible developments in the use of FIP concern the recording of paleoseismic events (Boullier and Robert, 1991), finite strain quantification (Onasch, 1990), in situ stresses analysis (Fleischmann, 1990), and more generally the microfissuring analysis in relation to deformation and fluid flow (Blenkinsop, 1990; Vollbrecht et al., 1991; O'Hara and Haak, 1992; Boullier, 1999).

Important applications of the FIP tool also concern economic geology. The geometry of the pore space at the time of pore fluid migration was investigated in several Au–quartz veins (for instance: Archean veins, Boullier and Robert, 1991; panAfrican veins, Zouhair et al., 1991; late Hercynian veins, Essaraj, 1992; Alpine veins, British Columbia, Boiron et al., 1992). In these different cases, the P, V, T, x composition of a fluid flowing in a specific direction during mineralizing events was determined by using the FIP tool. Therefore, the Au enrichment was attributed to fluids percolating in a specific direction. A precise chronology of the ore deposit formation was then established.

Is there a future for the FIP? The recent or future advances in the fluid inclusion analysis (such as quantitative determination of the elements present in the FI, dating of the FI,...) and in the characterization of the three-dimensional topology of the microcracks (tortuosity, roughness) by using different techniques such as cathodoluminescence (Boiron et al., 1992) strongly suggest that FIP would be considered as excellent records of the Earth's history. One can therefore consider that the future of research using FIP is full of possibilities.

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