Fractography of rock joints with special emphasis on granitic rocks

DRAFT MANUSCRIPT FROM PhD THESIS OF MARTIN ZIEGLER

Engineering Geology, ETH Zurich, Switzerland
martin.ziegler@erdw.ethz.ch

November 14, 2010

Table of contents

1. Definitions

Fractography is the study of micro- and macroscopic, natural as well as artificial fracture surfaces of various materials such as glasses, ceramics, metals, polymers, composites, or rocks. Fracture surface morphologies are described, analysed and interpreted to gain information about the stresses and mechanisms involved in the development of the fracture (Ameen 1995). Whereas most fractographic methods are used to ascertain reasons for failure of construction materials and engineering structures, tectonofractography interprets natural, mostly field macroscopic fracture surface phenomena of rocks and regional fracture systems in terms of the characterization of tectonic processes and accompanied geomechanical conditions (Bahat 1991). With tectonofractography, geological information as to fracture modes, fracture type and processes, paleo and in-situ stress directions and magnitudes, timing of the formation of differently orientated joint sets, location of joint origins, and (changes in) joint propagation velocities and directions can be obtained.

2. History of research

Most publications concerning fractography deal with glasses, ceramics or metals. In contrast, rock as a material is underrepresented, particularly in experiments. Most of the features described in glasses and various other materials can also be found in rocks. Understanding the fundamental processes may lead to a better understanding how fracture markings on rock surfaces can be interpreted. Quinn (2009) gives an overview about the history of fractography, and the interested reader is forwarded to his article. At this point, only the milestones of research regarding fractography of rock joints shall be presented. However, in the following chapters similarities and differences between observations on glass, metal or ceramic surfaces compared to rock surfaces will be explained (see also Kulander & Dean 1995; Hull 1999:250 et seqq.).

Woodworth (1895, 1896) was one of the first to document descriptions of macroscopic features of rock fractures and wrote the first classification. However, his ideas were initially not regarded as important and greater emphasis was put on the recognition of geometric joint patterns (Pollard & Aydin 1988). In addition, "the controversy over the tensile versus shear nature of rock joints with plumose markings need never have arisen if it had been recognized widely that plumose markings are characteristic of the tensile fracture" (Lutton 1969:2061), as can be seen in the articles of Sheldon (1912) or Swanson (1927). It was Hodgson (1961a) to conclude that the occurrence of plumose markings (see chapter 3.2.2) forbids discernible movement along the joint plane. He also modified Woodworth's classification. Not until the second half of the 20th century joints were studied fractographically in detail in many different rocks. Ameen (1995) presents an overview of the history of research in the field of rock fractography. According to his article, important contributions to the field following Woodworth's fundamental approach were made by Parker (1942), Hodgson (1961a, b), Roberts (1961), Bankwitz (1966), Syme Gash (1971), Kulander et al. (1979), and Bahat (1991). In addition, Lutton (1969) should be mentioned, who pointed out that the scale of plumose markings varies at least six orders of magnitude, from 10^{-6} m on crystal surfaces to at least 10 m in the field. Furthermore, Bahat & Rabinovitch (1988) were the first to develop the principles of estimating paleo stress magnitudes on jointed granite using fractographic markings. Later, Bahat et al. (1999) assigned the method to assess initial fracture stresses leading to the formation of exfoliation joints. Bucher (2006) and Sutter (2008) were the first to examine fractographic features in the Central Aar granite (Switzerland). Table 1 shows selected authors and their research related to rock fracture surface studies.

Tab. 1: Selected authors and research related to rock fracture surfaces

Year	Authors	Research topics
1895, 1896	J. B. Woodworth	First detailed description and classification of fracture surfaces
1942	J. M. Parker	Description of a few plumose markings on joint surfaces
1961	R. A. Hodgson	Modification of Woodworth's classification
	R. A. Hodgson	Fractographic markings on sedimentary rock joints (Arizona and Utah)
	J. C. Roberts	Discussion of the mechanisms causing feather-structures

1965	J. Gramberg	The cause of axial cleavage fracturing (brittle tensile) in rock samples; artificially caused fractographic markings on fractured rock surfaces
	P. Bankwitz	Some descriptions and interpretations of joint surfaces from Thuringia (Thüringisches Schiefergebirge)
1966	P. Bankwitz	Detailed description and interpretation of the system of fracture surface morphologies; parallelization of terms
1968	Z. T. Bieniawski	First experiments of terminal fracture velocities of rocks
1969	R. J. Lutton	Compilation of fractographic mapping symbols; scale range of plumose markings
1971	P. J. Syme Gash	Stress wave induced microfracturing as the cause of fracture surface markings
1977	O. L. Anderson & P. C. Grew	Review of stress corrosion theory concerning rocks
1979	B. R. Kulander, C. C. Barton & S. L. Dean	Review of the subject of fractography concerning core and outcrop investigations
	D. Bahat	Mechanical parameters of joint surfaces
1984	P. & E. Bankwitz	Interpretation of fractographic markings in terms of qualitative analyses of paleo stress orientations
1985	B. R. Kulander & S. L. Dean	Short review of the geometry of hackle plumes
1986	D. Bahat	Criteria to distinguish hackle from en echelon fringes and (natural) joints from artificial cracks
1988	D. D. Pollard & A. Aydin	Review of the mechanisms of jointing with respect to surface structures
1991	D. Bahat	Comprehensive review of the subject of fractography and new insights
1995, 1997	P. & E. Bankwitz	Drilling-induced fractures; determination of maximum horizontal in-situ stress orientations (KTB German Continental Deep Drilling Project)
1995	B. R. Kulander & S. L. Dean	Description of morphological features and evolution of fractures produced in the laboratory
1999	D. Bahat, K. Grossenbacher & K. Karasaki	Interpretation of fractographic markings concerning the process of exfoliation; estimation of tensile fracture stress
2003	D. Bahat, P. & E. Bankwitz	Pre-uplift joints (South Bohemian Pluton, Czech Republic); introduction of a quantitative method of joint classification
2004	D. Bahat, P. & E. Bankwitz	Introduction of the index of hackle raggedness
2005	L. Savalli & T. Engelder	Subcritical growth of joints in layered rocks
	D. Bahat, V. Frid & A. Rabinovitch	Comprehensive review of tensile rock fractures and new insights
2006	S. Bucher	Descriptions of fractographic markings along the Schöllenen Gorge (Switzerland)
2008	B. Sutter	A few descriptions of fractographic markings at the Grimsel Pass area (Aar Massif, Switzerland)

3. Fractographic features

"The features are produced when the fracture or joint deviates locally from its mean plane of propagation. Deviations are attributed to tensional crack-tip stresses that are affected by the amount of strain energy locally released, the passage of sonic waves, changes in propagation rate, changes in far-field stress, and material discontinuities related to varying elastic moduli and textures. Even though these far-field stresses may be attributed to Mode II or III shear loading, resultant crack-tip stresses in brittle material are tensile." (Kulander & Dean 1995:72)

Given that the surface morphology of an opening mode[1] joint was not changed or destroyed due to shearing (see e.g. Woodworth 1896; Hodgson 1961a; Bankwitz 1965), weathering (and subsequent erosion), stylolitization (especially in marly chalks), or mineralization after its formation one will be able to distinguish different types of fractographic markings and thus subdivide the joint's structure. Experiments as well as theoretical calculations (e.g. Nemat-Nasser & Horii 1982) have shown that fractures form at Griffith-cracks (Griffith 1921, 1924), which are assumed to be randomly distributed flaws occurring in brittle solids. The interaction, specifically the merging of these (micro)cracks, leads to the development of macroscopic fractures. The initial fracture point is surrounded by an area called the mirror (Fig. 1).

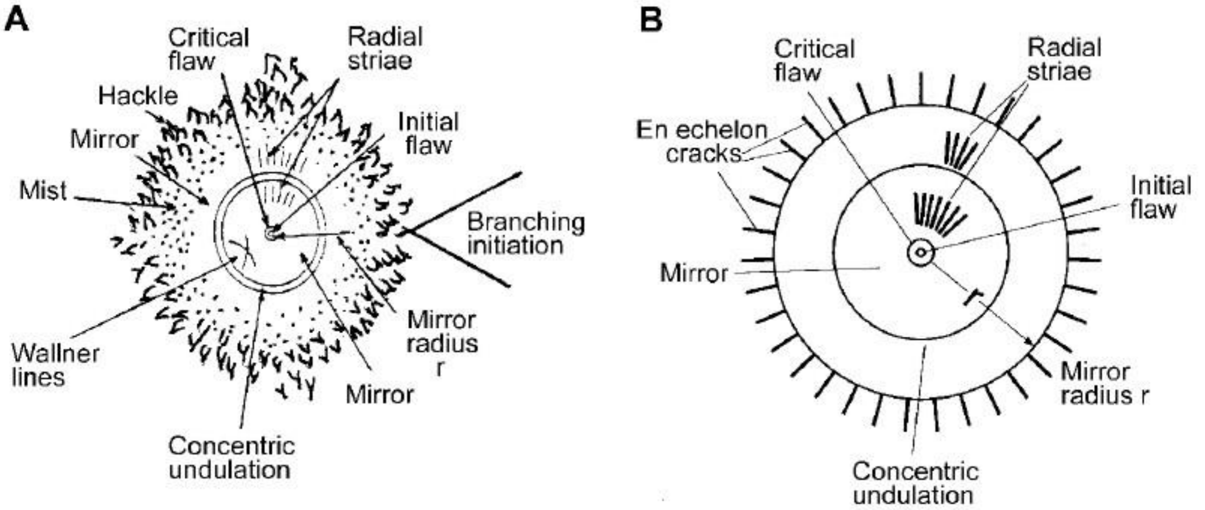

Fig. 1: (A) Schematic representation of a fracture surface that forms under postcritical conditions (see chapter 4 for explanation), showing the concept of initial flaw (fracture origin), mirror plane, radial striae (plumes), concentric undulations (ripple marks), Wallner lines, mist, hackle, branching initiation (bifurcation), and the mirror-plane radius, r (arrow). (B) Schematic representation of a fracture surface that often forms under subcritical conditions. The mirror plane radius, r (arrow) is measured from the critical flaw to the inner boundary of the fringe of en echelon cracks. Most features in A and B are similar. However, in B, en echelon cracks replace hackles (from Bahat et al. 2003).

The mirror region is a relatively smooth surface on which different surface markings can occur. If the mirror becomes successively rougher, it passes into the mist region. The mirror and mist are surrounded by a fringe, which 'end members' can either be an en echelon fringe or a hackle fringe (Bahat et al. 2005). The different features of a fracture surface are shown in Fig. 1 and will be described below. In addition, possible terminations of joints will be discussed. Fractographic markings occur on burial (sedimentary rocks), cooling (magmatic rocks), syntectonic (pre-uplift), uplift, and post-uplift joints (Bahat 1991; Engelder et al. 1993, cited in Bahat et al. 1999. Characteristic tectonofractographic features of joints can help to distinguish between different joint categories and to understand joint genesis (Bankwitz & Bankwitz 1994).

---

[1] According to e.g. Tapponnier & Brace (1976), rocks under stress fail through extension of (natural, pre-existing) cracks. Like Kranz (1979) explains, this does not mean that mode II and III stresses do not exist but it states that mode I stresses are dominant.

3.1 Fracture origin

Variable discontinuities at different scales can be considered as Griffith-flaws. Synonymously, Griffith-flaws are also called crack or fracture origins, microcracks, nucleation or initiation points. As Bahat et al. (2005) point out, joint nuclei can vary widely in size, shape and orientation. Their scale depends on the homogeneity of the material. Whereas Griffith-flaws in monocrystalline quartz have dimensions of 1 nm to 1 µm, in heterogeneous materials they range from 1 µm to 1 mm or greater. Nucleation points can be crystal inhomogeneities (preferential cleavage, twin lamellae, inclusions, defects), grain boundaries, cavities, layer boundaries, bedding irregularities, fossils, concretions, large grains or clasts. Generally, microcracks in crystalline rocks develop perpendicular to the least normal stress (e.g. Kranz 1983). However, crystallographic directions and stress heterogeneities at the grain scale lead to deviations of microcrack orientations (e.g. Vollbrecht et al. 1991). In granitic rocks, microcracks are usually too small to be identified macroscopically (Pollard & Aydin 1988). Although many granitic rocks are thought to be isotropic, quarrymen have long known and taken advantage that granites may split in different directions unequally. The rift, grain, and hardway are the three directions of 'easiest splitting'[2]. Fujii et al. (2007), for instance, found the anisotropy of Inada granite from Japan to be due to preferentially orientated microcracks. Uniaxial tensile fractures parallel to the rift, grain, and hardway show different surface topographies as well as different mineral ratios. This may also be true for many other granitic rocks. Since most rocks contain any number of microcracks, it is likely that a joint consists of numerous merging fractures (Kulander et al. 1979; Bahat 1991; cf. Bahat et al. 2005:Fig. 3.24c). Even if the location of the fracture origin cannot be directly identified, there may exist several indirect alternatives such as striations or rib marks to trace back the joint's origin(s) (see below).

3.2 Mirror and mist

As stated above, the term mirror describes a polished-like surface and surrounds the critical flaw. This is comprehensible for most glasses and metals, since very smooth surfaces reflect light comparable to real mirrors. In rocks, this is only true for volcanic glasses. However, in most rocks, the mirror describes a region on the joint surface that is relatively smooth compared to the fringe and hence constrained not only by the stress intensity, but by the coarseness of the rock, its mineralogical composition, and also by crystal orientations. Different flaw sizes are the cause for unequal surface morphologies between glasses and joints. While fractures in glasses or metals begin with a smooth mirror surface, rock joints commonly show plumose structures on the mirror from the beginning (Savalli & Engelder 2005). The mirror is also known as the joint plane (Woodworth 1895, 1896), master joint, main joint face, or the parent plane in contrast to the surrounding fringe zone. Well developed, i.e. smooth mirror surfaces, are generally limited to e.g. fine grained granitic or sedimentary rocks. A mirror surface may become relatively flat if "the two surfaces of the fracture contain statistically the greatest number of perturbed bonds" (Bahat 1979:82).

---

[2] *Rift*: planes of easiest splitting in quarries; *grain*: planes slightly more difficult to split than the rift; *hardway*: planes of more difficult splitting and often orthogonal to the orientation of the rift and grain (Johnson 1970). Various German quarryman terms for cleavage properties of granite are given in Kieslinger (1951, 1958).

The orientation of the mirror plane is perpendicular to the remote (far-field) tensile stress (or least principal compressive stress) and parallel to the maximum compressive stress direction.

The mirror is terminated by the development of velocity hackles, which are chaotic in nature and mark the inception of the mist region. Velocity hackles are formed by uncontrolled (unstable) fracture propagation at high speeds (0.6 of the shear wave velocity), or high stresses (postcritical stress intensities) through rotation of the principal tension direction (Hull 1999; Quinn 2009, and references therein). They cause the foggy appearance of the mist region. The mirror boundary lies between the mirror and the mist. However, differentiation between mirror and mist is scale dependent and continuous and hence not possible in every material. In the latter case, the mirror boundary is located between the mirror and the fringe. Velocity hackles are generally not observed on rock joint surfaces, which implies that joints form at stable propagation rates (Kulander et al. 1979; Kulander & Dean 1995). Bahat & Rabinovitch (1988) explain the missing of the mist band in ceramics and rocks by the granular nature of the material.

As indicated above, several surface features occur on the mirror plane (but are not necessary restricted to it), which can be used to determine the fracture origin and the propagation history of the fracture. Two common types of surface markings can be found: arrest marks and striations. On (coarse) rock surfaces, however, these features are often difficult to detect. In addition, the light angle and the viewing direction also play an important role in gaining the best contrast of surface features (Bucher & Löw 2009).

3.2.1 Arrest marks

Arrest marks[3], also called hesitation lines, ripple or rib marks, originate when the propagation of a fracture decreases in velocity, stops and moves on again. Closely spaced rib marks can indicate slow propagation velocities (Kulander & Dean 1985). Rib marks are also described as conchoidal structures and appear as curvilinear ridges or furrows. Arrest lines indicate a simultaneous halt of the fracture front. A crosscut through an arrest mark in the direction of fracture propagation will show that the ripples are mostly asymmetrical in forms of cuspate waves or lines separating tilted panels (see Fig. 2). Rounded forms are not thought to indicate an arrest of the fracture front. Rib marks are orientated perpendicular to striations (Bahat et al. 1999; Pollard & Aydin 1988, and references therein) and surround the fracture origin in circles or elliptical rings. Circular shapes originate when the fracture velocities in different directions are equal, other velocity distributions will lead to elliptical arrest mark shapes (e.g. Savalli & Engelder 2005). The concave side points to the fracture origin, making rib marks invariable indicators of the direction of propagation. This can be helpful if only parts of a joint plane are visible. Bucher & Löw (2009) mentioned that arrest lines are rather rare at the Schöllenen Gorge.

A second type of ripple marks are undulations. In contrast to arrest marks, undulations originate through rapid fracture propagation and differ in shape. They show rather symmetrical crosscuts (Bahat et al. 1999).

---

[3] The term arrest mark should be used with caution, because it postulates a halt of the fracture front, while the term ripple or rib mark is often used in a broader sense (Bahat 1991; Bahat 2005:128 et seq., 530).

The formation of rib marks and undulations is an expression of tilt. They require an out of plane (mixed mode I/II) stress field (Pollard & Aydin 1988). In other words, rib marks or undulations result when the tensile stresses at the fracture front interact with elastic vibrations normal to the fracture front, which are caused by broken bonds and emanate from the crack front (Poncelet 1958).

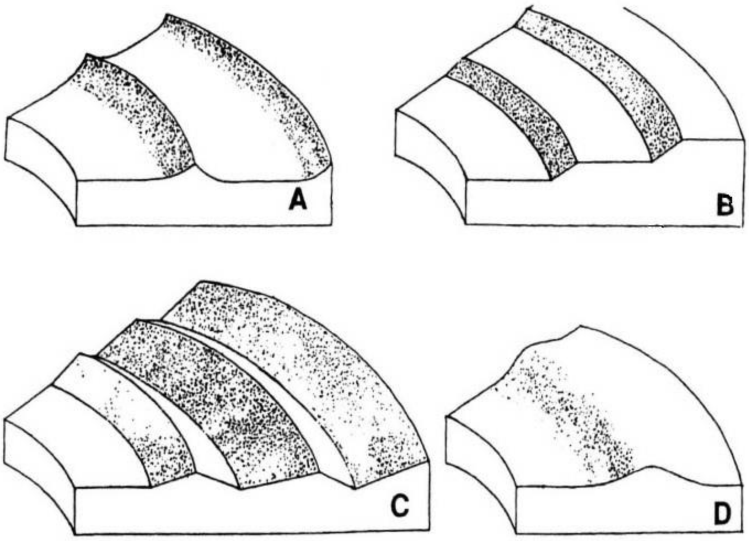

Fig. 2: Cross-section morphologies of rib marks (from Kulander & Dean 1995). A: cuspate waves; B, C: lines separating tilted panels; D: rounded forms that may not indicate an arrest of the fracture front.

3.2.2 Striations

Striations[4], also called striae, (en echelon) steps, barbs, lances, or river lines, are very common features of fracture surfaces and are parallel to the direction of crack propagation (Bahat 1991). According to Pollard & Aydin (1988:1187), striations can be described as "curvilinear boundaries with differential relief between adjacent surfaces" and "either radiate from the origin [...] or fan away from a curvilinear axes." The chevron-like pattern consisting of an origin, an axis, and hackle marks is most commonly called a feather structure (Woodworth 1896), plumose structure (Pollard & Aydin 1988), or simply plume. Further, the terms chevron and herringbone mark are used. Plumes exhibit different morphologies (see Fig. 3): straight (S-type), curved (C-type), rhythmic, with common origin (radial), spiral, bifurcating, or chaotic, and also combinations are known but less common (Bahat 1991, and references therein). Bahat (1991) summarizes that striae or river line patterns are cracks bridging en echelon segments. Striae as well as en echelon fringes originate by similar mechanisms (mixed mode I/III). Plumose structures consist of a "complex micro-structure of tensile and shear zones (Bahat et al. 2007:289). Morphologically, however, striae originate within the mirror region, while en echelon steps are restricted to the fringe (see chapter 3.3). Among others, Hull (1999) points out that the development of river line patterns causes progressive roughening of the fracture surface. Several small radial steps unite during propagation and form single larger steps.

---

[4] Synonymously the term hackle is widely used, but according to Rice (1974) and Rice (1984, cit. in Bahat 1991:118 et seq.), it should be used only for fractures outside the mirror region. While striations reflect relatively low stress intensities, hackles possibly refer to higher stress intensities. Although often used interchangeably, the terms striation and striae respectively should only be used within the mirror plane, and the terms steps, lances, or river patterns for the fringe (Bahat 1991).

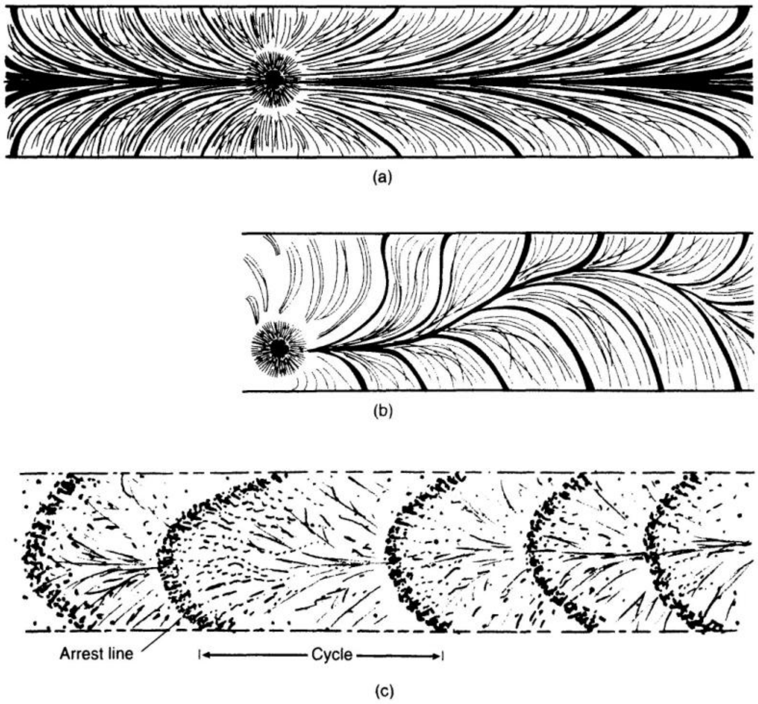

Fig. 3: Types of plumose structures (from van der Pluijm & Marshak 2004:143). (a) straight (S-type), (b) curved (C-type), (c) rhythmic with multiple arrest lines.

3.2.3 Wallner lines

"Rapidly propagating fractures in glass do not show subtle Wallner lines because of the rough fracture surface. Wallner lines, unless of large wavelength and amplitude, are generally not apparent on rocks because their subtle nature is easily disrupted by coarse transient features and the polycrystalline granular texture of the rock itself. However, Wallner line amplitudes are directly proportional to the size of the flaw [...] responsible for the sonic waves. This fact enables discernible Wallner lines to form even in coarse rocks if conditions permit [...]." (Kulander et al. 1979:30 et seq.)

"The Wallner lines are generally very faint and their limited detectability in polycrystalline materials has made them of little value for the determination of crack velocity in studies of ceramics and rocks." (Bahat 1991:83)

Wallner lines, named after the researcher who first explained their origin in 1939, are wavelike undulations[5], which are "caused by a temporary excursion of the crack front out of plane in response to a tilt [...] in the axis of principal tension" (Fréchette 1990:14). The excursion is caused by an elastic pulse. They are similar to rib marks, but oblique to striation. Sometimes different sets of Wallner lines can be distinguished. If the distortional wave velocity is known, one can extract the crack velocity. Wallner lines are mostly studied in materials like glasses (e.g. Kerkhof 1973) or single crystals such as quartz (e.g. Payne & Ball 1976). It is all but impossible to detect them in coarse, polycrystalline rocks as Kulander et al. (1979) or Bahat (1991) mention, and the author hasn't found any report about Wallner lines on rock surfaces. In contrast, Hull (1999:252) mentions that Wallner lines (and also mist regions) may occur on rock joints, but without giving any reference. However, he alludes

---

[5] The author uses the term undulation on purpose, because the term ripple mark, which is often used to explain Wallner lines is also often found as a synonym of arrest lines, but Wallner lines do not indicate a (temporary) termination of the fracture front.

that the scale of fracture phenomena and Wallner lines respectively is dependent on microstructure and micro-deformation processes.

3.3 Fringe

According to Woodworth (1895, 1896), an en echelon[6] fringe consists of border planes (b-planes) and cross-fractures (c-fractures), which were later called en echelon cracks and steps respectively (Fig. 4). Bahat (1991), Hull (1999) and Bahat et al. (2005:130 et seqq.) point out that en echelon fringes are common on natural joints, while other fringe types, the hackle fringe and the cuspate hackle fringe, are rather uncommon in nature. The formation of hackle fringes requires unstable propagation (see chapter 4), which is rare in rocks but common in e.g. glasses (e.g. Frid et al. 2005). In contrast to en echelon fringes, the formation of hackle fringes creates more new area per given nominal area of fringe, i.e. en echelon fringes lose energy during their propagation not used in the creation of new surfaces (Bahat et al. 2002). This energy loss is attributed to the 'mode III operation' which causes plastic deformation and subsequent heat loss. While en echelons originate essentially in mixed mode I and III (mode II is thought to be of minor importance), hackle fringes are interpreted to develop in mode I (Bahat et al. 2002, and references therein). Bahat (1986) and Bahat et al. (2004:Tab. 1) specify how to distinguish between the different fringe types and introduce an index of hackle raggedness, called IHR. The development and mechanical interpretation of en echelon cracks is e.g. reviewed by Pollard et al. (1982). Below, only the en echelon fringe end member shall be explained.

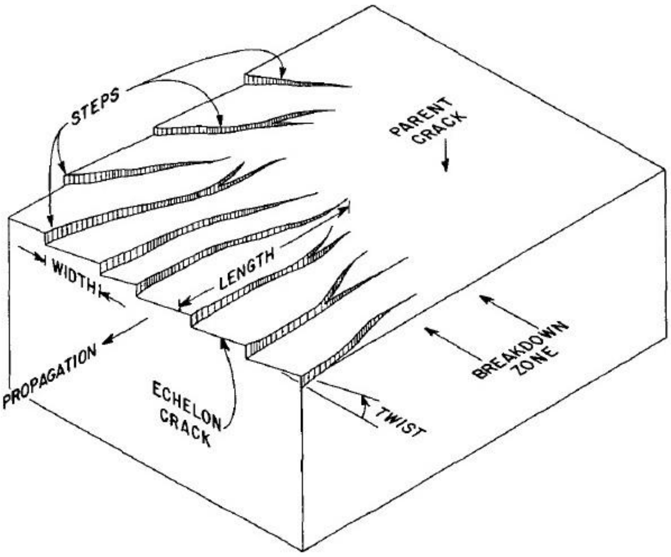

Fig. 4: The geometry of an en echelon fringe zone surrounding a parent crack (mirror) (after Pollard et al. 1982, modified from Lutton 1971).

The fringe en echelon cracks and steps can be described in terms of width, length, orientation, tilt, and twist angle as shown in Fig. 4. En echelon cracks overlap each other sub-parallel like shingles with a uniform orientation in relation to the master joint (see quotation from Pollard & Aydin 1988 in

---

[6] The French term *en échelon* describes the staggered or overlapping arrangement of linear geological features such as joints, faults or folds (MacDonald & Burton 2003).

chapter 3.4.2). The surface of the en echelon joints is rather smooth. In contrast, the steps between adjacent cracks are rough (Bahat 1986). These steps that bridge the tensile en echelon segments are shear fractures, and they systematically approach orthogonal relationships (Bahat 2010). En echelon crack surfaces may exhibit plumose structures (Woodworth 1896; Hodgson 1961a). The height of the steps increases with increasing distance from the fracture origin (river line pattern, see chapter 3.2.2). The sequence in which fringe steps and en echelon cracks develop is variable. Steps can form after (Woodworth 1896; Bahat 1986:203), before (Bankwitz 1966), or even contemporaneously with cracks (Bahat 1991). In addition, Bankwitz & Bankwitz (2004) show that the relationship of tilt and twist angles of early formed steeply dipping joints in granitic plutons may indicate the depth of first joint formation and intrusion depths respectively.

The formation of en echelon and hackle fringes requires a rotation of the local principal stresses near the parent-crack tip, which means an out of plane stress field (mixed mode I/III). Fringes are an expression of twist (Pollard et al. 1982).

3.4 Intersection, interaction and termination of joints

Intersection, interaction, and termination characteristics of joints can be interpreted in terms of relative age differences and formation sequences between joints or joint sets and estimations of the change of principle stress orientations.

3.4.1 Intersection

Several intersections can be distinguished. X-intersections develop when an older, continuous joint set is (systematically) cut by a second joint set at an acute angle. Sometimes the latter joints are not as continuous as the pre-existing ones and incidentally stop or show slight offsets. If both joint sets are continuous, it is often not possible to gain relative ages without additional observations. T-interactions, however, are invariable evidence to differentiate between older and younger joints. With approximately right angles, T-intersections are also well known as orthogonal joint systems. Younger joints stop at older ones. Y-intersections are typical of discontinuous contraction joints and well known in mud crack patterns and columnar joints of basalt for instance. However, these joints may originate through very different processes. Ideally, the joints intersect at 120° angles (Davis & Reynolds 1996). Minor forms are hexagonal joint patterns on igneous rock surfaces (see e.g. Ziegler 2008:126), which are counted among superficial weathering phenomena and hence not of interest for tectonofractographic studies.

3.4.2 Interaction

> "Each joint enhances propagation of its neighbour by inducing a tensile stress in the vicinity of the neighbor`s tip. As the tips overlap, however, the induced stress becomes compressive, and the propagation energy drops precipitously, so that joint growth will tend to stop. This explains why slightly overlapped en echelon joints are so common in nature." (Pollard & Aydin 1988:1199 et seq.)

The stress field around a joint tip can change the stress field of a neighbouring one and affect growth, growth orientation or the manner of termination (Pollard & Aydin 1988). Bahat (1991) describes several interaction types of adjacent propagating cracks in chalks (Fig. 5). Kranz (1979:Fig. 1), on the other hand, mentions three basic types of interactions: en echelon (≈ d, e), en passant (≈ c), and crack-pore interaction.

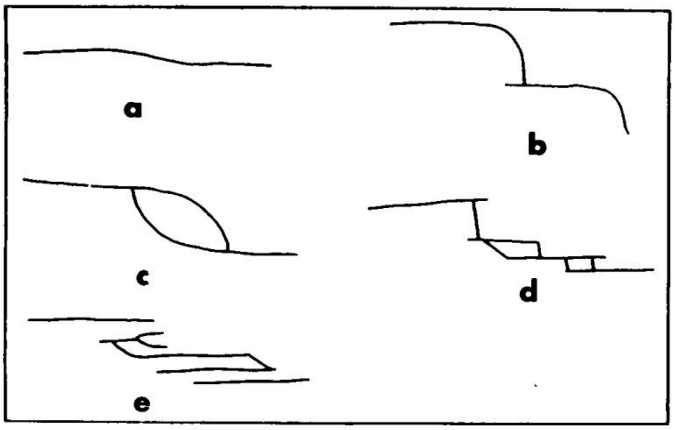

Fig. 5 a-e: Fracture interaction types observed from the Middle Eocene chalks at Beer Sheva (from Bahat 1991). (a) Asymptotic and (b) suborthogonal abutting curves, (c) oppositely propagating cracks along curved trajectories (also called tip-to-plane), (d) parallel joints with or (e) without straight, suborthogonal or oblique single or multiple connecting cracks.

3.4.3 Termination

According to Pollard & Aydin (1988, and references therein), experimental data show that fractures terminate either in a plastic zone or a zone of microcracking, which reduces the stress concentration at the joint tip and may stop fracture propagation. Consequently, if fracture growth starts again, arrest marks are formed (see chapter 3.2.1). Joints often terminate at discontinuities such as other joints, lithologic boundaries, or faults. Depending on the intersection location of a joint with any other visible surface, various traces of joints and their terminations can be studied. Joint traces may hook and stop, hook radically to form T-intersections with adjacent joints, break up into en echelon segments (see chapter 3.3), or simply stop without changing orientation (Davis & Reynolds 1996). Another type of termination is branching before the joint dies out. Joint branching, also called bifurcation or forking, occurs at terminal velocities and high stress intensities respectively (see chapter 4 and 5.2). The high K_I (mode I stress intensity) values enable the propagating crack to form additional cracks at an angle to the original one. With fracture velocity experiments on norite, Bieniawski (1968b:Fig. 6) showed that forking is associated with attainment of the rock's terminal fracture velocity.

4. Estimation of fracture velocity

One of the most challenging questions concerning joints is how long it takes them to form and whether it is possible to gain information about relative as well as absolute fracture velocities from fractographic markings. Not without cause, Kulander & Dean (1985:90) wrote that "in all but a few special cases it is impossible to determine actual fracturing velocities [...] for natural joints." First, we need to know the parameters that control fracture velocity. Bieniawski (1968a, 1968b) was one of, if not the first to investigate fracture velocities in rocks. He experimented with norite using an ultra-high speed camera and showed that the rock's terminal fracture velocity is dependent on the velocities of shear, longitudinal, or Rayleigh waves of the rock, which had been previously theoretically predicted. Bahat et al. (2003) were the first to introduce quantitative fractographic analysis of joints based on experimental and theoretical data from glass, ceramics, polymeric materials, as well as metals. In addition, they outlined analogies between rocks and other materials regarding their behaviour while subjected to tensile stresses. Also Pollard (2000:1365) states that "the mode [in which glass for example fractures] is eminently suitable for the taxonomic classification and description of fractures in the field". Rocks may thus be treated as natural ceramic bodies (Bahat 1979).

Wiederhorn & Bolz (1970) and Evans (1974) studied glass and ceramic fractures and showed how crack velocity is dependent on the tensile stress intensity under subcritical[7] conditions. They obtained a graph plotting fracture velocity v normal to the stress intensity factor (mode I), K_I. Bahat et al. (2003) assigned this curve in relation to a pre-uplift joint set in granite from the South Bohemian Pluton (Fig. 6). According to Wiederhorn & Bolz (1970), Evans (1974), Bahat et al. (2003) and Savalli & Engelder (2005, and references therein), region I (A-B) shows slow crack growth at essentially 'equilibrium' conditions which is attributed to stress corrosion, also called static fatigue. Stress corrosion may be defined as environmentally induced subcritical crack growth under sustained stress (Anderson & Grew 1977, and references therein). Furthermore, we can see that the curve in point A does not start at zero K, indicating that below a certain threshold stress intensity factor, K_0, subcritical crack growth is thought to stop (Fig. 6 right). This threshold is dependent on the environmental conditions (Darot & Gueguen 1986). Region II (B-C) is constrained by the rate of fluid transport to the crack tip. Here, v is constant and independent of K_I. With regard to the studied granite, the shape of the curve between B and D is unclear. Region III (C-b) exhibits high velocities and is independent of reactants and stress corrosion. Point E (or 'o' in the right plot in Fig. 6) shows a velocity overshoot within region III and is the beginning of postcritical crack propagation, when K_I exceeds K_{Ic}, the critical stress intensity factor. The fracture velocity reaches a terminal velocity plateau implying that crack velocity becomes independent of K. Dynamic fracture propagation occurs at velocities greater than 10^{-1} m/s, while static fracture propagation (e.g. stress corrosion, creep) takes place at velocities less than 10^{-1} m/s (e.g. Anderson & Grew 1977; Swanson 1984, and references therein). Olson (1993) summarizes that subcritical fracture growth velocities can be as low as 10^{-10} m/s. As Fig. 6 indicates, the formation of a mirror plane may cover a broad range of velocities up to a magnitude of 10^3 m/s (Kerkhof 1975). Furthermore, en echelon formation is interpreted to occur at reduced velocities (Müller & Dahm 2000; Bahat et al. 2001). One has to keep in mind that there are still many uncertainties with respect to the shape of the curve and limitations of the proposed methodology (Bahat et al. 2003).

As shown in Fig. 6, ripple marks (arrest marks) on the mirror plane are formed by relatively slow fracture velocities. According to Bahat et al. (2003), ripple marks are often superposed by plumose structures (plumes) and hence the latter are interpreted to occur under similar conditions (but slightly

---

[7] A (tensile) crack propagates under subcritical conditions if the stress intensity factor K (see equation 2, chapter 5.2) at its tip is below the critical stress intensity factor K_{Ic} ($K < K_{Ic}$). Subcritical crack growth is also called stable, in contrast to unstable fracture propagation with $K \geq K_{Ic}$.

more intense). En echelons are thought to form over a broad velocity spectrum and it appears that they are not sensitive to fracture velocity. Furthermore, it is not clear if en echelons originate under low K_I values or close to K_{Ic}. However, "the appearance of [an en echelon fringe] signifies a stage of fracture characterized by an increase in the stress intensity factor K_I and the velocity of fracture propagation" (Bahat 1979:84). Mist (m) and hackle fringe (h) are diagnostic of postcritical conditions at rapid, terminal velocities. Hackle fringes are thought to be the 'dynamic end member' and en echelons the 'quasi-static end member' of fringe types (Kerkhof 1975; Kulander et al. 1979; Bahat et al. 2002; Bahat et al. 2004).

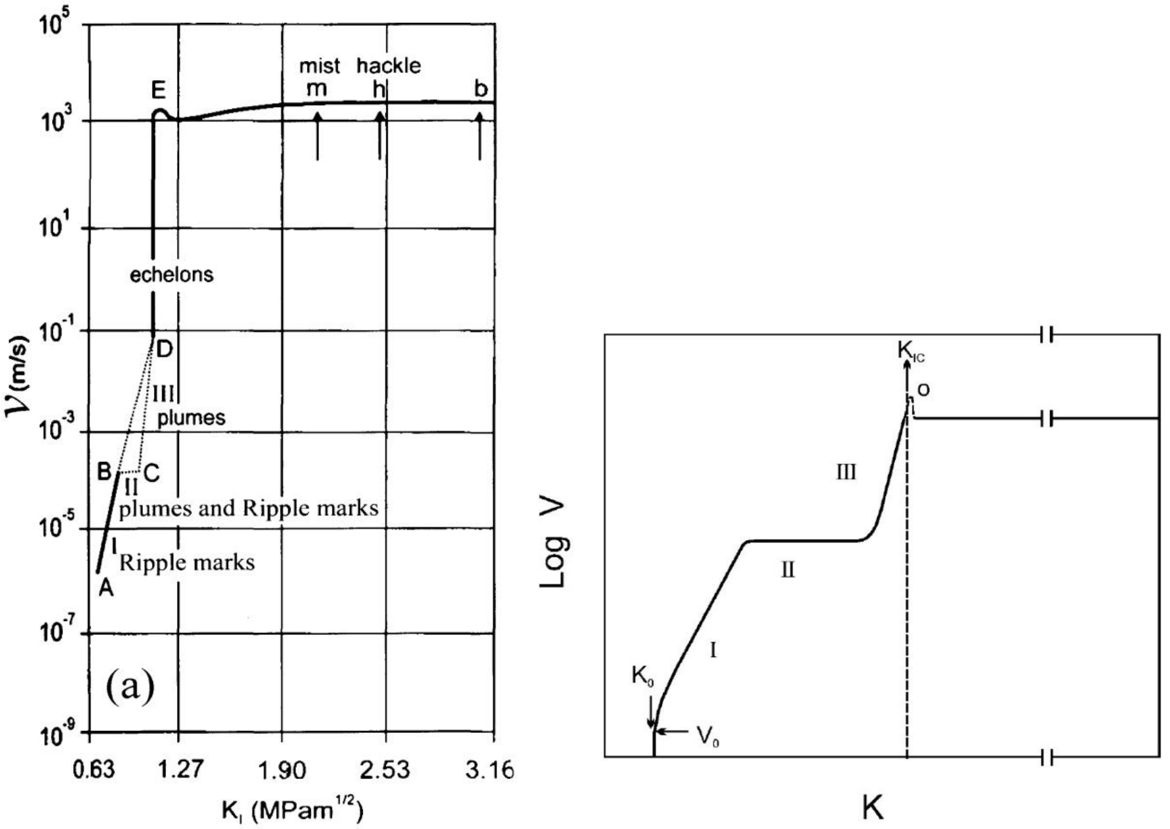

Fig. 6: (left) V versus K_I diagram for pre-uplift joints in granite from the South Bohemian Pluton, synthesized from laboratory experiments (from Bahat et al. 2003). (right) Schematic variation of stress intensity factor (K) with crack velocity (v). K_0: stress corrosion limit, v_0: velocity at stress corrosion limit, K_{Ic}: mode I critical stress intensity factor, I/II/III: distinct regions (from Bahat et al. 2003, and references therein).

In addition, a single joint is thought to form at different stages with different propagation velocities. Consequently, a joint plots not as a single point onto the curve in Fig. 6 (left) but as an area ranging between minimum and maximum velocities. Also cyclic fracture processes are recorded by alternating mirror and fringe regions, related to multiple bifurcations (Bankwitz 1965:Fig. 5; Bahat et al. 2001; Bahat 2005:Fig. 2-30c-e), or by multiple ripple marks on the mirror plane. Bahat et al. (2005:Fig. 4.35b) show a v-K_I plot of several pre-uplift joints in a granite quarry in the South Bohemian Pluton (Czech Republic) based on the joints' fractographic morphologies. They interpret the joints to form within a broad spectrum from subcritical to postcritical conditions and assume drastic changes in fluid pressure to be the main cause leading to diverse fractographic features. However, as indicated above many assumptions are necessary and fractographic morphologies must be identified and interpreted, which may lead to imprecise estimations of fracture velocity.

An important conclusion to be drawn is that crack propagation not only occurs when the stress intensity factor K_I equals the critical value K_{Ic}. Subcritical crack growth and stress corrosion respectively must be considered, making brittle failure of rocks an environmental- and time-dependent process. Stress corrosion is facilitated by a chemically active environment and controlled by the supply of reactants to the growing crack tip and by chemical reactions between the fluid and the rock. Under special conditions like high temperatures or reaction rates, diffusion, dissolution, ion exchange, and microplasticity also enhance subcritical crack propagation (Atkinson & Meredith 1987). Anderson & Grew (1977), Swanson (1984) and Schultz (2000) give reviews of fracture propagation with special emphasis on subcritical crack growth. Particularly Swanson (1984) explains the difficulties when applying techniques originally developed for synthetic materials to complex materials like rocks.

Furthermore, the peak stress obtained from laboratory experiments, such as compressive strength tests, at high strain rates is rate-dependent, and at lower strain rates stress corrosion may become increasingly important (e.g. Paterson & Wong 2005). Especially in geological studies, time-scales and time-dependent processes should be considered. Savalli & Engelder (2005:436, and references therein) conclude that "a number of models support the assertion that stable or quasi-static crack growth is the most common type of joint propagation within the Earth". Not only stress intensity or time, but also the rock's structure plays an important role. Besides chemical or mineralogical differences, the large number of potential fracture origins in rocks, coupled with low tensile elastic strain energy, is thought to cause low fracture velocities, which stay well below the rock's sonic velocities (Bieniawski 1968a, 1968b; Kulander et al. 1979). Swanson (1984) experimented with Westerly granite and found that small fracture velocities ($v \approx 10^{-7}$ m/s) and high relative moisture content (100 %) increase the amount of intergranular in contrast to transgranular fractures, especially in quartz. This shows that fracture velocity and pre-existing flaws have a major impact on the fracture path. These observations have to be kept in mind when assigning laboratory results from different materials to rocks.

Radial plumose markings and concentric undulations from the exfoliation joint described by Bahat et al. (1999, see chapter 5.2) indicate simultaneous formation, and Bahat et al. (1999) assume that the joint propagated subcritically with velocities below 4×10^{-5} m/s for dry conditions and below 10^{-2} m/s for water-saturated conditions. The mirror plane radius was calculated to be 96.4 ± 7.7 m. Accordingly, the formation of the exposed exfoliation joint would have required more than two weeks (dry condition), or more than an hour (wet condition). The location of the joint on the Half Dome cliff above the water table, and the low porosity of the rock, suggest dry fracture conditions.

Even if we are able to roughly estimate fracture propagation velocities on the basis of certain fractographic markings, there are some major problems. For instance, arrest marks indicate stops in the joint's development. At stress intensity conditions below the threshold intensity factor K_0 there is no evidence of the period of time of propagation arrest. Furthermore, many joints do not show (clear) fractographic markings and one has to identify if this indicates slow joint propagation due to stress corrosion in a relatively homogeneous stress field, or if the absence of surface structures is a secondary effect due to weathering for example.

5. Estimation of paleo and in-situ stresses

5.1 Stress directions

The diverse morphological features of joints are interpreted to be restricted to joint formation primarily in tensile mode. Such joints are parallel to the maximum compressive principal stress orientation. Fractography allows both paleo (Bankwitz 1978; Bankwitz & Bankwitz 1984) and in-situ stress orientation (Bankwitz & Bankwitz 1995, Bankwitz & Bankwitz 1997) estimations. Bankwitz (1978) mentions two important facts. First, the orientation of joints is independent of the orientation of the initial crack. Secondly, the orientation of the fringe, which deviates from a (small) parent joint, is a more convenient indicator for the principal stress orientation than the orientation of the parent joint. Joints without fringes directly show the principal stress orientation. In an early model, Bankwitz & Bankwitz (1984) show how different joint morphologies correspond to the principal stress orientations (Fig. 7). Although this model is out-dated and in some parts there are major doubts concerning basic assumptions (e.g. a single stress field is assumed, no influence of pore pressures on the effective stress), it gives an idea of the principal stress field. The shortcomings are discussed in Bahat (1991).

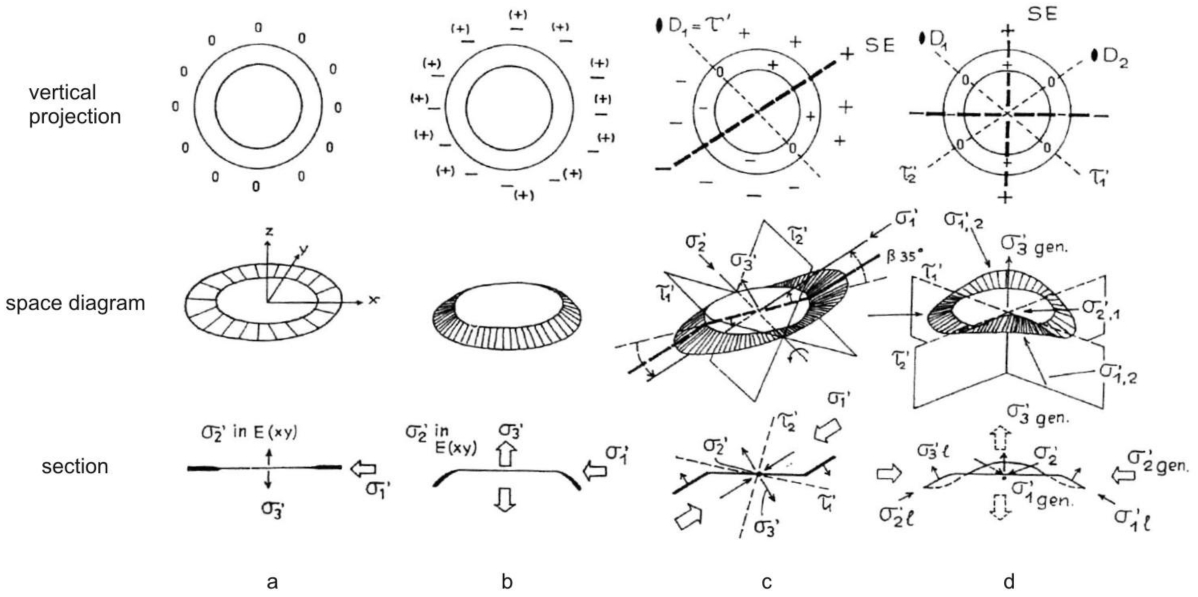

*Fig. 7 a-d: Basic morphological types of joint surfaces concerning their fringe zones and their stress interpretation (after Bankwitz & Bankwitz 1984). (**a**) Nearly planar (two-dimensional) joint; rarely realized in rock. (**b**) Fringe zone of typical tensional joints. (**c**) Characteristic shape of joints formed under compressive stress with superposition of a tectonic compressive stress; the fringe zone is parallel to the plane of pressure. (**d**) Frequent special type of natural joints with geometry similar to core disking. **a**, **b**: no segments; **c**: two segments; **d**: four segments; +, - signs: joint curvature, deviation from the main plane; SE: plane of symmetry; D_1, D_2: rotation symmetry axes and trace of the plane of shear stress τ; $\sigma_1 > \sigma_2 > \sigma_3$ principal stresses.*

Bankwitz & Bankwitz (1995, 1997) studied core samples from the KTB drilling (German Continental Deep Drilling Project) near Windischeschenbach in the Bohemian Massif, and showed that coringinduced fracture surface morphologies yield an estimation of in-situ maximum and minimum horizontal stress orientations. Comprehensive fractographic studies on drill cores were rarely performed previously.

5.2 Stress magnitudes

The distance between the joint origin (critical crack) and the mirror plane boundary is called the mirror radius, r (see Fig. 1). The radius is used to estimate the (far field paleo tensile) fracture stress, σ_f. However, this implies that mirror planes without fringes have no mirror boundary and no r value and hence no paleo stresses can be calculated. Bahat (1979) and Bahat & Rabinovitch (1988, and references therein) used the following (semi-empirical) equations (1-3) to estimate paleo stresses[8]:

$$\sigma_f = \frac{A_H}{\sqrt{r}} \qquad (1)$$

The mirror constant, A_H, is a material property and can be experimentally defined by a bending test (Mecholsky et al. 1974). Equation (1) relates fracture stress, σ_f, and mirror radius, r. The smaller the mirror, the larger the stress at the origin. Corresponding to Quinn (2009, and references therein), the actual shape of the mirror may be determined more accurately with a branching stress intensity factor criterion than with the simple stress - mirror radius relation. However, equation (1) is still widely used, because of its convenience and accuracy for small mirror sizes. Within the scope of this introduction, it is not possible to specify to which extent equation (1) is accurate for large mirrors like Bahat et al. (1999) used to account for stresses driving rock fractures (see below).

Furthermore, Irwin (1958, cited in Bahat & Rabinovitch 1988) found that for an infinitely sharp elastic crack in a plate, the stress intensity factor, K, is defined by

$$K = \sigma_f \times \sqrt{\alpha \pi c_{cr}} \qquad \text{(2) \textit{stress intensity method}}$$

with α being a correction for shape and for deviation from the infinite body assumption. The crack tip functions as a stress concentrator. A_H and K have the same units (MPa·m$^{1/2}$). The fracture stress, σ_f, represents the 'far-field' (tensile) stress. In addition, there is a constant relation between the mirror radius, r, and the critical flaw size, c_{cr}, for glasses as well as ceramic materials. A constant ratio is assumed to be also relevant for rocks.

For fracture branching, Congleton & Petch (1967, cited in Bahat & Rabinovitch 1988) found the following equation:

$$\sigma_f = \frac{2G \times \sqrt{\frac{E\gamma}{\pi}}}{\sqrt{r}} \qquad \text{(3) \textit{surface energy method}}$$

---

[8] Since the principal concepts about crack formation are assumed from studies of materials other than rocks, it is important to understand the basic concepts, especially stress intensity and the wing crack model. The author is aware that a comprehensive review of the topic of fracturing is out of the scope of this introduction to rock fractography. The history of research and basic concepts regarding fracture mechanics, starting with the classic papers from Inglis (1913) and Griffith (1920, 1924), are reviewed and theories are discussed e.g. by Atkinson (1987), Gramberg (1989), Bahat et al. (1999), and Paterson & Wong (2005).

where G, E and γ refer to an enhancement for fast or slow propagation, Young's modulus, and surface energy respectively. The equations show that there are two different ways to estimate fracture stress. First, by measuring r in the field and using the constant ratio for r and c_{cr}, and values for K from laboratory measurements (equations (1) and (2)). Second, by measuring r in the field and using values for surface energy from laboratory measurements (equation (3)). Both methods relate to the mirror radius. The surface energy method is assumed to be more reliable than the stress intensity method, because it is based directly on the mirror radius (Bahat & Rabinovitch 1988; Bahat et al. 1999).

For a granite from East Sinai, Bahat & Rabinovitch (1988) obtained σ_f results ranging for a shallow flaw from 2.9 to 10.1 MPa with an average value of 6.5 MPa, and for a semi-circular flaw from 4.6 to 15.8 MPa with an average value of 10.2 MPa for the first method. For branching occurring in feldspar (feldspar exhibits higher surface energy values compared to micas or quartz, see references cited in Bahat & Rabinovitch (1988)), the second method yielded fracture stresses from 2.4 to 6.0 MPa with an average value of 4.2 MPa. Bahat & Rabinovitch (1988) assumed the low stress values to result from persistent fatigue conditions. Studying the formation of an exfoliation joint on Half Dome (granodiorite) in Yosemite National Park, Bahat et al. (1999) even calculated significantly lower σ_f values. The stress intensity method yielded stresses ranging from 0.01 to 1.48 MPa, while the surface energy method yielded values ranging from 0.06 to 0.37 MPa. The low fracture stress indicates a prolonged, subcritical propagation process. Based on the assumed constant ratio $r/c_{cr} = 15.3$, Bahat et al. (1999) obtained a 6.3 m critical crack radius. This implies that the examined exfoliation joint had developed from a pre-existing joint. Concerning the formation of exfoliation joints, also Mitchell (2008) assumes that "the long-term strength of rock may be at least an order of magnitude less than that indicated in laboratory strength tests over short time scales", but gives no explanation to the cause (see also Mandl 2005).

[Displacement measurements of joints and estimations of strain and joint density respectively in granodiorite of the Sierra Nevada Batholith (at Ward Lake and Florence Lake, Mt Givens) yield initial stresses, which range between 1.2 ± 1.0 MPa (1.7 ± 0.9 MPa) and 10 ± 6 MPa (26 ± 11 MPa) (Segall & Pollard 1983). These values represent the magnitude of the sum of far-field, tensile stresses including internal fluid pressure. The studied joints are thought to have formed syntectonically, as suggested by chlorite-epidote joint fillings.]

In addition, equation (1) implies that joints, in terms of natural fractures formed at low stresses, exhibit large mirror planes compared to the size of hackle (fringes), whereas artificial joints caused by high stresses (for example in quarries) show small mirrors surrounded by relatively long and narrow radial hackles (see Bahat 1986:Fig. 5). This may help to distinguish between natural and artificial fractures. Furthermore, artificially induced fractures often are caused by a single event, while natural joints often are due to a cyclic fracture process (see chapter 4).

6. References

Ameen, M. S. (1995): Fractography: fracture topography as a tool in fracture mechanics and stress analysis. An introduction. In: Ameen, M. S. (ed.): Fractography: fracture topography as a tool in fracture mechanics and stress analysis (Geological Society, London, Special Publications, 92): 1–10.

Anderson, O. L. & Grew, P. C. (1977): Stress corrosion theory of crack propagation with applications to geophysics. In: Reviews of geophysics and space physics, **15** (1): 77–104.

Atkinson, B. K. (ed.) (1987): Fracture mechanics of rock. 534 pp. London: Academic Press (Academic Press Geology Series).

Atkinson, B. K. & Meredith, P. G. (1987): The theory of subcritical crack growth with applications to minerals and rocks. In: Atkinson, B. K. (ed.): Fracture mechanics of rock. London: Academic Press (Academic Press Geology Series): 111–166.

Bahat, D. (1979): Theoretical considerations on mechanical parameters of joint surfaces based on studies on ceramics. In: Geological Magazine, **116:** 81–92.

Bahat, D. (1986): Criteria for the differentiation of en echelons and hackles in fractured rocks. In: Tectonophysics, **121:** 197–206.

Bahat, D. (1991): Tectonofractography. 354 pp. Berlin, Heidelberg, New York: Springer.

Bahat, D. (2010): Personal correspondence. 12.03.2010.

Bahat, D.; Bankwitz, P. & Bankwitz, E. (2001): Changes of crack velocities at the transition from the parent joint through the en échelon fringe to a secondary mirror plane. In: Journal of Structural Geology, **23:** 1215–1221.

Bahat, D.; Bankwitz, P. & Bankwitz, E. (2003): Preuplift joints in granites: evidence for subcritical and postcritical fracture. In: Geological Society of America Bulletin, **115:** 148–165.

Bahat, D.; Bankwitz, P. & Bankwitz, E. (2004): The index of hackle raggedness on joint fringes. In: Cosgrove, J. W. & Engelder, T. (ed.): The initiation, propagation, and arrest of joints and other fractures (Geological Society, London, Special Publications, 231): 103–116.

Bahat, D.; Bankwitz, P. & Rabinovitch, A. (2002): Comparison of the new fracture areas created by the formation of en echelon and hackle fringes on joint surfaces. In: Z. geol. Wiss., **30** (1-2): 1–12.

Bahat, D.; Frid, V. & Rabinovitch, A. (2005): Tensile fracturing in rocks. Tectonofractographic and electromagnetic radiation methods. 569 pp. Berlin, Heidelberg: Springer.

Bahat, D.; Grossenbacher, K. & Karasaki, K. (1999): Mechanism of exfoliation joint formation in granitic rocks, Yosemite National Park. In: Journal of Structural Geology, **21:** 85–96.

Bahat, D. & Rabinovitch, A. (1988): Paleostress determination in a rock by a fractographic method. In: Journal of Structural Geology, **10** (2): 193–199.

Bahat, D.; Rabinovitch, A.; Frid, V. & Brosch, F. J. (2007): Cycles of sub-critical tensile and shear alternating fracturing in diminishing dimensions, under tensile loading. In: Int. J. Frac., **148:** 281–290.

Bankwitz, P. (1965): Über Klüfte I. Beobachtungen im Thüringischen Schiefergebirge. In: Geologie, **14**: 241–253.

Bankwitz, P. (1966): Über Klüfte II. Die Bildung der Kluftfläche und eine Systematik ihrer Strukturen. In: Geologie, **15**: 896–941.

Bankwitz, P. (1978): Über Klüfte IV. Aspekte einer bruchphysikalischen Interpretation geologischer Rupturen. In: Z. geol. Wiss., **6** (3): 301–311.

Bankwitz, P. & Bankwitz, E. (1984): Über Klüfte V. Die Symmetrie von Kluftoberflächen und ihre Nutzung für eine Paläospannungsanalyse. In: Z. geol. Wiss., **12** (3): 305–334.

Bankwitz, P. & Bankwitz, E. (1994): Event related jointing in rocks on Bornholm island (Denmark). In: Z. geol. Wiss., **22** (1/2): 97–114.

Bankwitz, P. & Bankwitz, E. (1995): Fractographic features on joints of KTB drill cores (Bavaria, Germany). In: Ameen, M. S. (ed.): Fractography: fracture topography as a tool in fracture mechanics and stress analysis (Geological Society, London, Special Publications, 92): 39–58.

Bankwitz, P. & Bankwitz, E. (1997): Fractographic features on joints in KTB drill cores as indicators of the contemporary stress orientation. In: Geol. Rundsch., **86:** 34–44.

Bankwitz, P. & Bankwitz, E. (2004): The relationship of tilt and twist of fringe cracks in granite plutons. In: Cosgrove, J. W. & Engelder, T. (ed.): The initiation, propagation, and arrest of joints and other fractures (Geological Society, London, Special Publications, 231): 183–208.

Bieniawski, Z. T. (1968a): Fracture dynamics of rock. In: The International Journal of Fracture Mechanics, **4** (4): 415–430.

Bieniawski, Z. T. (1968b): The phenomenon of terminal fracture velocity in rock. In: Felsmech. Ingenieurgeol., **6:** 113–125.

Bucher, S. P. (2006): Talklüfte im östlichen Aarmassiv (Schöllenen-Schlucht). Unveröff. Diplomarbeit. Betreut von Prof. Simon Löw. 104 pp. ETH Zürich, Geologisches Institut.

Bucher, S. P. & Löw, S. (2009): Talklüfte im zentralen Aaregranit der Schöllenen-Schlucht (Kanton Uri, Schweiz). In: Swiss J. Geosci., **102** (3): 403-421.

Congleton, J. & Petch, N. J. (1967): Crack branching. In: Philosophical Magazine, **16:** 749–760.

Darot, M. & Gueguen, Y. (1986): Slow crack growth in minerals and rocks: Theory and experiments. In: Pageoph, **124** (4/5): 677–692.

Davis, G. H. & Reynolds, S. J. (1996): Structural geology of rocks and regions. 2. Aufl. 776 pp. New York: Wiley.

Engelder, T., Fischer, M. P., Gross, M. R. (1993): Geological aspects of fracture mechanics, a short course manual notes. Geological Society of America Annual Meeting, Boston.

Evans, A. G. (1974): Slow crack growth in brittle materials under dynamic loading conditions. In: Int. J. Frac., **10** (2): 251–259.

Fréchette, V. D. (1990): Failure analysis of brittle materials. 136 pp. Westerville, Ohio: The American Ceramic Society (Advances in Ceramics, 28).

Frid, V.; Bahat, D. & Rabinovitch, A. (2005): Analysis of en échelon/hackle fringes and longitudinal splits in twist failed glass samples by means of fractography and electromagnetic radiation. In: Journal of Structural Geology, **27:** 145–159.

Fujii, Y.; Takemura, T.; Takahashi, M. & Lin, W. (2007): Surface features of uniaxial tensile fractures and their relation to rock anisotropy in Inada granite. In: Int. J. Rock Mech. Min. Sci., **44:** 98–107.

Gramberg, J. (1965): Axial cleavage fracturing, a significant process in mining and geology. In: Eng. Geol., **1** (1): 31–72.

Gramberg, J. (1989): A non-conventional view on rock mechanics and fracture mechanics. 250 pp. Rotterdam, Brookfield: A. A. Balkema.

Griffith, A. A. (1921): The phenomena of rupture and flow in solids. In: Phil. Trans. R. Soc. Lond., **A221:** 163–198.

Griffith, A. A. (1924): The theory of rupture. In: Proc. 1st Int. Congr. Appl. Mech., Delft : 55–63.

Hodgson, R. A. (1961a): Classification of structures on joint surfaces. In: American Journal of Science, **259:** 493–502.

Hodgson, R. A. (1961b): Regional study of jointing in Comb Ridge-Navajo Mountain area, Arizona and Utah. In: AAPG Bulletin, **45:** 1–38.

Hull, D. (1999): Fractography. Observing, measuring and interpreting fracture surface topography. 366 pp. Cambridge: Cambridge University Press.

Inglis, C. E. (1913): Stresses in a plate due to the presence of cracks and sharp corners. In: Trans. Inst. Nav. Archit. London, **55:** 219–241.

Irwin, G. R. (1958): Fracture. In: Flügge, S. (ed.): Handburch der Physik. Vol. 6. Berlin, New York: Springer : 551–590.

Johnson, A. M. (1970): Physical processes in geology. A method for interpretation of natural phenomena - intrusions in igneous rocks, fractures and folds, flow of debris and ice. 577 pp. San Francisco: Freeman, Cooper & Company.

Kerkhof, F. (1973): Wave fractographic investigations of brittle fracture dynamics. In: Sih, G. C. (ed.): Proceedings of an international conference on dynamic crack propagation. Lehigh University, Bethlehem, Pennsylvania, USA. July 10-12, 1972. Leyden: Noordhoff International Publishing : 3–35.

Kerkhof, F. (1975): Bruchmechanische Analyse von Schadensfällen an Gläsern. In: Glastechn. Ber., **48** (6): 112–124.

Kieslinger, A. (1951): Gesteinskunde für Hochbau und Plastik. Fachkunde für Steinmetzen, Bildhauer, Architekten und Baumeister. 211 pp. Wien: Österreichischer Gewerbeverlag.

Kieslinger, A. (1958): Restspannung und Entspannung im Gestein. In: Geol. u. Bauw., **24** (1): 95–112.

Kranz, R. L. (1979): Crack-crack and crack-pore interactions in stressed granite. In: Int. J. Rock Mech. Min. Sci., **16** (1): 37–47.

Kranz, R. L. (1983): Microcracks in rocks: a review. In: Tectonophysics, **100:** 449–480.

Kulander, B. R.; Barton, C. C. & Dean, S. L. (1979): The application of fractography to core and outcrop fracture investigations. United States Department of Energy, Morgantown Energy Technology Center, METC/SP-79/3. 174 pp.

Kulander, B. R. & Dean, S. L. (1985): Hackle plume geometry and joint propagation dynamics. In: Stephansson, O. (ed.): Fundamentals of rock joints. Proceedings of the international symposium on fundamentals of rock joints. 15-20 September 1985, Björkliden. Luleå: Centek : 85–94.

Kulander, B. R. & Dean, S. L. (1995): Observations on fractography with laboratory experiments for geologists. In: Ameen, M. S. (ed.): Fractography: fracture topography as a tool in fracture mechanics and stress analysis (Geological Society, London, Special Publications, 92): 59–82.

Lutton, R. J. (1969): Systematic mapping of fracture morphology. In: Geological Society of America Bulletin, **80:** 2061–2066.

Lutton, R. J. (1971): Tensile fracture mechanics from fracture surface morphology. In: Clark, C. B. (ed.): Dynamic rock mechanics. 12th US Symposium on rock mechanics : 561–571.

MacDonald, J. & Burton, C. (2003): Collins dictionary geology. 2. Aufl. 480 pp. Glasgow (UK): HarperCollins Publishers.

Mandl, G. (2005): Rock Joints. The mechanical genesis. 221 pp. Berlin, Heidelberg: Springer.

Mecholsky, J. J.; Rice, R. W. & Freiman, S. W. (1974): Prediction of fracture energy and flaw size in glasses from measurements of mirror size. In: J. Am. Ceram. Soc., **57** (10): 440–443.

Mitchell, K. J. (2008): Mechanics of sheeting joints. Paper 147-1. In: Geological Society of America Abstracts with Programs, **40** (6): 150.

Müller, G. & Dahm, T. (2000): Fracture morphology of tensile cracks and rupture velocity. In: Journal of Geophysical Research, **105** (B1): 723–738.

Nemat-Nasser, S. & Horii, H. (1982): Compression-induced nonplanar crack extension with application to splitting, exfoliation, and rockburst. In: Journal of Geophysical Research, **87** (B8): 6805–6821.

Olson, J. E. (1993): Joint pattern development: effects of subcritical crack growth and mechanical crack interaction. In: Journal of Geophysical Research, **98** (B7): 12251–12265.

Parker, J. M. (1942): Regional systematic jointing in slightly deformed sedimentary rocks. In: Geological Society of America Bulletin, **53** (3): 382–408.

Paterson, M. S. & Wong, T. (2005): Experimental rock deformation - the brittle field. 2. Aufl. 347 pp. Berlin, Heidelberg, New York: Springer.

Payne, B. W. & Ball, A. (1976): The determination of crack velocities in anisotropic materials by the analysis of Wallner lines. In: Philosophical Magazine, **34** (5): 917–922.

Pollard, D. D. (2000): Strain and stress: discussion. In: Journal of Structural Geology, **22:** 1359–1367.

Pollard, D. D. & Aydin, A. (1988): Progress in understanding jointing over the past century. In: Geological Society of America Bulletin, **100:** 1181–1204.

Pollard, D. D.; Segall, P. & Delaney, P. T. (1982): Formation and interpretation of dilatant echelon cracks. In: Geological Society of America Bulletin, **93:** 1291–1303.

Poncelet, E. F. (1958): The markings on fracture surfaces. In: Journal of the Society of Glass Technology, **42** (209): 279–288.

Quinn, G. D. (2009): A history of the fractography of brittle materials. Fractography of advanced ceramics III. In: Key Engineering Materials, **409:** 1–16.

Rice, R. W. (1974): Fracture topography of ceramics. In: Fréchette, V. D.; La Course, W. C. & Burdick, V. L. (ed.): Surfaces and interfaces of glass and ceramics. New York: Plenum Press : 439–472.

Rice, R. W. (1984): Ceramic fracture features, observations, mechanisms and uses. In: Mecholsky, J. J. & Powell, S. R. (ed.): Fractography of ceramic and metal failures. Philadelphia: ASTM : 1–51.

Roberts, J. C. (1961): Feather-fracture, and the mechanics of rock-jointing. In: American Journal of Science, **259:** 481–492.

Savalli, L. & Engelder, T. (2005): Mechanisms controlling rupture shape during subcritical growth of joints in layered rocks. In: Geological Society of America Bulletin, **117:** 436–449.

Schultz, R. A. (2000): Growth of geologic fractures into large-strain populations: review of nomenclature, subcritical crack growth, and some implications for rock engineering. In: Int. J. Rock Mech. Min. Sci., **37:** 403–411.

Segall, P. & Pollard, D. D. (1983): Joint formation in granitic rock of the Sierra Nevada. In: Geological Society of America Bulletin, **94:** 563–575.

Sheldon, P. (1912): Some observations and experiments on joint planes. In: The Journal of Geology, **20** (2): 164–183.

Sutter, B. (2008): Kluftmuster und Kluftgenese am Grimselpass. Geologisch-geotechnische Eigenschaften und Tiefenwirkung der Trennflächensysteme. Masterarbeit. Betreut von Prof. Dr. Simon Löw und Dr. Keith Evans. 97 pp. ETH Zürich, Geologisches Institut.

Swanson, C. O. (1927): Notes on stress, strain, and joints. In: The Journal of Geology, **35** (3): 193–223.

Swanson, P. L. (1984): Subcritical crack growth and other time- and environment-dependent behavior in crustal rocks. In: The Journal of Geophysical Research, **89** (6): 4137–4152.

Syme Gash, P. J. (1971): A study of surface features relating to brittle and semibrittle fracture. In: Tectonophysics, **12:** 349–391.

Tapponnier, P. & Brace, W. F. (1976): Development of stress-induced microcracks in Westerly Granite. In: Int. J. Rock Mech. Min. Sci., **13:** 103–112.

van der Pluijm, B. A. & Marshak, S. (2004): Earth structure. An introduction to structural geology and tectonics. 2. Aufl. 656 pp. New York: W. W. Norton & Company.

Vollbrecht, A.; Rust, S. & Weber, K. (1991): Development of microcracks in granites during cooling and uplift: examples from the Variscan basement in NE Bavaria, Germany. In: Journal of Structural Geology, **13** (7): 787–799.

Wallner, H. (1939): Linienstrukturen an Bruchflächen. In: Z. Physik, **114**:368-378.

Wiederhorn, S. M. & Bolz, L. H. (1970): Stress corrosion and static fatigue of glass. In: J. Am. Ceram. Soc., **53** (10): 543–548.

Woodworth, J. B. (1895): Some features of joints. In: Science, **2** (52): 903–904.

Woodworth, J. B. (1896): On the fracture system of joints, with remarks on certain great fractures. In: Proceedings of the Boston Society of Natural History, **27** (9): 163–183.

Ziegler, M. (2008): Geologie ausgewählter Steinbrüche im Hauzenberger Granitmassiv und gesteinstechnisch-petrographische Untersuchungen an einem granitischen Verwitterungsprofil. Unveröff. Diplomarbeit- und kartierung. Betreut von Prof. Kurosch Thuro und Dr. Gerhard Lehrberger. 177 pp. TU München, Lehrstuhl für Ingenieurgeologie.