DESK COPY

2

JOINTS AND SHEAR FRACTURES IN ROCK

Terry Engelder

Department of Geosciences, Pennsylvania State University, USA

2.1. INTRODUCTION

The upper crust contains a variety of brittle structures from microcracks to continent-scale strike–slip fault zones. This paper presents a qualitative review of those brittle phenomena developed during the rupture of intact rock. Fault zones, most of which develop during a long history of repeated rupturing are excluded. The review progresses from the simplest situation, the single crack, through multiple cracking, toward rupture in shear. Such a progression is also based on the number, scale, and relative orientation of the cracks that contribute to joint and shear fracture patterns. Finally, the review moves to the present debate concerning where, within the crust, joints and shear fractures form and the relationship between joint propagation and the distribution of stress within the upper crust.

2.2. HISTORICAL BACKGROUND

By the end of the nineteenth century structural geologists recognized that some breaks in rocks propagated under tensile stresses and others propa-

Fracture Mechanics of Rock
ISBN: 0-12-066265-5

pore pressure on rocks at depth in the same manner as that outlined by Terzaghi (1943) for soils. Secor's (1965) solution was that joints develop at increasingly greater depths in the crust of the earth as the ratio of fluid pressure to overburden weight approaches one. This solution provided the link among the petroleum industry's hydraulic fractures, natural dikes, and joints formed at great depth. Some geological rock mechanics literature refers to joints as extension fractures in recognition that the lithostatic stress at depth is compressive and that tensile stresses develop only as the compressive lithostatic pressures are counterbalanced by larger pore pressures.

Price (1966) presented an analysis for jointing upon uplift associated with removal of overburden. The driving stress for jointing arises from the thermal–elastic contraction of rock upon uplift. Price's (1966) analysis was parallel with Secor's in the sense that both offered an explanation for joint propagation at depth within the crust. With Secor's (1965) and Price's (1966) papers the dilemma posed 83 years before by Crosby (1882) had largely been resolved.

2.3. THE ISOLATED CRACK

In rocks differences exist between microscopic and mesoscopic cracks. On the microscopic scale the ideal crack is a penny-shaped (planar) opening with long dimensions less than the diameter of individual grains within the host rock. These microcracks may be isolated planar discontinuities or may be linked by many jogs and sharp bends, particularly if cracking occurs within grains containing an easily developed cleavage face. An isolated joint (a crack on a mesoscopic scale) is a discontinuity left by a complicated rupture event cutting a large number of grains. The rupture may consist of a single initiation point and either one main arrest or synchronous secondary arrests. On a mesoscopic scale one rupture may appear to have propagated smoothly without stopping but on a microscopic scale the same rupture propagated quickly but in a discontinuous manner following many branches of microcracking.

In the fracture mechanics literature mesoscopic cracks are subdivided into three types based on the orientation of the load relative to the plane of the initial crack (see Atkinson, Chapter 1 of this volume). Mode I loading occurs when the plane of the crack carries no shear tractions whereas during mode II and mode III loading a shear traction parallels the plane of the crack. Evidence from the geologic record suggests that microcracks and joints in rock propagate under mode I loading where the component of maximum effective tensile stress is normal to the plane of the crack. Evidence for mode I loading includes: (1) joints cutting objects such as individual grains, fossils, likely indicators of shear rupture in sandstone (Aydin and Johnson, 1978). Another indicator is the presence of microscopic feather fractures within the host rock (Friedman and Logan, 1970). In contrast, detailed joint surface morphology indicative of mixed-mode loading on the microscopic scale is usually solid evidence for average mode I loading on the mesoscopic scale.

2.4. MICROCRACKS

2.4.1. Open Cracks

Microcracks are those planar discontinuities that are too small to be seen within a hand specimen. Their longest dimension is of the order of one to several grain diameters (about 100 to 1000 microns) with the small dimension of the order of 1 micron. Based on their location within igneous rock, Simmons and Richter (1976) recognize three general classes of microcracks: grain boundary cracks (located at the boundary between grains); intergranular cracks (cracks cutting more than one grain); and intragranular cracks (cracks contained within one grain). Minerals with a well defined cleavage often crack parallel to cleavage to form a type of intragranular crack. In sedimentary rock, particularly those with a clay matrix, the clay often contains many small cracks of both the grain boundary and cleavage type.

The walls of intragranular as well as grain boundary microcracks are often irregular or ragged (Padovani *et al.*, 1982). The irregular or ragged surface morphology implies that in-plane (mode I) crack propagation may not dominate even at the microscopic scale. Commonly the walls of open cracks have embayments filled with foreign material that has been deposited by hydrothermal processes. Natural intragranular microcracks are rarely fresh but presumably when fresh they should resemble experimentally generated microcracks that follow cleavage planes (e.g. a zigzag pattern that follows the rhombohedral faces of quartz (Martin and Durham, 1975)). This zigzag pattern comes about when the average microcrack plane is misoriented from a dominant cleavage plane. The tendency is for the crack to twist and segment or tilt with abrupt cross-cleavage steps and then propagate for short distances within the cleavage planes even though the cleavage plane is subject to mixed-mode loading. This stepping across cleavage is an energy-controlled rather than a stress-controlled twist or tilt as seen on the surface of mesoscopic joints.

Microcracks have two characteristic distributions within the crust. One is a uniform distribution of the type found within igneous rocks where the microcrack density is not related to local structures but rather a pervasive process such as homogeneous cooling. The other type of distribution varies as a function or distance with respect to local structures such as joints, shear minerals suggest that microcracking can occur shortly after solidification of intrusions at depths greater than 10 km (Savage, 1978). Experimental work indicates that microcracks heal rapidly after formation (Smith and Evans, 1984).

2.4.2. Healed microcracks

Hydrothermal processes including pressure solution/reprecipitation act rapidly to fill open microcracks with cement derived from either the host grain or rock matrix. Bubble tracks in quartz grains are common indicators of a filled microcrack where the filling material has a silica content (Tuttle, 1949). The combinations of host and filling material are numerous as documented by Padovani *et al.* (1982) (e.g. iddingsite filling in hypersthene and epidote-like filling in hornblende).

Information on healing of microcracks comes from experiments by Smith and Evans (1984). In the absence of fluid, microcracks show no tendency to heal but when fluid is introduced, the microcracks heal using the cement that is locally derived and transported by diffusion along the crack surface. The healing process proceeds with the formation of cylindrical voids at the crack tip followed by a pinching of the voids to form spherical inclusions. Estimation of thermal activation parameters for crack healing suggest that the cracks will heal rapidly compared to geological time scales at temperatures of 200°C or greater. See also Chapter 4 of this volume.

The rate of crack healing is undoubtedly a function of many parameters and local conditions. Limited information about relative crack-sealing rates in Ries impact crater shows that the rate of filling is highest in quartz, followed by those in plagioclase and amphibole, respectively (Padovani *et al.*, 1978).

2.5. JOINTS: RECORD OF A RUPTURE

Many planar discontinuities, visible on the mesoscopic scale, show opening displacements with no appreciable shear displacement (Bankwitz, 1966). They propagate during one continuous rupture or during a series of interrupted ruptures. These breaks are variously referred to as joints (Badgley, 1965), extension fractures (Griggs and Handin, 1960), veins (Ramsay, 1980), or a crack as in en echelon cracks (Beach, 1975, 1977). The term joint is used here. Joints often occur in evenly spaced sets that may be correlated from outcrop to outcrop over lightly deformed regions as large as several hundred km wide (Nickelsen and Hough, 1967). In other situations joints may have an irregular spacing (Segall and Pollard, 1983).

Joints, particularly those that are not filled, have a distinctive surface morphology called a plumose pattern (e.g. Woodworth, 1896; Lutton, 1971) homogeneous rocks such as granites barbs trace to the point of origin which is likely to be one of the main microcracks or slightly larger fractures commonly distributed throughout the body (e.g. Segall, 1984). In more inhomogeneous rocks, such as sandstones or shales, barbs radiate from either a bedding plane discontinuity or an inclusion within the bed such as a fossil, concretion or clast (e.g. Kulander *et al.*, 1979).

Within bedded shales and siltstone the point of origin may vary from bed to bed. If joints initiate along the bedding surface, they are more likely to originate from irregularities such as ripples or sole marks than from smooth portions. Joints in one bed often initiate from a common feature such as the upper bedding surface (e.g. Bahat and Engelder, 1984). In adjacent beds, fossils or concretions may host the point of origin for most joints. In sedimentary rocks the flaw from which propagation initiates is usually a feature larger than a microcrack within a grain or at a grain boundary. The close association between points of origin and small cracks, irregularities, or inclusions within the rock leaves little doubt that these are the points of microscopic stress concentration where far-field stress is magnified to locally exceed the tensile strength of the rock at the point of stress concentration.

2.5.2. Rupture Propagation

One or more of at least four patterns may be printed on the surface of joints during progress of a rupture from origin to final arrest (Kulander and Dean, 1985). The surface morphology of joints resembles fractured surfaces of glass (Johnson and Holloway, 1966) and much that can be said about the propagation of a rupture in rock is based on studies of the rupture of glass (Bahat, 1979). The main patterns on glass include: a mirror zone, wallner lines, a mist, and twist hackles.

Mirror zone. The mirror zone covers the area immediately adjacent to the point of origin. This region of essentially flat surface represents the area cut by slow but accelerating rupture where the tip stresses have not increased to the point that many bonds oblique to the rupture can be broken (Bahat, 1979; Kulander *et al.*, 1979). The mirror is more commonly seen on cracks in glass, an amorphous material, than on joints in rocks, a more coarse grained polycrystalline material. Within a polycrystalline material like rocks, local inhomogeneities such as grain boundaries and misoriented grains will react with the far-field tensile stress to twist or tilt the crack on a small scale. This strength-controlled tilting or twisting will roughen the crack next to the nucleation point. Kulander *et al.* (1979) suggest that joint surfaces that are without undulations larger than grain size may represent some sort of mirror region. In this case joint propagation would have been relatively slow.

planes oblique to the main crack and their long axes parallel to the direction of the rupture (Poncelet, 1958). The mist divides the mirror zone from the hackle zone, a zone of crack bifurcation and severe surface roughening (Lawn and Wilshaw, 1975).

The hackle zone. In glass, cracks that propagate at a critical velocity tend to branch or bifurcate (Lawn and Wilshaw, 1975). Several models have been proposed to explain crack bifurcation including distortion of the stress field at the crack tip and initiation of secondary cracks. In either case there seems to be evidence for local components of twist during crack propagation. The Yoffe (1951) hypothesis is that crack tips at their terminal velocity shift the maximum local tension away from the existing crack plane and onto an inclined plane. The rupture then follows that inclined plane and branches away from the original crack plane. A second hypothesis for crack branching is that the local stress intensity is so high at the tip of a crack moving at terminal velocity that secondary cracks form. The main crack then branches to follow a number of the secondary cracks. The effect of this crack branching is to form a hackle zone which records crack motion at a critical high velocity. In the ceramics literature the hackle zone refers only to that portion of the crack surface where the crack was moving fast enough to cause crack bifurcation and, hence, hackles signify rapid fracture (Bahat, 1979).

In the geological literature there is some confusion about the use of the term hackle. Kulander *et al.* (1979) states that, "Twist hackles form when a propagating fracture abruptly enters a region of different stress orientation. The fracture breaks (twists) into a series of en echelon individual (blade-like) twist hackle faces, each perpendicular to a resultant tension". These regions of different stress orientation are local in scale and not large enough to cause the rotation of the entire joint plane (Fig. 2.3). Twist hackles will form en echelon within the plane of the original joint and often form as the joint approaches the edge of sedimentary beds (Hodgson, 1961). Kulander *et al.* (1979) suggest that the twist hackle is most pronounced on portions of the

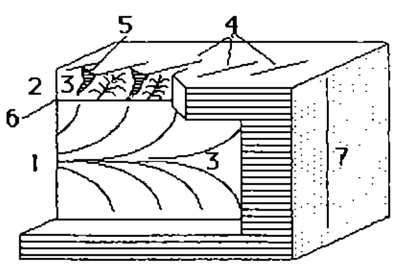


Figure 2.3. The surface morphology on joints (after Hodgson, 1961).

2.6. RECRACKING

Recracking (starting a rupture after it has come to a complete halt) may take three forms: intermittent growth, crack–seal growth, and joint–zone growth. All three of these forms occur on the mesoscopic scale. Some cracks may grow intermittently with time so that recracking is manifested by a single-rupture plane with multiple arrest lines (Bahat and Engelder, 1984). With time a single crack may heal by the introduction of cements including silica and calcite. Recracking as manifested by crack–seal veins occurs when a second rupture follows the original crack either by propagating through the cement or by following the cement-intact rock boundary (Ramsay, 1980). Recracking may also occur with ruptures propagating side-by-side in the form of a zone of intense jointing (Hodgson, 1961). Crack-seal veins and joint zones have many features in common. Recracking is distinguished from multiple cracking which involves cracking in more than one orientation to form a regular pattern within a rock outcrop or over a large region.

Recracking indicates that stress magnitudes change in time. In particular, tensile stresses cycle from low to high causing a single crack to propagate further, a sealed crack to recrack, and multiple cracks in a narrow zone. The nature of stress cycling is not known with certainty but evidence points to changes in fluid pressure along joints as one mechanism for significant effective stress changes within the rock.

Here a clear distinction must be made between pore pressure and fluid pressure in joints. The first should mean only pressure of fluid in pores and it is likely to be different in magnitude from the fluid pressure along a joint. The fluid pressure within a joint acts against the remote stress normal to the joint. This stress is likely to be the least compressive stress. In contrast, the pore pressure acts against all components of normal stress on the scale of the individual pore. It is the pore pressure that creates an effective stress within a rock.

2.6.1. Intermittent growth

Some joints of the Appalachian Plateau USA display a plume pattern that looks rather like a fan which is repeated several times (Bahat and Engelder, 1984). One interpretation is that the edge of the fan is an arrest of a rupture which is further driven forward when the outward pressure of pore fluid on the crack wall exceeds the tensile strength of the rock (Engelder, 1985). This rupture stops within a short distance because of the sudden drop in pore pressure accompanying the expansion of the crack on initiation of the rupture. In specific examples shown in Bahat and Engelder (1984) the crack tip advanced less than a metre during individual events (Fig. 2.1). The effective pressure starts to increase each time the crack tip stops advancing.

2.6.3. Joint Zones

Hodgson (1961) recognized that individual members of a joint set often consist of several closely spaced joints in a joint zone (Fig. 2.7). When the joint zones are traced parallel to strike, single joints terminate and are replaced by others that are slightly offset within the same zone. A joint zone may consist of many individual joints closely enough spaced that when viewed from a distance the joints appear as an individual joint cutting the length of the outcrop. The formation of these joints again requires the repeated cracking of the outcrop. The tips of individual joints pass each

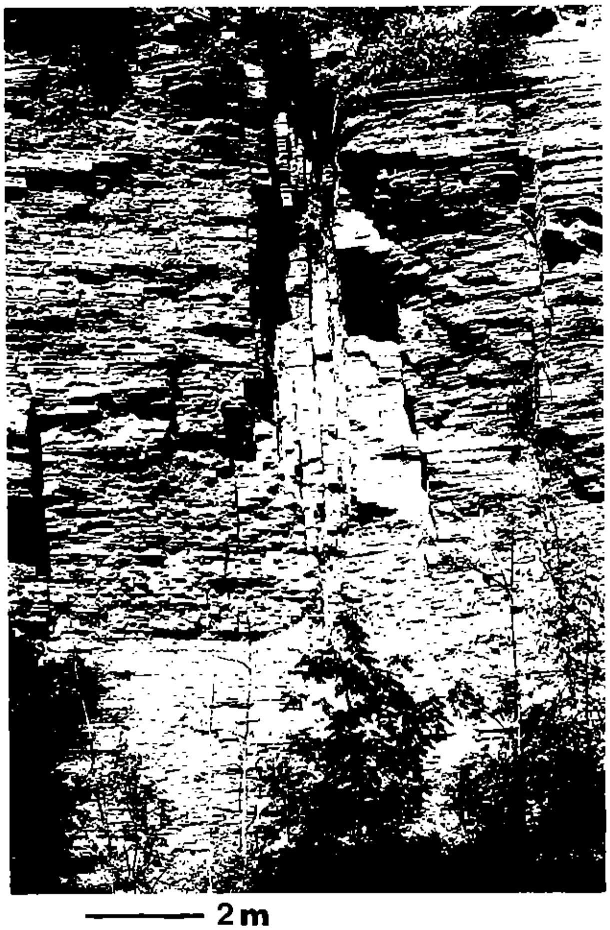


Figure 2.7. Joint zone within Devonian siltstones from the Appalachian Plateau, USA.

between these microfractures and the shear zone points in the direction of shear along the fault zone. Microcracks within the intact rock at an acute angle to a joint are one of the surest signs of a shear origin for the fracture.

Although common, a cloud of microcracks is not always found with the development of a shear fracture. This is particularly true for very porous sandstones and carbonates (Nelson, 1985). In the former case the existing pores probably serve as a sufficiently dense set of microcracks that the generation of more is not necessary to the fracture process.

2.7.1. Braided shear fractures

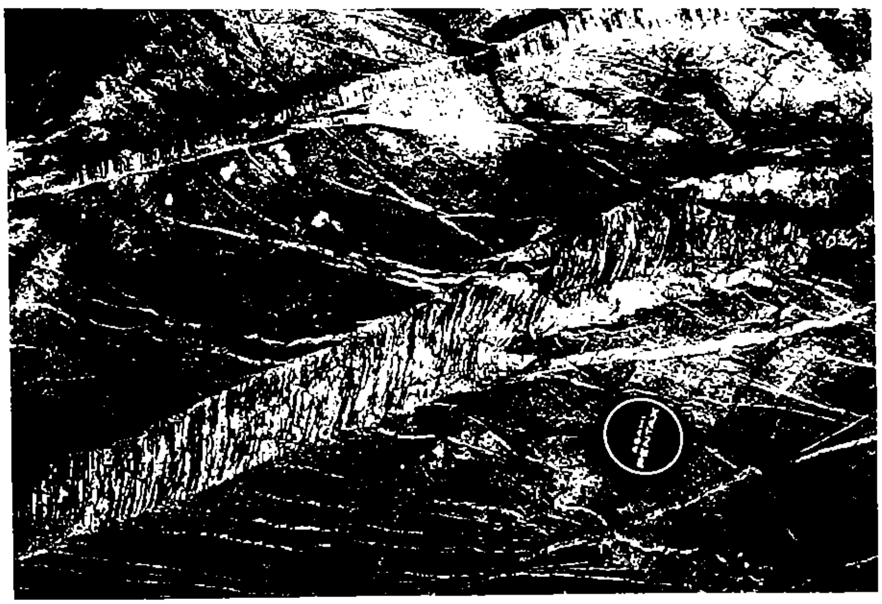
Shear fractures are often graphically developed within lightly cemented sandstone where zones of highly milled gouge interfinger with bits and pieces of intact rock (Fig. 2.8). The small pieces of the intact rock often have a lens

Figure 2.8. Braided shear fractures from the Bonita Fault zone, New Mexico, USA.

2.7.2. Extensional (Hybrid) Shear Fractures

Brittle failure for rock may be predicted by the Coulomb–Mohr failure envelope. Two important characteristics of the Coulomb–Mohr failure envelope are that it slopes toward the tensile portion of the normal stress axis and it becomes parabolic in shape as it crosses the shear stress axis and approaches the normal stress axis (see Price, 1966). At large differential stresses the Coulomb–Mohr failure envelope predicts that conjugate shears will have a dihedral angle of the order of 60° depending on the lithology. A consequence of the slope to the Coulomb–Mohr failure envelope is that shear failure under conditions of decreasing confining pressure requires decreasing differential stresses. A consequence of the parabolic shape is that shear failure will occur along conjugate planes with progressively smaller dihedral angles as the confining pressure decreases. As the conjugate angle becomes smaller the normal stress across the planes decreases to the point where it becomes tensile. Under tensile normal stresses an oblique extension fracture (Dennis, 1972) or hybrid shear fracture (Hancock, 1985) forms with the walls of the discontinuity moving apart.

Evidence for extensional shear is found in veins with fibres growing at angle to the wall of the fracture (Fig. 2.10). Presumably the fibres track the



5 cm

Figure 2.10. Extensional shear joint within a sandstone from Tasmania, Australia. Photo taken by Steve Cox.

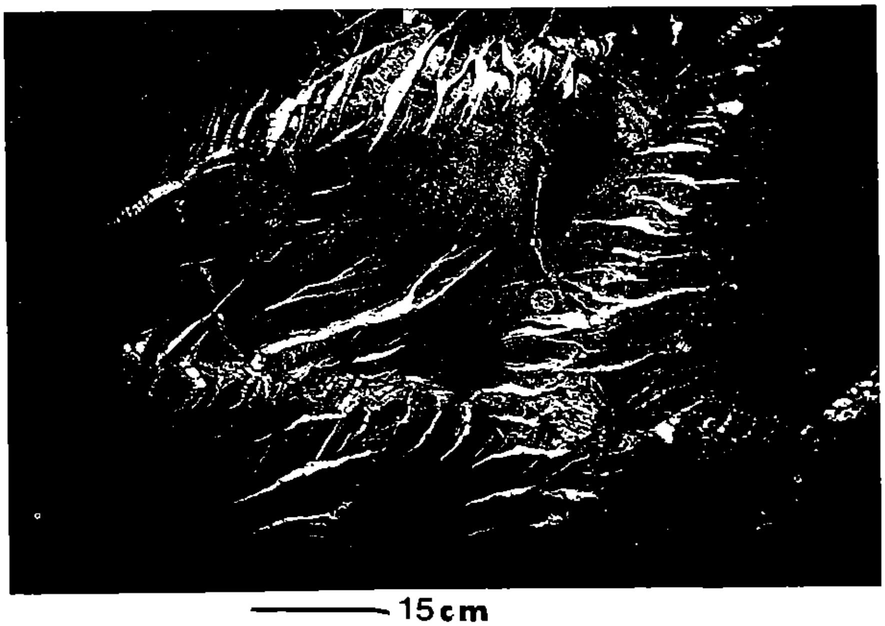


Figure 2.11. Conjugate en echelon vein arrays within the Merrimbula sandstone of Australia (Rickard and Rixon, 1983). This array shows veins in one set parallel to the zone defined by the veins within the other set. The veins in one set are not parallel to veins within the other set.

other shear zone and vice versa (Roering, 1968). Genesis of this pattern is a matter of some debate (Rickard and Rixon, 1983). The other pattern develops with veins in both zones initially parallel (Beach, 1975). The literature is in general agreement that this second pattern is consistent with mode I cracks forming in conjugate zones of shear.

The question concerning the en echelon veins parallel to conjugate zones, is whether the veins formed as mode I cracks or shear cracks. The veins are mode I cracks, if the principal stresses are locally rotated by the shear zone (Lajtai, 1969). In a study of the Merrimbula sandstone of Australia, Rickard and Rixon (1983) argue for mode I crack origin by stating that there is no evidence for shear displacement at vein tips and veins do not continue as shear fractures (Fig. 2.11). Likewise, the orthogonal relationship between veins and local cleavage supports a mode I origin. These observations suggest a local rotation of the principal stresses within the shear zone defined by the en echelon veins (Lajtai, 1969).

Roering (1968) argues that the local stresses cannot be rotated by local shearing because "first-order strains do not, in general, accompany these

Figure 2.13. Pinnate joints within the Cooma complex, Australia. The pinnate joints parallel the pencil and form an acute angle with the master shear joint.

2.8. PATTERNS OF MULTIPLE FRACTURES AND JOINTS

Patterns may be grouped according to structural associations and regional persistence. Here a distinction is made between regional patterns that cut across all local structures and outcrop patterns that are limited in extent to local bodies such as plutons and local structures such as faults and folds.

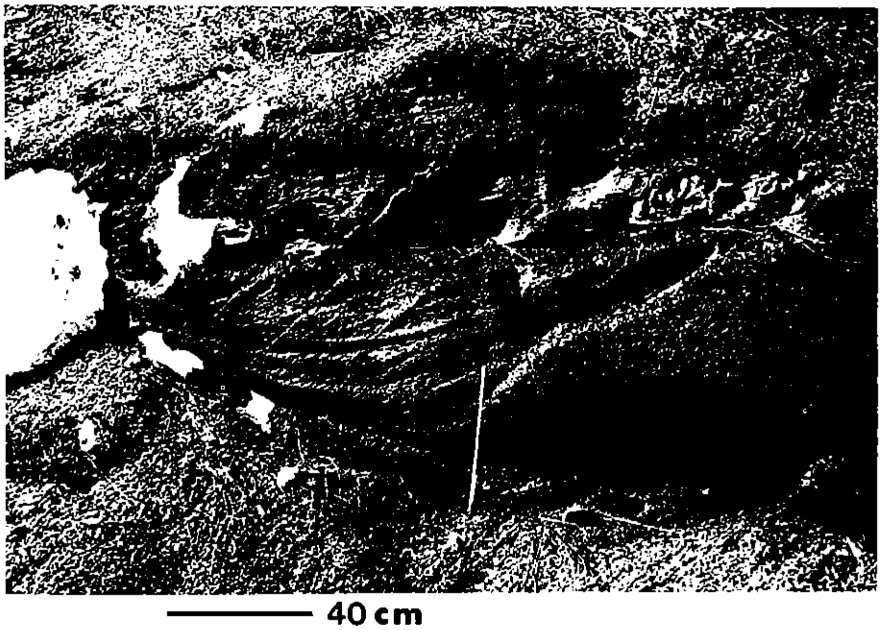


Figure 2.14. Two sets of cooling joints within the Moruya Batholith, Australia. The most prominent set is orthogonal whereas the more pervasive set appears to form Y intersections which are characteristic of columnar jointing. Swiss army knife for scale.

The spacing between the columnar joints and, hence, the cross-sectional area of the columns is believed to be proportional to cooling rates (Spry, 1962) with larger spacing associated with slower cooling rates (Ryan and Sammis, 1978). Cooling rates are inferred from the observation that joint spacing increases toward the interior of igneous dikes in Hawaii (Wentworth and Jones, 1940; Ryan and Sammis, 1978). One explanation for this relationship between joint spacing and depth in an igneous body is that not all thermally induced cracks continue to grow into the interior of a cooling body. Some cracks become stationary with time (Nemat-Nasser *et al.*, 1978).

2.8.2. Outcrop Patterns Associated With Local Structures

Occasionally joint patterns on the outcrop scale appear to be non-systematic in orientation (Fig. 2.15). Most random patterns develop in association with unloading and weathering. Other random jointing develops when local folding and faulting had a complicated history. The effect of such a history was to continuously reorient the local stress field during the propagation of joints. The joint pattern that evolves under these circumstances is difficult or

Folds. Hancock (1985) distinguishes six fracture sets that form about folds (Fig. 2.16). Using an orthogonal coordinate system with ***b*** parallel to the fold axes and *c* normal to bedding Hancock (1985) points out that joints within planes defined by two of the three axes are usually extension fractures. In general shear fractures form oblique to two of the three axes and parallel the third. Assuming that shear fractures form according to a Coulomb–Mohr fracture criterion with the stress system orthogonal to the coordinate system in Fig. 2.16, then there may be up to six possible sets of conjugate shear fractures associated with any fold.

Around the Teton Anticline in central Montana, four of the six possible sets of conjugate shear fractures have formed (Stearns, 1968). Conjugate *hk*0 fractures (Hancock's (1985) terminology) enclosing an acute angle about either ***a*** or ***b*** must form when the intermediate stress is vertical and suggests that *hk*0 about ***a*** forms early and *hk*0 about ***b*** forms only when extension normal to the fold becomes large (i.e. after folding has started). Within the Teton Anticline the conjugate set *hk*0 about ***a*** are the most common. Conjugate sets *h*0*l* about ***a*** and ***c*** form only in response to local buckling and bending. Set *h*0*l* enclosing the acute angle about ***a*** is in the orientation to accompany thrust faulting as the fold grows to become a break thrust (Geiser, 1985). Large-scale folds within the thrust-fold zone of the Variscan externides in southwest Wales (Pembrokeshire) also contain four of the six possible conjugate shear sets (Hancock *et al.*, 1982, 1983). Here set *hk*0 about ***b*** is missing and set 0*kl* about ***c*** is present.

Faults. Fractures forming in the vicinity of a fault can be illustrated using the Bonita fault which has a relatively simple pattern. The Bonita Fault is a well exposed Laramide normal fault which cuts Mesozoic sandstone south of Tucumcari, North Mexico (Stearns, 1972). Throw on the main fault plane is a maximum of 150 m. Faulting occurred at an effective confining pressure which never exceeded 20 MPa based on an estimate of less than 1.5 km of overburden. Associated with the Bonita fault are two antithetic faults with throws of less than 35 m. Shear fractures, of which many form a complete conjugate system, pervade the 200 m wide fault zone which is defined based on the distribution of shear fractures (Fig. 2.17). A regional joint set also cuts across the Bonita fault zone. The average strike of the shear fracture parallels the Bonita Fault and its antithetic fault; the average dip is 60°. Many of the shear fractures are offset up to 5 cm and contain several mm of quartz fault gouge in the form of braided zones.

The Bonita fault is a fine example of the formation of shear fractures in an Andersonian (Anderson, 1951) normal faulting environment. Two conjugate fractures dipping at 60° give a clear indication of a vertical maximum principal stress and horizontal minimum principal stress. Regional strain across the Bonita fault was accommodated with the formation of two shear

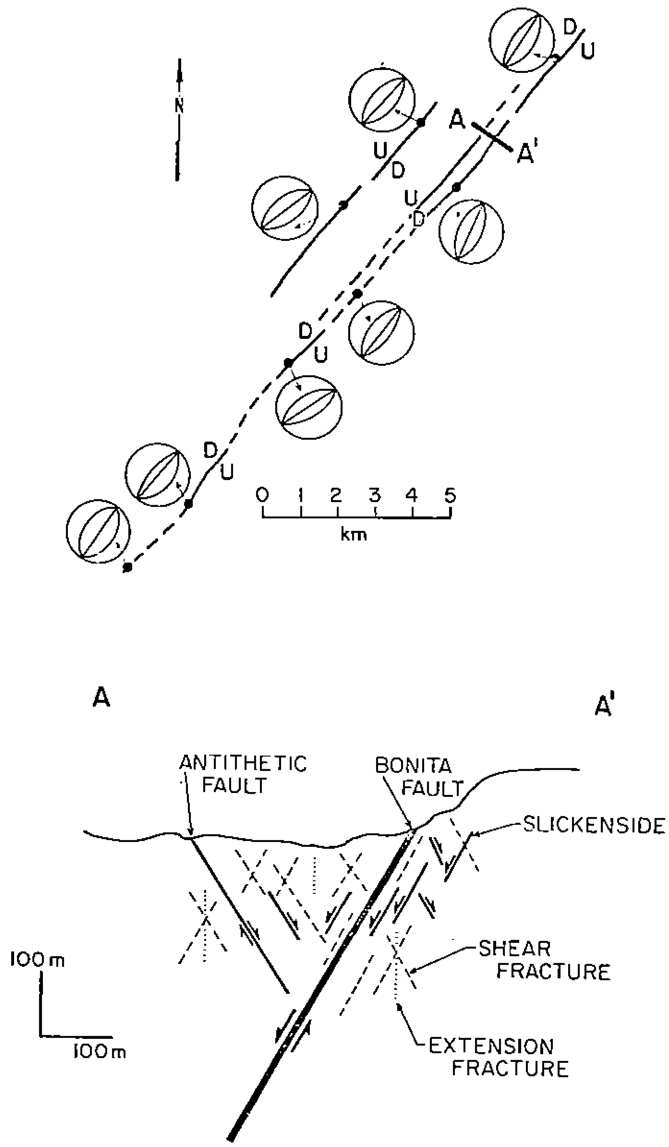


Figure 2.17. Map of the Bonita Fault showing in stereographic projection the orientation, the conjugate shear fractures and extension fractures (after Stearns, 1972). AA′ is a cross-section through the Bonita Fault.

orthogonal. Central Indiana shows such a pattern (Powell, 1976). Three sets appear in central Ohio (Ver Steep, 1942) and on the northern rim of the Michigan Basin four sets appear at about 45° to each other (Holst and Foote, 1981). In New England Wise *et al.* (1979) observe fracture sets that cut across the tectonic grain.

Shear fractures. On a regional scale shear fractures are less common than joints. Systematic shear fractures can form over a region undergoing three-dimensional strain (Reches, 1978). One example of this behaviour is found in central Oregon where Cenozoic normal faults form in a rhomboid pattern (Donath, 1962). Thompson and Burk (1974) describe the same pattern for fault patterns of the entire Basin and Range.

A system of orthogonal shear faults cut the strata of northern Umbrian Apennines fold belt (Marshak *et al.*, 1982). The fault surfaces are associated with stylolites and are coated with elongate calcite fibres, suggesting that movement occurred by the mechanism of pressure solution slip. Here the deformation mechanism is independent of normal stress and so the plane of highest shear stress is most easily activated in slip. Because that plane is at 45° to the maximum compressive stress, the shear fractures in the Apennines have an orthogonal pattern on the regional scale.

Continental scale fractures and joints. Wise (1974) has used ERTS images to show that pervasive fracture systems extend over much of the northeastern United States. The most predominant set strikes N70°E and correlates with outcrop scale joints that appear over much of the northeastern United States.

2.9. LOADING CONDITIONS LEADING TO PROPAGATION OF JOINTS

Three basic loading conditions may lead to brittle failure within the crust of the earth. One loading condition leads to shear failure whereas the other two lead to mode I crack propagation. Shear failure is associated with high shear stresses generated by stress magnification around mesoscopic to macroscopic scale structures. This loading condition has no depth restriction. The generation of abnormal fluid pressures leads to effective tensile stresses at depth and subsequent propagation of joints. Thermal–elastic contraction during erosion and uplift is another mechanism for the generation of effective tensile stresses and the propagation of joints. These two loading conditions for joint propagation are more likely to be restricted in depth with the former deep and the latter shallow.

joints in that tectonic compaction or a tectonically produced hydraulic head is responsible for the abnormal pore pressure leading to joint propagation. Hydraulic joints propagate in response to high pore pressures generated by compaction accompanying overburden loading. Jointing by the former mechanism may occur during a tectonic event such as the Alleghanian Orogeny (~320 Ma to 280 Ma) of the Appalachian Mountains whereas jointing by the latter mechanism may occur in a Gulf of Mexico setting.

Although subtle, the distinction between unloading and release joints is based on the mechanism controlling the joint's orientation. Both types form in response to the thermal–elastic contraction developed during erosion and uplift. The orientation of unloading joints is controlled by either a residual or contemporary tectonic stress with propagation normal to the least principal stress. In either case the actual stress differences are less than four times the tensile strength of the rock. Release joints are controlled by a rock fabric not including a residual stress. On the Appalachian Plateau of New York strike joints open parallel to a cleavage within the clastic rocks of the Appalachian Basin. In a sense these joints are responding to the release of the Alleghanian stress field under which the cleavage formed.

2.9.3. Fracture Orientation versus Depth: A case for Unloading Joints

The fracture process is illustrated by considering the depth at which unloading joints are found (Engelder, 1985). The orientation of unloading joints is controlled by either the tectonic stress field at the time of denudation and uplift or a residual stress left from some previous tectonic event. A set of joints (set III of Parker, 1942) on the Appalachian Plateau is aligned with the contemporary tectonic stress field and because of this relationship Engelder (1982a) proposed that the joints were genetically related to the contemporary tectonic stress field. These set III joints are related to neither a residual stress nor a structural fabric left by the Alleghanian tectonic compression.

The distribution of natural fractures within core from the Appalachian Basin supports the hypothesis that unloading joints have developed. From 1975 through 1981 selected wells were core-drilled in the northern end of the Appalachian Basin. The cores were oriented and later examined for natural joints. The location of these wells is plotted on a map showing that joints at depth include some that correlate with cross-fold joints and some that correlate with the ENE direction of the maximum horizontal compression in the contemporary tectonic stress field (Fig. 2.19). The most common joint within the top 0.5 km strikes subparallel to the contemporary tectonic stress field whereas those observed below the top 0.5 km are almost exclusively cross-fold joints. Joints within the core from OH-1 are the only joints that do not fit into the scheme of unloading joints at <0.5 km depth. The argument

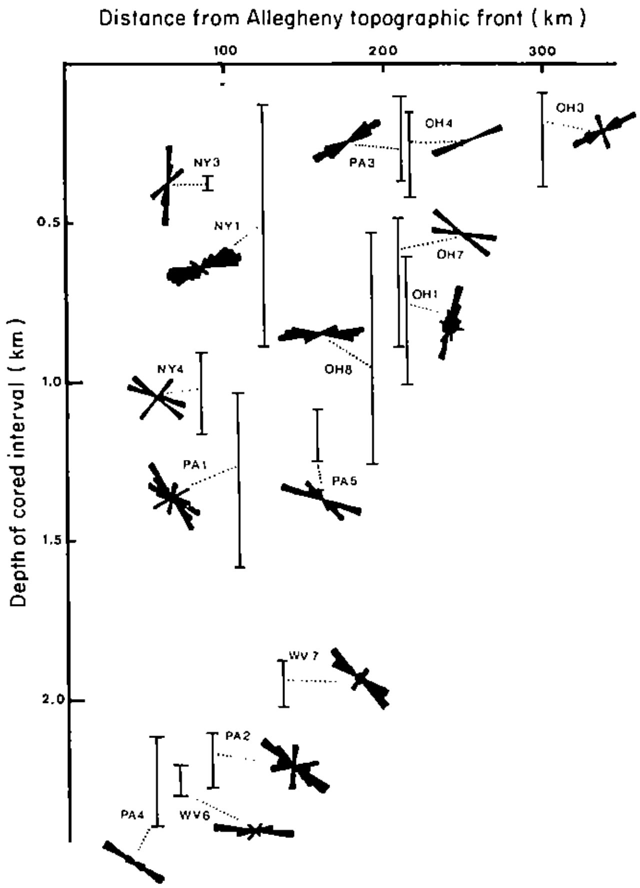


Figure 2.19. A plot of depth of cored interval versus distance from the Allegheny topographic front. As in Fig. 2.18 the strike of the natural joints is plotted in map view with north toward the top of the figure. Here it is seen that the deep cores contain cross-fold joints whereas shallow cores contain joints parallel to the contemporary tectonic stress field. An interpretation is that deep joints are tectonic whereas the shallow joints are unloading.

depth, the zone of effective tensile stresses extends down to about 400 m. The orientation of the least principal stress is about 5° west of north. These tensile stresses would be relieved by the propagation of unloading joints oriented slightly north of east. Although stress data from this well do not come from the Devonian sequence, the depth at which tensile stresses develop is within the depth range of Devonian core containing unloading joints striking parallel to the contemporary tectonic stress field.

This denudation is compatible with the propagation of unloading joints at a few hundred metres to 1 km in depth. Data from both Hickman *et al.* (1984) and Cliffs Minerals (1982) suggest that the depth of propagation of unloading joints is between 200 and 500 m. These observations and this interpretation further reinforce the proposal that thermal–elastic contraction during erosion and uplift is the third general process leading the fracturing within the crust of the earth.

2.10. CONCLUSIONS

Joints are those breaks in the upper crust that display opening displacements and no appreciable shear displacement. Shear fractures are breaks with appreciable displacement parallel to the plane of the break. The formation of both joints and shear fractures starts on the scale of microcracks which may be open or healed. The surface morphology of joints records a rupture event including nucleation, propagation, and arrest. Joints recrack in the form of crack–seal veins, joint zones, or intermittent growth of a single joint. The process of shear fracturing is reflected in braided shear fractures, extensional shear fractures, en echelon cracks, and pinnate joints next to the shear fracture. Patterns of multiple fractures and joints include contraction cracks, fractures around local structures, and regional patterns. Jointing and shear fracturing occur under different stress conditions within the crust.

ACKNOWLEDGEMENTS

Numerous geologists have introduced me to various aspects of the fracture process. These people include: Dov Bahat and Bryon Kulander (joint surface morphology), Steve Cox and Paul Hancock (hybrid shear fractures), Mel Friedman and Bob Kranz (microcracks), Chris Powell (veins and kink bands associated with braided shear fractures). Barry Atkinson, Nigel Higgs, Ron Nelson, David Pollard, and Don Wise are all thanked for comments on this manuscript. This work was supported by NSF grant EAR-83-06146, a grant from Schlumberger-Doll Research Center, Ridgefield, Connecticut, and DOE contracts DE-AC21-83MC20337 and DE-AS02-76ER04054.

REFERENCES

Adams, F. D. (1910). *J. Geol.* **18**, 489–525.

Anderson, E. M. (1951). "Dynamics of Faulting and Dyke Formation with Application to Britain". Oliver and Boyd, Edinburgh, 1–206.

Atkinson, B. K. (1984). *J. Geophys. Res.* **89**, 4077–4114.

Aydin, A. and Johnson, A. M. (1978). *Pure Appl. Geophys.* **116**, 931–942.

Griggs, D. T. and Handin, J. (eds) (1960). *Geol. Soc. Am. Mem.* **79**, 1–352.
Grout, M. A. and Verbeek, E. R. (1983). Proc. 16th Oil Shale Symposium, (ed. Gray, J. H.). Colorado School of Mines, Golden, Colorado, 68–80.
Hancock, P. L. (1972). *Geol. Mag.* **109**, 269–276.
Hancock, P. L. (1985). *J. Struct. Geol.* **7**, 437–458.
Hancock, P. L., Dunne, W. M. and Tringham, M. E. (1982). *In* "Geological Excursions in Dyfed, South-West Wales", (ed. Bassett, M. G.). National Museum of Wales, Cardiff, 214–248.
Hancock, P. L., Dunne, W. M. and Tringham, M. E. (1983). *In* "The Variscan Fold Belt in the British Isles", (ed. Hancock, P. L.). Adam Hilger, Bristol, 47–73.
Harrison, J. R. and Moench, R. E. (1961). *U.S.G.S. Prof. Pap.* **374-B**.
Hickman, S. H., Healy, J. H. and Zoback, M. D. (1985). *J. Geophys. Res.* **90**, 5497–5512.
Hodgson, R. A. (1961). *Am. J. Sci.* **259**, 493–502.
Holst, T. B. and Foote, G. R. (1981). *Geol. Soc. Am. Bull.* **92**, 85–93.
Hubbert, M. K. and Rubey, W. W. (1959). *Geol. Soc. Am. Bull.* **70**, 115–166.
Hubbert, M. K. and Willis, D. G. (1957). *Trans. AIME* **210**, 153–168.
Huber, N. K. and Rinehart, C. D. (1967). *U.S.G.S. Prof. Paper* **554-D**, 1–21.
Johnson, J. W. and Holloway, D. G. (1966). *Phil. Mag.* **14**, 731–743.
Kranz, R. L. (1979). *Int. J. Rock Mech. Min. Sci.* **16**, 37–47.
Kranz, R. L. (1983). *Tectonophys.* **100**, 449–480.
Kulander, B. R. and Dean, S. L. (1985). *In* "Fundamentals of Rock Joints", (ed. Stephansson, O.). Cevek Publishers, Sweden, 85–94.
Kulander, B. R., Barton, C. C. and Dean, S. L. (1979). "The Application of Fractography to Core and Outcrop Fracture Investigations": Tech. Report for U.S. Dept. of Energy Contract EY-77-Y-21-1391, METC/SP-79/3, 174.
Lajtai, E. Z. (1969). *Geol. Soc. Am. Bull.* **80**, 2253–2272.
Lajtai, E. Z. (1971). *Tectonophys.* **11**, 129–156.
Lachenbruch, A. H. (1962). *Geol. Soc. Am. Sp. paper* **70**, 1–69.
Lawn, B. R. and Wilshaw, T. R. (1975). "Fracture of Brittle Solids". Cambridge University Press, Cambridge.
LeConte, J. (1882). *Am. J. Sci.* Ser. 3, **23**, 233–234.
Leith, C. K. (1913). "Structural Geology". Holt, New York.
Lockner, D. and Byerlee, J. (1977). *Bull. Seism. Soc. Am.* **67**, 247–258.
Lutton, R. J. (1971). *In* "Dynamic Rock Mechanics", Proc. 12th Sym. Rock. Mech. (ed. Claye, G. B.). Port City Press Inc., Baltimore, Md, 561–571.
Magara, K. (1975). *Bull. Am. Ass. Petrol. Geol.* **59**, 2017–2045.
Magara, K. (1981). *Bull. Am. Ass. Petrol. Geol.* **65**, 123–132.
Marshak, S., Geiser, P. A., Alvarez, W. and Engelder, T. (1982). *Geol. Soc. Am. Bull.* **93**, 1013–1022.
Martin, R. J. and Durham, W. B. (1975). *J. Geophys. Res.* **80**, 4837–4844.
McGarr, A. (1980). *J. Geophys. Res.* **85**, 6231–6238.
Murgatroyd, J. B. (1942). *J. Soc. Glass Tech.* **26**, 155–171.
Narr, W. and Currie, J. B. (1982). *Bull. Am. Ass. Petrol. Geol.* **66**, 1231–1247.
Nelson, R. (1985). "Contributions in Petroleum Geology and Engineering" No. 1. Gulf, Houston, 320 pp.
Nemat-Nasser, S., Keer, L. M. and Parihar, K. S. (1978). *Int. J. Solids Structures* **14**, 409.
Nickelsen, R. P. (1976). *Utah Geol. Assoc.* **5**, 193–199.
Nickelsen, R. P. and Hough, V. D. (1967). *Bull. Geol. Soc. Am.* **78**, 609–630.
Oertel, G. (1965). *Tectonophys.* **2**, 343–393.
Olsson, W. and Peng, S. 1976. *Int. J. Rock Mech. Min. Sci.* **18**, 53–59.

Vernon, R. H., Etheridge, M. A. and Wall, V. J. (1983). *Geol. Soc. Austral. Abstr.* **9** (Canberra Convention).
Voight, B. and St. Pierre, B. H. P. (1974). *In* "Advances in Rock Mechanics". Proc. 3rd Cong. ISRM, Vol. II, 580–582.
von Karmon, T. (1911). *Z. Ver. Deutsch. Ingeniere* **55**, 1749–1757.
Walker, F. (1940). *Geol. Soc. Am. Bull.* **51**, 1059–1106.
Wang, H. F. and Heard, H. C. (1985). *J. Geophys. Res.* **90**, 10,342–10,350.
Wang, H. F. and Simmons, G. (1978). *J. Geophys. Res.* **83**, 5849–5856.
Weaire, D. and O'Carrol, C. (1983). *Nature* **302**, 240–241.
Wentworth, C. K. and Jones, A. E. (1940). *J. Geol.* **48**, 975–1006.
Wise, D. U. (1964). *Geol. Soc. Am. Bull.* **75**, 287–306.
Wise, D. U. (1974). *Utah Geological Association Publication* **5**, 416–422.
Wise, D. U. and students (1979). "Fault, Fracture and Lineament Data from Western Massachusetts and Western Connecticut": Final Report to the Nuclear Regulatory Commission for Contract No. AT(49-24-0291).
Woodworth, J. B. (1896). *Proc. Boston Soc. Nat. Hist.* **27**, 163.
Yoffe, E. H. (1951). *Phil. Mag.* **42**, 739.

Padovani, E. R., Shirey, S. B. and Simmons, G. (1982). *J. Geophys. Res.* **87**, 8605–8630.

Parker, J. M. (1942). *Bull. Geol. Soc. Am.* **53**, 381–408.

Plumb, R., Engelder, T. and Sbar, M. (1984). *J. Geophys. Res.* **89**, 9333–9349.

Pollard, D. D., Muller, O. H. and Dockstader, D. R. (1975). *Geol. Soc. Am. Bull.* **93**, 1291–1303.

Pollard, D. D., Segall, P. and Delany, P. T. (1982). *Geol. Soc. Am. Bull.* **93**, 1291–1303.

Poncelet, E. F. (1958). *J. Soc. Glass Tech.* **42**, 279–288.

Powell, R. L. (1976). PhD Thesis. Purdue University, 1–169.

Preston, D. A. (1968). *Trans. AGU* **49**, 302.

Preston, D. A. (1926). *J. Soc. Glass Tech.* **10**, 234–269.

Price, N. J. (1968). "Fault and Joint Development in Brittle and Semi-brittle Rock". Pergamon, London, 1–176.

Price, N. J. (1974). *In* "Advances in Rock Mechanics", Proc. 3rd Cong. ISRM, Vol. 1, 487–496.

Ramsay, J. G. (1980). *Nature* **284**, 135–139.

Reches, Z. (1978). *Tectonophys.* **47**, 109–129.

Rickard, M. J. and Rixon, L. K. (1983). *J. Struct. Geol.* **5**, 573–578.

Roberts, J. C. (1961). *Am. J. Sci.* **259**, 481–492.

Rispoli, R. (1981). *Tectonophys.* **75**, T29–T36.

Robin, P.-Y. and Currie, J. B. 1971. *Can. J. Earth Sci.* **8**, 1302–1313.

Robinson, C. S. (1956). *USGS Bull.* **1021-I**, 289–302.

Roering, C. (1968). *Tectonophys.* **5**, 107–123.

Ryan, M. P. and Sammis, C. G. (1978). *Geol. Soc. Am. Bull.* **89**, 1295–1308.

Savage, W. Z. (1978). *Geophys. Res. Lett.* **5**, 633–636.

Scholz, C. H. (1968). *J. Geophys. Res.* **73**, 3295–3302.

Segall, P. (1984). *Bull. Geol. Soc. Am.* **95**, 454–462.

Segall, P. and Pollard, D. D. (1983). *Bull. Geol. Soc. Am.* **94**, 563–575.

Secor, D. T. (1965). *Am. J. Sci.* **263**, 633–646.

Secor, D. T. (1969). *Proc. Conf. Res. Tectonics, Geol. Surv. Canada* **68–52**, 3–48.

Shainin, V. E. (1950). *Geol. Soc. Am. Bull.* **61**, 509–517.

Simmons, G. and Nur, A. (1968). *Science* **162**, 789–791.

Simmons, G. and Richter, D. (1976). *In* "The Physics and Chemistry of Minerals and Rocks", (ed. Strens, R. G. J.). Wiley, New York.

Smith, D. L. and Evans, B. (1984). *J. Geophys. Res.* **89**, 4125–4136.

Spry, A. H. (1962). *J. Geol. Soc. Austral.* **8**, 191–216.

Stearns, D. W. (1965). *In* Rock Mechanics Seminar Special Report, (ed. Riecker, R. E.). Terrestrial Sciences Lab. Air Force Cambridge Research Labs, Bedford, MA, 97–116.

Stearns, D. W. (1972). *New Mexico Geological Society Guidebook* **23**, 161–164.

Swolfs, H. A., Handin, J. and Pratt, H. R. (1974). *In* "Advances in Rock Mechanics", Proc. 3rd Cong. ISRM, Vol. II, 563–568.

Terzaghi, K. (1943). "Theoretical Soil Mechanics". Wiley, New York.

Thompson, G. A. and Burke, D. B. (1974). *Ann. Rev. Earth Planet. Sci.* **2**, 213–238.

Tuttle, O. F. (1949). *J. Geol.* **57**, 331–356.

Van Hise, C. R. (1896). 16th Ann. Rept. U.S. Geol. Surv. U.S. Govt Printing Office, Washington.

Van Tyne, A. (1948). MS Thesis. Syracuse University, Syracuse, NY.

Ver Steeg, K. (1942). *Econ. Geol.* **37**, 503–509.

Ver Steeg, K. (1944). *J. Geol.* **52**, 131–138.

Badgley, P. C. (1965). "Structural and Tectonic Principles". Harper and Row, New York.
Bahat, D. (1979). *Geol. Mag.* **116**, 81–92.
Bahat, D. (1980). *Tectonophys.* **68**, T1–T7.
Bahat, D. and Engelder, T. (1984). *Tectonophys.* **104**, 299–313.
Bankwitz, P. (1966). *Geologie* **15**, 896–941.
Barker, C. (1972). *Bull. Am. Ass. Pet. Geol.* **56**, 2068–2071.
Bauer, S. J. and Johnson, B. (1979). Proc. 20th U.S. Symp. Rock Mech. Austin, Texas, 7–18.
Beach, A. (1975). *Tectonophys.* **28**, 245–264.
Beach, A. (1977). *Tectonophys.* **40**, 201–225.
Becker, G. F. (1893). *Bull. Geol. Soc. Am.* **4**, 13–90.
Brace, W. F. and Bombolakis, E. G. (1963). *J. Geophys. Res.* **68**, 3709–3713.
Brown, A. (1973). *Tectonophys.* **19**, 383–397.
Bucher, W. H. (1920). *J. Geol.* **28**, 707–730.
Carter, N. L. and Kirby, S. H. 1978. *Pure Appl. Geophys.* **116**, 807–839.
Cliffs Minerals, Inc. (1982). "Analysis of the Devonian Shales in the Appalachian Basin". U.S. DOE contract DE-AS21-80MC14693, Final Report. Springfield Clearing House, Washington.
Conrad, R. E. and Friedman, M. (1976). *Tectonophys.* **33**, 187–198.
Crosby, W. O. (1882). *Proc. Boston Soc. Nat. Hist.* **XXII**, 72–85.
Dale, T. N. (1923). *U.S.G.S. Bull.* **738**, 1–488.
Daubree, A. (1879). "Géologie Experimentale". Libraire des Corps des points et Chausser des Mines et des Télégraphs, Paris.
Dennis, J. G. (1972). "Structural Geology". Ronald Press, New York.
Delaney, P. T., Pollard, D. D., Ziony, J. I., and McKee, E. H. (1986). *J. Geophys. Res.* **91**, 4920–4938.
Donath, F. A. (1962). *Geol. Soc. Am. Bull.* **73**, 1–16.
Durney, D. W. and Ramsay, J. G. (1973). "Gravity and Tectonics", (eds de Jong, K. A. and Scholten, R.). Wiley, New York, 67–96.
Dyer, R. 1981. *Geol. Soc. Am. Abstracts with Programs* **13**, 444.
Eisbacher, G. H. and Bielenstein, H. U. (1971). *J. Geophys. Res.* **76**, 2012–2021.
Engelder, T. (1974). *Geol. Soc. Am. Bull.* **85**, 1515–1522.
Engelder, T. (1982a). *Tectonics* **1**, 161–177.
Engelder, T. (1982b). *Tectonics* **1**, 465–470.
Engelder, T. (1985). *J. Struct. Geol.* **7**, 459–476.
Engelder, T. and Geiser, P. (1980). *J. Geophys. Res.* **85**, 6319–6341.
Engelder, T. and Sbar, M. L. (1977). *Pure Appl. Geophys.* **115**, 41–55.
Etheridge, M. A. (1983). *Geology* **11**, 231–234.
Friedman, M. (1964). *In* "State of Stress in the Earth's Crust", (ed. Judd, W. R.). Elsevier, New York, 451–552.
Friedman, M. and Logan, J. M. (1970). *Geol. Soc. Am. Bull.* **81**, 3417–3420.
Friedman, M., Handin, J. and Alani, G. (1972). *Int. J. Rock Mech. Min. Sci.* **9**, 757–766.
Gallagher, J. J., Friedman, M., Handin, J. and Sowers, G. (1974). *Tectonophys.* **21**, 203–247.
Gash, P. J. S. (1971). *Tectonophys.* **12**, 349–391.
Geiser, P. A. (1985). *Am. Ass. Pet. Geol. Bull.* (submitted).
Gilbert, G. K. (1882). *Am. J. Sci.* **23**, Ser. 3, 25–27.
Gray, N. H., Anderson, J. B., Devine, J. D. and Kwasnik, J. M. (1976). *J. Int. Assoc. Math. Geol.* **8**, 617–626.
Griffith, A. A. (1920). *Phil. Trans. R. Soc. Lond.* **A221**, 163–198.

2.9.4. The Consequence of Jointing: Strain Relaxation

Griffith (1920) treated the problem of prediction of crack growth by considering the balance between available strain energy and the new surface energy that would be created upon crack propagation. This is the phenomenon of strain relaxation during crack propagation. Abundant laboratory data confirms the validity of the Griffith relationship even on the scale of large cracks (cf. Atkinson, 1984). Although there is little reason to doubt that the same behaviour occurs in nature, confirmation in the field is rare.

Strain relaxation associated with jointed rock was measured in outcrops of the basal Cambrian Potsdam Sandstone of New York State (Engelder and Sbar, 1977). The magnitude of strain relaxation was proportional to the joint density with the most heavily jointed rocks containing the least elastic strain. This behaviour may be interpreted as a gradual decoupling of the outcrop from the tectonic stress field either during or after propagation of the joints. The decoupling may take two forms: (1) joints act as very soft springs preventing reloading of the outcrops; and (2) joints have relieved a strain imparted as either a residual elastic strain or an elastic strain from the presence of the contemporary tectonic stress field. If decoupling is the sole cause of low strain within highly fractured outcrops, then it must be assumed that the contemporary tectonic stress field is not capable of reloading the outcrops.

Outcrops containing a dominant fracture set commonly contain an *in situ* strain that is maximum parallel to the strike of the fracture set (Preston, 1968; Eisbacher and Bielenstein, 1971; Brown, 1973; Swolfs *et al.*, 1974; Engelder and Sbar, 1977; Engelder and Geiser, 1980). Again there are two points of view concerning this behaviour. One is that the joints relieve strain normal to their surfaces and then act as very soft springs that prevent the reloading normal to their surfaces. The other point of view is that the driving stress controlling the orientation of the joints was not relieved during joint propagation and is still present. The driving stress may be either a residual stress or a contemporary tectonic stress. A complex relationship between joints and stress is found within the central Adirondack Mountains where the strike of joints and topographic trends align with the contemporary tectonic stress field (Plumb *et al.*, 1984). Although *in situ* stress measurements detect a maximum compression parallel to the joints, the technique is incapable of distinguishing between stresses caused by the tectonic stress field and stresses influenced by near surface jointing.

Unloading joints require the removal of overburden equal to more than 50% of their depth of burial depending on the change in Poisson's ratio during lithification (Engelder, 1985). Estimates for the removal of overburden from the Appalachian Plateau of western New York vary from 500 m (Van Tyne, 1948) to 2 km (conodont isograd index of Epstein *et al.*, 1975).

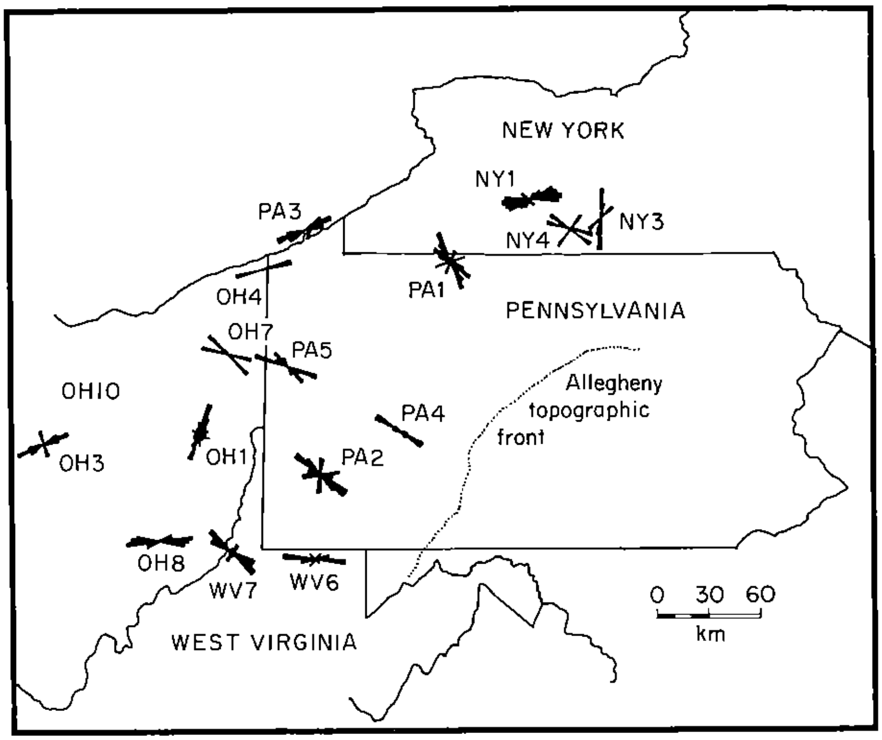


Figure 2.18. The location of fifteen wells drilled by the Department of Energy for their eastern gas shales program. The rose diagrams show the strike of natural joints found within core taken from each of the fifteen wells. Comparing the strike of joints at depth with the orientation of cross-fold joints (not shown), it is apparent that most joints from wells near the Allegheny Topographic Front are in the cross-strike orientation whereas those joints from wells most distant from the Allegheny Topographic Front are parallel to the contemporary tectonic stress field which has a maximum horizontal stress oriented ENE (Engelder, 1982a). This apparent geographic distribution is a function of depth from which the cores were taken as shown in Fig. 2.19.

is that rock at depths in excess of 0.5 km has not been denudated enough to develop the thermal–elastic stresses that favour the propagation of unloading joints. The deeper rocks are also those whose pore space is most likely to house the high pore pressures necessary for propagation of tectonic joints.

The deepest set of stress measurements from the Appalachian Basin was taken within a 1.6 km deep well at Auburn, New York (Hickman *et al.*, 1984). An extrapolation of the magnitude of the least principal stress towards the surface suggests that tensile stresses might develop at about 200 m depth. If normal pore–fluid pressure extends from the surface to 400 m

2.9.1. Stress Conditions

Both shear fractures and joints propagate under a variety of tectonic settings. However, the Coulomb–Mohr failure criterion states that they cannot propagate simultaneously. The stress difference necessary for initiation and propagation of shear fractures is significantly larger than for joints. The Coulomb–Mohr failure envelope predicts that unless the differential stress exceeds four times the tensile strength of the rock shear fracture will not occur (Etheridge, 1983). If the differential stress is less than four times the tensile strength, jointing or at best hybrid shear failure are the only brittle failure mechanisms.

Estimates of differential stress necessary for shear fracturing are based on laboratory tests. The magnitude of differential stress for shear fracture is much higher than the average regional stress differences within the crust of the earth as measured using *in situ* stress techniques (McGarr, 1980). Reactivation of joints in shear are not to be mistaken for a pattern of regional shear fractures. In fact, it is this very reactivation of joints that keeps regional differential stress at the level of the frictional strength of rocks.

High differential stresses are not common on a regional basis, so shear fractures usually do not develop as part of a regional pattern. Because shear fracturing is found within the crust, it follows that local processes such as stress magnification around mesoscopic to macroscopic scale structures must be responsible for higher than average differential stresses. Examples of local stress magnification include asperities along fault zones, structures developed over salt domes, and granite uplifts. Stresses may be magnified within certain portions of folds. However, shear fracturing on a regional basis may develop in the vicinity of extensional environments (e.g. the Basin and Range normal faulting) where shear fracturing does not require the magnification of a regional compressive stress but rather a sufficient decrease in the least horizontal stress to cause significant stress differences which are uniform on a regional basis.

2.9.2. Loading Paths to Joint Propagation

Joints develop on a regional scale (Ver Steeg, 1942; Nickelsen and Hough, 1967). Engelder (1985) has classified four types of joints based on the timing of joint propagation during burial, lithification, deformation, and denudation of clastic rocks within the sedimentary Appalachian Basin of North America. These joint types include tectonic, hydraulic, unloading, and release. Tectonic and hydraulic joints form at depth prior to uplift in response to abnormal pore pressures, whereas unloading and release joints form in the near surface in response to thermal–elastic contraction accompanying erosion and uplift. Tectonic joints are distinguished from hydraulic fractures (i.e. a conjugate set). This is an example of regional plane strain where two sets of shear fractures are sufficient for local deformation (Reches, 1978).

2.8.3. Regional Joint Patterns Associated With a Tectonic Grain

Sedimentary rocks subject to a major tectonic event have several joint sets that relate to burial, regional deformation, and uplift. At least five regional joint sets appear in the northwestern portion of the Appalachian Plateau (Engelder and Geiser, 1980; Engelder, 1985). In tracing the Appalachian fold belt around a major change in strike as many as a dozen joint sets appear in various rock types (Nickelsen and Hough, 1967). In the Piciance Basin of western Colorado multiple joint sets extend across the Basin (Grout and Verbeek, 1983). Subparallel joint sets appear within the Arches National Monument of Colorado (Dyer, 1981).

2.8.4. Regional Joint Patterns Cutting Across Tectonic Grain

On a regional scale the mechanism of joint propagation is often difficult to identify. Yet, joints mapped over a large region show patterns which are repeated from outcrop to outcrop. This is so even within old (>600 Ma) basement rock which has been subject to a variety of situations where the effective stresses were conducive to joint propagation. One example includes the joints of the Precambrian rocks of the Idaho Springs area of central Colorado (Harrison and Moench, 1961). Here joint patterns persist through a variety of rock types and structural styles.

Microfractures. Intrusive rocks of New England contain microscopic fabrics that allow quarrymen to split the rocks relatively easily in a consistent direction. This characteristic direction within granites is known as the rift plane. Usually the same rocks contain a secondary plane of easy splitting called the grain. One component of the microscopic fabric is an aligned set of microcracks which are either open or filled (Wise, 1964). These microcracks can be aligned over large regions. Wise (1964) compiled Dale's (1923) data on rift and grain in New England to show distinct regions where rift and/or grain has a preferred orientation. Metamorphic rocks may have the same aligned microcracks as is the case for the quartzites of the Piedmont west of Washington D.C. (Tuttle, 1949).

Joints. In most midcontinent settings regional joint patterns appear to be less complicated than in regions of one or more high strain deformational events. The simplest of patterns appear to be a double set of joints that are

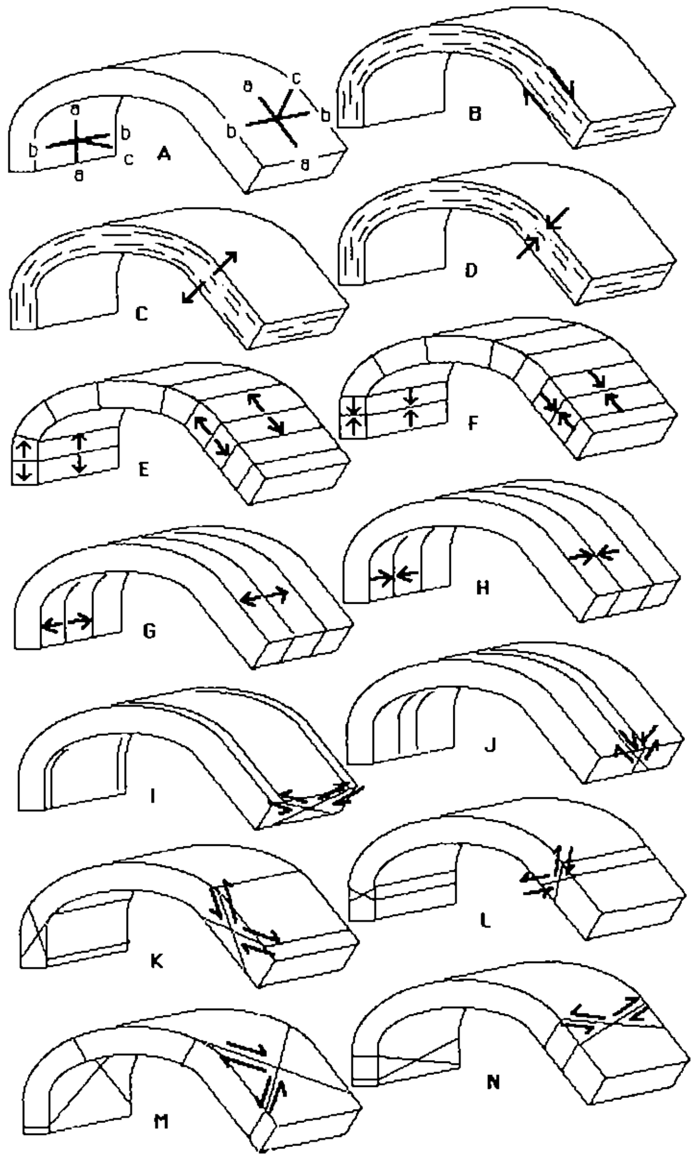


Figure 2.16. Block diagrams illustrating mesofracture sets and systems symmetrically arranged with respect to sedimentary layering and fold hinge lines (after Hancock, 1985). (a) Definition of the fabric orientations; (b) shear surfaces in *ab*; (c) extension fractures in *ab*; (d) stylolites in *ab*; (e) extension fractures in *bc*; (f) stylolites in *bc*; (g) extension fractures in *ac*; (h) stylolites in *ac*; (i) conjugate 0*kl* fractures enclosing an acute angle about *b*; (j) conjugate 0*kl* fractures enclosing an acute angle about *c*; (k) conjugate *h*0*l* fractures enclosing an acute angle about *a*; (l) conjugate *h*0*l* fractures enclosing an acute angle about *c*; (m) conjugate *hk*0 fractures enclosing an acute angle about *a*; (n) conjugate *hk*0 fractures enclosing an acute angle about *b*.

Figure 2.15. Non-systematic joints within the Ordovician Mallacoota beds of Australia.

impossible to decipher. In many other cases joint patterns are systematically related to the orientation of local folds and faults.

In some outcrops shear fractures form simultaneously in four or more sets (Oertel, 1965). Reches (1978) shows that simultaneous shear fractures may be associated with three-dimensional strain. Three or four sets of shear fractures are necessary to accommodate three-dimensional strain where the orientation of these shear fractures depends upon the strain to which the outcrop has been subjected. One example of simultaneous multiple faulting is found within the metavolcanic rocks of Modac, Canada where four fractures accommodate slip with a fifth occasionally accommodating additional slip (Robin and Currie, 1971).

2.8.1. Contraction Cracks

Some igneous rocks are highly jointed with single joints extending vertically for some distance but intersecting other joints within short lateral distances. These joints then give vertical exposures a columnar appearance because the rock has been broken into a set of locally parallel vertical prisms. This phenomenon, known as columnar jointing, is displayed in such localities as the Giant's Causeway, Northern Ireland (Weaire and O'Carroll, 1983), the Palisades Sill, New York (Walker, 1940), Devil's Tower, Wyoming (Robinson, 1956), and Devil's Postpile, California (Huber and Rinehart, 1967). Columnar joints form by tensile cracking that accommodates contraction during thermal cooling (Lachenbruch, 1962; Spry, 1962; Ryan and Sammis, 1978). Columnar joints are a subclass of contraction cracks which form under circumstances similar to the desiccation of mud and cooling after contact metamorphism. The nature of formation of columnar joints is inferred from the geometry of intersection and the spacing between the columnar joints. The same information may be inferred from cracks formed in soft sediment (Gilbert, 1882).

Intersections associated with columnar jointing include X, T, and Y geometries. In cooling igneous bodies T and X intersections are more common near contacts and free surfaces such as master joints (Spry, 1962) whereas Y intersections tend to be more common within the interior of the bodies (Gray *et al.*, 1976; Weaire and Carroll, 1983). The Moruya Batholith near Tuross Head, Australia contains a mafic magma which mixed with a sialic magma (Vernon *et al.*, 1983). During cooling the mafic magma solidified first and displays predominant orthogonal cooling joints. In Fig. 2.14 two sets of orthogonal cooling joints appear with the later set rotated by 45° from the earlier. Here X and T intersections appear in association with the nearby boundary of the mafic bleb which is about 2 m across. The X and T intersections are characteristic of joints propagating at different times.

Y intersections present conceptual problems. If it is assumed that Y intersections form by the outward propagation of three fractures from a point, then in a hexagonal pattern of columnar jointing these outward propagation fractures must meet at points where three fractures come together. In this case the Y intersections would form synchronously. Another solution to the formation of Y intersections is that they form by crack branching or bifurcation where the intersections do not have to form synchronously (Lauchenbruch, 1962; Spry, 1962). In their study of the basalt of Boiling Pots, Hawaii, Ryan and Sammis (1978) concluded that the majority of Y intersections did not form synchronously. Aydin (personal communication, 1986) suggested that mixed-mode crack tip behaviour was responsible for crack segmentation and the formation of Y intersections.

5cm

Figure 2.12. Later veins branching off sigmoidal veins within the Merrimbula sandstone of Australia.

structures" and, therefore, has taken Beach's (1975) point of view that they are shear cracks. In the case of the Merrimbula sandstone with multiple-vein arrays, later veins branch off sigmoidal veins or cut across them (Fig. 2.12). Synchronous propagation of all veins would have supported a shear origin for the parent vein and the branching veins would then be pinnate joints. Some of the veins do have small splay veins at an angle to the walls of the discontinuity as is the case for a shear vein formed as a hybrid shear. However, the prime evidence supporting the shear origin is the parallelism of veins and the conjugate shear zone (Beach, 1975).

2.7.4. Pinnate Joints Within a Shear Zone

Some larger shear fractures or shear zones have pinnate joints which are larger scale versions of the microcracks (microscopic feather fractures) forming an acute angle with the fault zone. Again the apex of the acute angle between the pinnate joints and shear zone points in the direction of shearing. Pinnate joints have the same geometry as en echelon twist hackles at the fringes of joints formed under mode I loading (Fig. 2.13).

opening of a mixed-mode crack even as the crack tip was propagating along a plane subject to a shear stress. There is a difference between the extensional shear joints and common shear fracture formed at higher confining pressure; in the former case process zones are narrow and difficult to identify. Microcracks should appear within the intact rock next to the hybrid shear but are rare. Figure 2.10 shows veins in the same orientation (east–west) that microcracks within the process zone should assume. Likewise, because the extensional shear joint pulls apart upon formation, cataclasis does not occur during shear offset.

2.7.3. En Echelon Cracks

En echelon cracks are found on several scales from microns to kilometres. On the microscopic scale en echelon microcracks may link during the brittle failure in shear (Kranz, 1979). A cross-section of twist hackles through a joint has the appearance of en echelon cracks (Woodworth, 1896). Large dikes may intrude in an en echelon manner over distances of more than a kilometre (Pollard *et al.*, 1982). En echelon veins are common in deformed sedimentary rocks and are often associated with solution cleavage (Shainin, 1950; Rispoli, 1981). En echelon veins may be sheared so that individual veins have a sigmoidal pattern with the sense of shear consistent with the development of an incipient shear fracture (Durney and Ramsay, 1973).

On the mesoscopic scale en echelon veins closely approximate the process zone associated with the propagation and subsequent coalescence of microcracks to form a shear fracture. Some en echelon veins represent the breakdown of the parent fracture due to mixed mode I and III loading (Pollard *et al.*, 1982). However, the regularity of the veins in orientation as well as spacing makes them distinct from the microcrack cloud that precedes the massive shear failures. One hypothesis is that en echelon veins form within zones in which shear has been localized by some mechanism such as the development of a cloud of microcracks (Lajtai, 1969). In this case the veins propagate as mode I cracks following a locally reoriented stress field within a material that is already weakened by incipient brittle shear. En echelon veins form before the material completely loses cohesion to form the standard shear fracture described above.

En echelon veins form a shear zone whose angle may vary from 10° to 45° to the outcrop trace of the veins (Hancock, 1972). This is the entire range expected for shear fracture angles predicted by a parabolic Coulomb–Mohr failure envelope that becomes horizontal at very high confining pressure. Two patterns are common for conjugate sets of en echelon veins (Beach, 1975). In one pattern the veins in one shear zone parallel the trace of the

Figure 2.9. Braided shear fracture within sandstone layers of a turbidite from Australia. The shear offset of the sandstone is reflected in a kinking of the shale layers.

shape which gives the gouge zones a braided appearance (Engelder, 1974). The thickness of the braided zone is a function of shear displacement (Aydin and Johnson, 1978).

The nature of the shear fracture depends on the lithology as is illustrated with shear zones cutting a turbidite sandstone in New South Wales, Australia. Where the shear zone cuts the coarser grained sandstone a well developed braided shear fracture forms (Fig. 2.9). As the shear zone passes into the finer grained shales with a strong bedding-plane anisotropy, the shear changes from a braided shear fracture to a kink band. In the kinked layers the shear is taken up by slip along bedding planes.

other, indicating that the initiation of one joint is not linked to the tip of the predecessor as was the case for joints formed by intermittent cracking.

2.7. SHEAR FRACTURING

The process of shear fracturing is best understood from an experimental point of view where the sequence of events leading to rupture can be traced in detail. Shear fractures develop as a plane of shear failure only after a long history of microfracturing (Scholz, 1968; Lajtai, 1971). Steps include the formation of individual microcracks, propagation and linking of these cracks, and then larger scale shear failure often but not exclusively with an accompanying cataclasis of a zone within the host rock. Unlike the propagation of mode I cracks the large scale shear failure follows a plane of high shear stress without twisting or tilting out of that plane.

Shear fracturing occurs even when the applied stress is compressive. Under these circumstances failure accompanies the propagation of microcracks. The compressive stress is locally modified at these microcrack tips to become tensile which process causes the microcracks to propagate further under mode I loading. Gallagher *et al.* (1974) showed that stress risers, associated with elastic mismatches and pointed contacts in an aggregate, will also generate local tensile stresses capable of causing microcrack initiation and then propagation. These stress risers are activated when the aggregate is subject to large differential stresses. On average the microcracks propagate normally to the direction of the least principal stress (Brace and Bomolakis, 1963). There is no distinction between these microcracks and those discussed earlier in this paper.

Further compression will add microcracks until they start to link. This is a feed back process because the largest stress risers are at the tip of the longest cracks (i.e. the linked cracks). This feed back process has been mapped by Scholz (1968) and Lockner and Byerlee (1977) as a cloud of microcracks along the plane of future shear fracture. Such a cloud of microcracks constitutes the same type of process zone that forms at the tip of a joint during propagation of a microcrack under mode I loading (Friedman *et al.*, 1972). Delaney *et al.* (1986) have documented a process zone associated with dike emplacement that extends 100 m. During shear failure the cloud of microcracks eventually focuses along a zone inclined to the maximum compression at about 30°.

Massive failure occurs with the formation of a zone of cataclastic material having little or no cohesion. During further slip the cataclastic gouge zone will grow in width and become finer grained. Intact rock within the process zone on either side of the gouge zone may contain microfractures parallel with the direction of maximum compression (e.g. microscopic feather fractures (Friedman and Logan, 1970)). These are remnants of the cloud of microcracks that preceded massive shear failure. The apex of the acute angle

In this example of intermittent growth of a single joint, the rupture must come to a complete stop until the fluid pressure again exceeds the tensile strength of the rock. This model requires the pore pressure within individual pores of the rock to be higher than the fluid pressure necessary to drive a mesoscopic joint through the rock. Hence, there must always be leakage from the pores to the joint.

2.6.2. Crack–seal veins

Veins are joints that have been filled with a cement derived from hydrothermal processes including pressure solution. Quite often veins are formed by an accretionary process during which narrow cracks propagate followed by the filling of the narrow open space with a cement (e.g. Ramsay, 1980). Successive ruptures follow the cement–host rock boundary or cut a crack entirely within the cement (Fig. 2.6). The accretionary process, known as the crack–seal mechanism of rock deformation, may account for as much as 50% extension in some local situations (Ramsay, 1980). Individual components of the crack–seal veins are of the order of a few microns wide, whereas the entire vein may be as much as several cm wide. As was the case for the intermittent crack discussed above, hydraulic fracturing is believed to be the mechanism for crack propagation. The stress cycling involves effective stress where fluid pressure repeatedly exceeds the tensile strength of the vein material or vein–host rock boundary.

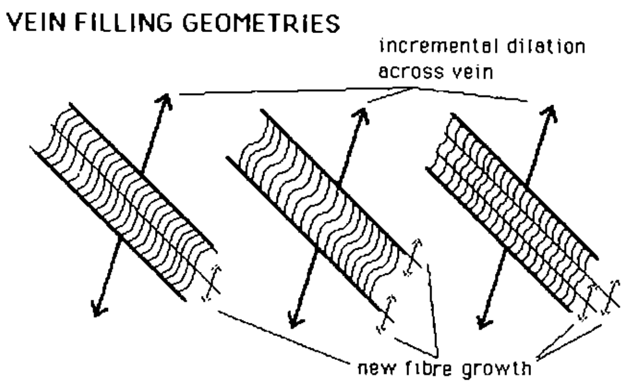
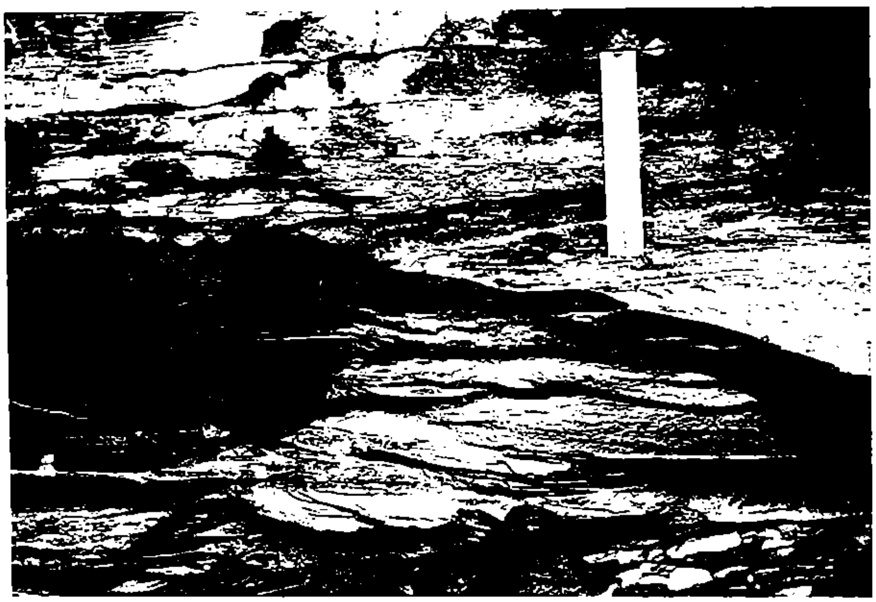


Figure 2.6. Schematic of the crack–seal process (after Durney and Ramsay, 1973). Type I veins show syntaxial growth. Type II veins show antitaxial growth. Type III veins show changing increments of strain.

1942)) or that the rupture comes to a complete stop until stress conditions change enough to reactivate the rupture. In the latter situation the short-term rupture velocities may be relatively fast.

Arrest lines are found on all scales from thin beds to 25 m tall faces in massive sandstone. Massive sandstone beds in the Sinai of Egypt have joints displaying large undulations which Bahat (1980) compares with conchoidal fractures. These undulations seem to have the characteristics of arrest lines as described above.

Velocity of the rupture. Because joint growth has never been observed directly, estimates of velocity can only be inferred by comparing the fracture geometries with known experiments. Relatively high velocities are reflected in the branching of crack tips (Lawn and Wilshaw, 1975). The ridges of arrest lines in shales resemble a form of branching and may represent the arrest of a relatively higher velocity crack. Crack branching in a siltstone exhibits another form of arrest (Fig. 2.5). In contrast Segall (1984) argues that joint sets in the granitic rocks of the Sierra Nevada do not show evidence of branching and, hence, must have propagated quasi-statically (i.e. relatively slowly).



25 cm

Figure 2.5. Crack branching of a joint within a Devonian siltstone bed from the Appalachian Plateau, New York. Common plume structures are seen on the left. As the rupture moved to the right branches formed as seen within the right-hand portion of the photo. Scale is 15 cm.

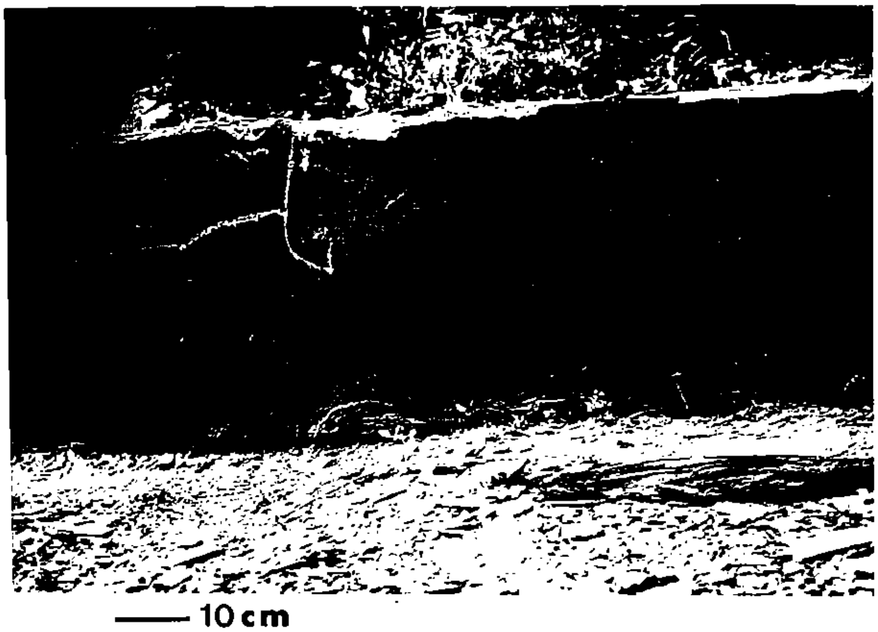


Figure 2.4. The plumose pattern on a joint within the Genesee Group of the Appalachian Plateau, New York. Note the initiation point at the bottom of the bed and the barbs of the plume propagating to the right from the initiation point and then curving back on itself to propagate to the left towards the top of the bed. Lens cap for scale.

crack that were subject to the lowest tensile stresses. Early workers (e.g. Parker, 1942; Roberts, 1961; Gash, 1971) interpreted joints with twist hackles as shear fractures but both Kulander *et al.* (1979) and Engelder (1982a) have presented evidence (e.g. no shear offset) that joints with twist hackles are tensile in origin on the scale of the entire joint. The problem with referring to these en echelon features that bound beds as hackles is that there is no evidence that the local stress reorientation was caused by high crack velocity. Equally likely is that the stress at the edge of the bed was rotated prior to and independent of the fracture process.

On joint surfaces the most commonly seen features that record rupture motion are plume structures called feathers (Woodworth, 1896), hackle plumes (Kulander *et al.*, 1979), striations (Bahat, 1979), or barbs (Bahat and Engelder, 1984) (Figs 2.1 and 2.4). The plume structure consists of an axis from which the striations or barbs mark the direction of the rupture front as portions diverge away from the plume axis. Barbs often become more pronounced towards the edge of beds and furthest from the plume axis. The mechanism for the formation of barbs seems to be similar to that for the hackle marks where the barbs were long narrow planes oblique to the main

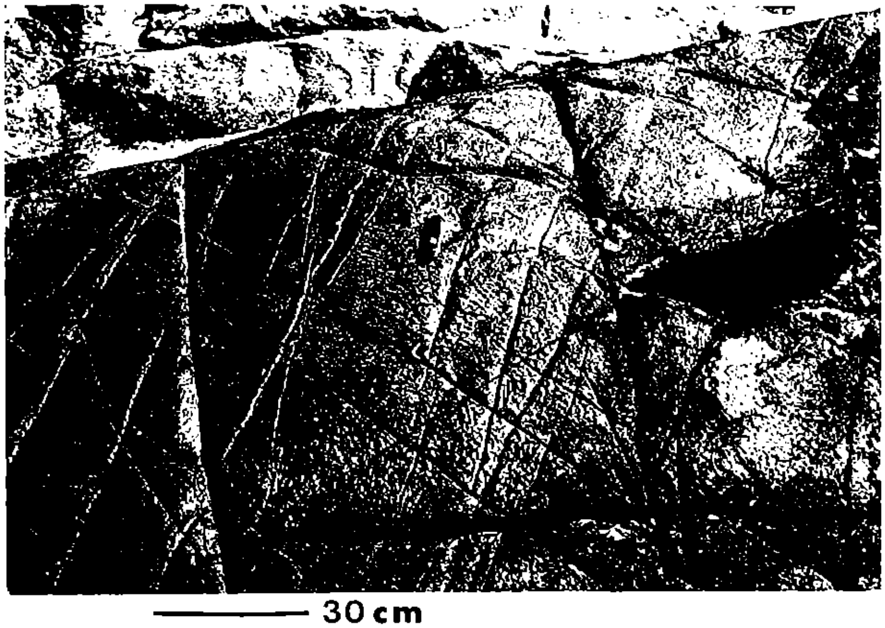


Figure 2.2. Subhorizontal Wallner lines on the surface of a vertical joint within a Devonian shale from the Appalachian Plateau, New York, USA.

Wallner lines. Ripples observed in the mirror region of cracks in glass are called Wallner lines. These ripples arise as the rupture front of the crack and higher velocity sonic waves couple while crossing each other (Kulander *et al.*, 1979). Because the sonic waves do not intersect the rupture front simultaneously along the front, the Wallner lines are not a true record of the shape of the rupture front at any point in its history. Sonic waves are generated as the rupture front moves through existing flaws within the material. Wallner lines may not be common on joint surfaces because sonic waves may immediately outdistance the slowly moving rupture front. Some subhorizontal ripples on joint faces in shale are thought to be Wallner lines (Fig. 2.2).

The mist zone. As the velocity of the rupture front increases through a critical value, the rupture front starts to deviate from the plane of the crack. This happens as the stresses at the crack tip become large enough to break the material at oblique angles to the crack plane (Bahat, 1979). Once oblique cracking on a fine scale has started a crack in glass no longer maintains a mirror-like surface but rather becomes misty. Within the mist region the first fine striations form looking like microscopic blades with their which permits interpretation concerning rupture nucleation, propagation, and arrest (Kulander *et al.*, 1979). The plumose pattern develops largely because of local twists and tilts during propagation that otherwise would be planar. Focus in this section is on evidence for the single rupture event which is characterized by an initiation point and either a single arrest or multiple synchronous arrests.

2.5.1. Nucleation

For joints with a well developed surface morphology, progress of the rupture may be followed backwards along surface irregularities called barbs (Bahat and Engelder, 1984) to a focus point at the origin (Fig. 2.1). In relatively

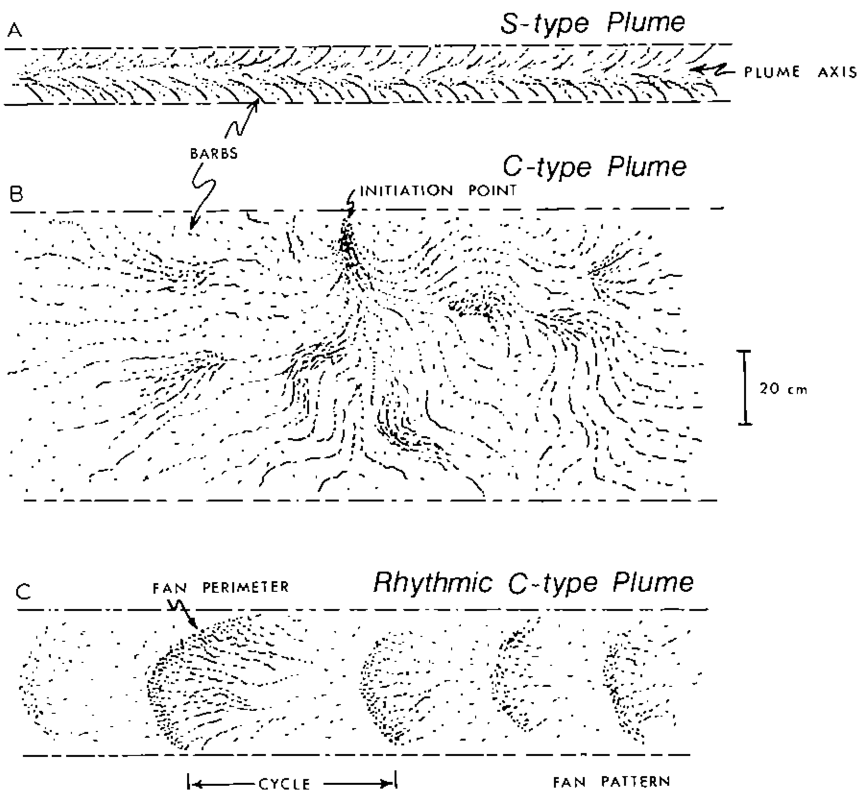


Figure 2.1. Various plume patterns observed on the siltstones of the Appalachian Plateau (after Bahat and Engelder, 1984). The barbs mark the direction of local fracture propagation. The fan perimeters in the rhythmic C-type plumes designate loci of arrest lines. They are convex toward the direction of fracture propagation.