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MICROCRACKS IN ROCKS: A REVIEW

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ABSTRACT

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In the past decade the number of studies about microcracks in rocks has rapidly increased. This review of recent work concentrates on microcracks in rock as separate entities, emphasizing microcrack morphogenesis, kinematics, dynamics, population statistics and observational techniques.

Cracks are produced when the local stress exceeds the local strength. The local stress may be augmented by twin lamellae interactions, kink bands and deformation lamellae, stress concentrations at grain boundary contacts and around intracrystalline cavities. Local strength may be reduced along cleavage planes, along grain boundaries, and along any internal surface as a result of corrosion by chemically active fluids. Dislocations appear not to be a significant factor for crack nucleation below about 500°C in silicates.

Spatial and temporal changes in temperature can also induce microcracking as a result of differential thermal expansion between grains with different thermoelastic moduli and thermal conductivities. The amount of quartz in the rock has a significant effect on thermally induced microcracks because of its large and variable thermal expansivity.

The application of hydrostatic pressure between 100 and 200 MPa effectively closes most cracks, but the closure may not be uniform if crack wall asperities exist. Hydrostatic pressure appears to stabilize cracks and make crack growth more difficult. The number and average size of mechanically induced microcracks is greater in rock deformed at higher pressures.

The application of a deviatoric stress field on the boundaries of a rock mass results, on a microscopic scale, in a very complex stress system which greatly affects nucleation and propagation paths. The relative amount of intragranular and intergranular cracking appears to depend upon mineralogy, rock type and stress state. The vast majority of stress-induced microcracks in rocks appear to be extensional. Statistically, they are predominantly oriented within 30° of the macroscopic maximum stress direction. Crack densities increase as macroscopic deviatoric stress increases above a threshold level. Crack size distributions may be either lognormal or exponential.

Fracture in rock under compressive boundary loads is a result of the coalescence of many microcracks, not the growth of a single crack. Some crack configurations are more favorable for coalescence than others. As deviatoric stress increases and rock failure is approached, the microcrack population changes

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spatially from random to locally intense zones of cracking. Away from the fault, the crack density dies off rapidly to the background level a few grains away.

Under lesser deviatoric stresses, slow, subcritical microcrack growth can occur as a result of stress-aided corrosion at the crack tip. The rate governing mechanism may be either the chemical reaction rate or the rate at which water can get to the crack tip. Important details still remain to be worked out.

INTRODUCTION

Within the last 10 years, an enormous amount of work has been done to further our knowledge about microcracks in rock. For obvious reasons, much of this work has been directed toward understanding how microcracks affect physical rock properties such as compressibility, strength, elastic wave velocities, permeability and electrical conductivity. Of equal interest are microcracks as separate entities: how they form, respond to various stress states, grow and interact, as well as their actual morphologies and population statistics.

Studies of microcracks in rock may throw light on a rock's history. They may be used to infer local stress domains. They may be used for comparisons between natural and laboratory conditions, and because of apparent morphological and mechanical similarities between microcracks and joints and faults, they may be useful for geophysical scaling problems. Knowledge about microcrack populations is a necessary input to model studies of the micromechanics of fracture and fault formation. Microcracks may be pathways for the migration of chemically active fluids through rock.

For these reasons, it appears timely and worthwhile to review the literature as it pertains to microcracks in rock. Paterson (1978) discusses the role of microcracks in physical rock properties. Friedman (1975) and Logan (1979) cover the state of knowledge about rock fracture and brittle phenomenon up to the time of their reviews. Simmons and Richter (1976) describe microcrack characteristics in igneous rocks. While this review will overlap these somewhat, I intend to concentrate on microcracks in rock as separate entities, emphasizing recent work on microcrack morphogenesis, kinematics, dynamics and statistical studies. I have elected to cover as many aspects of the recent work as possible at the expense of in-depth analysis and theoretical framework which is covered in the references.

For completeness, a brief review of observational techniques used in microcrack studies will be given in an Appendix.

Though intended to be comprehensive, this review is not exhaustive and omissions or oversights are unintentional. Much of this work is still progressing as fast as new technology can be adapted to the study of microcracks in rock. I have therefore included references to recent Ph.D. theses and abstracts. The majority of references are, however, to accessible English language journals and conference proceedings. Perhaps the major value of this review will be its bibliography.

MICROCRACK MORPHOGENESIS

Morphology

Simmons and Richter (1976) define the word *microcrack* as:

> "an opening that occurs in rocks and has one or two dimensions smaller than the third. For flat microcracks, one dimension is much less than the other two and the width to length ratio, termed crack aspect ratio, must be less than 10^{-2} and is typically 10^{-3} to 10^{-5}. The length... typically is of the order of 100 μm or less."

They subdivide microcracks into *grain boundary cracks* (associated and perhaps coincident with the grain boundary), *intragranular* or *intracrystalline* cracks (lying totally within the grain) and *intergranular* or *intercrystalline cracks* (extending from a grain boundary crossing into one or more other grains). The term *transgranular crack* is sometimes used for cracks running across a grain from grain boundary to grain boundary. Most author's terminology conforms to these definitions, but descriptions are often ambiguous. It is often difficult, for example, to distinguish several connected intragranular cracks from a single transgranular crack. An intergranular crack may be a grain boundary crack along part of its length (e.g. Dunn et al., 1973, fig. 11). Thus more detailed descriptions of observations are to be encouraged. Crack type labels obviously depend to some extent on the resolution of the observation.

The terms *microcrack, microfracture* and *crack* will here be used synonymously. We adopt the above divisions for clarity.

A fair number of published photomicrographs illustrating various details of different crack types now exist in the literature. Simmons and Richter (1976) suggest that there is a unique relationship between microcrack characteristics and the process which produced the characteristics. Accordingly, they distinguish (and illustrate with photomicrographs from thin sections of igneous rocks) various crack types by mode or origin. While one may take exception to some of their divisions, their descriptions are excellent and need not be repeated here. Their observations of the various crack types are compatible with and supplemented by other studies by Richter et al. (1976), Batzle and Simmons (1976), Richter and Simmons (1977), Wang and Simmons (1978), and Shirey et al. (1980). An additional crack type, *microscopic feather fractures,* is described by Friedman and Logan (1970) and Conrad and Friedman (1976).

Descriptions of cracks observed with the aid of the scanning electron microscope (SEM) are usually more detailed due to the higher resolution possible. Such descriptions, with photomicrographs, may be found in Brace et al. (1972), Sprunt and Brace (1974a, 1974b), Tapponnier and Brace (1976), Kranz (1979a, 1979b), Dengler (1976, 1979) and Padovani et al. (1982). Microcracks in rock-forming minerals seen with the transmission electron microscope (TEM) are illustrated and discussed by Boland and Hobbs (1973), Tullis and Yund (1977), Marshall and McLaren (1977) and Dunning et al. (1980).

Data collected from the literature on microcrack sizes, numbers and orientations will be presented in the section on crack statistics. It must be emphasized, however, that statistics alone are inadequate to describe the variety and complexity of cracks found in rock, especially as it now appears that rock type, composition, history and even sample preparation influence what has been seen. With that caution, a synoptic description of the four basic microcrack types is given below.

Grain boundary cracks

Grain boundary cracks may be subdivided into those coincident and those non-coincident with the actual crystal boundary (Simmons and Richter, 1976). It is often difficult to make this distinction because the grain boundary may not be obvious.

Non-coincident grain boundary cracks may extend for short distances into the grain at a high angle to the grain boundary (Simmons and Richter, 1976, figs. 5 and 6) or may be close to and subparallel to the boundary, possibly running through grain boundary asperities or adjacent cement.

Coincident grain boundary cracks, to the resolution attainable, are separations (open space) between part or all of the crystal boundary and adjacent material. Many grain boundaries found in unstressed, virgin crystalline rock are partially healed, and only a row of slot-like elongated cavities mark the grain boundary (Sprunt and Brace, 1974b; Kranz, 1979a; Padovani et al., 1982). Sprunt and Brace (1974b) comment that in gabbro and diabase, the grain boundaries they observed were crack free, while in granites, the percentage of the grain boundary which was cracked varied. In metamorphic rocks, Padovani et al. (1982) observed that grain mineralogy was a factor in determining whether grain boundaries were partially open or completely sealed.

In thermally or mechanically stressed rock, grain boundaries are often highly separated (Sprunt and Brace, 1974b, fig. 10; Kranz, 1979a, fig. 4) or they may be crushed by pressure and loads applied normal to the boundary (Sprunt and Brace, 1974a; Batzle et al., 1980, fig. 7). Coincident grain boundary cracks in thermally or mechanically stressed rock may be continuous along the boundaries of several grains.

Intracrystalline cracks

Intracrystalline cracks are relatively small, usually much less than a grain diameter in length and about 1 μm or less in width. Natural intracrystalline cracks may have rough-edged walls which may be filled in or bridged by other material (i.e. all or partially healed). Quite often, they are simply slot-like cavities within the grain with aspect ratios of 10^{-2} to 10^{-4} and with blunt tips. In contrast, thermally or mechanically induced intracrystalline cracks have sharp walls, are generally narrow, and tips are sharp or tapered (Sprunt and Brace, 1974b; Tapponnier and Brace, 1976; Kranz, 1979a).

Just as for transgranular and intergranular cracks, intragranular cracks may be tilted or twisted in three dimensions, or have traces in two dimensions which are curved. Their opposing sides may be skewed or in register. Based on post-deformation observations, most stress-induced cracks appear to be extensional, Mode I type cracks (Tapponnier and Brace, 1976; Kranz, 1979a; Wong, 1980), with little shear motion between the crack surfaces.

Intercrystalline cracks

Intercrystalline cracks are longer and often wider than the intracrystalline cracks but are otherwise morphologically similar. They are generally transgranular and are easily seen at low magnifications. Orientation of the crack may vary approaching a grain boundary or be deflected by the grain boundary and other cracks (Simmons and Richter, 1976, fig. 9). In porous sedimentary rock, they may run from and through grain contact points, following stress trajectories in each grain (Gallagher et al., 1974), or merge with appropriately oriented grain boundary cracks. In mechanically stressed rock, the tip-to-tip intercrystalline crack orientations are usually subparallel to the macroscopic maximum stress direction.

Cleavage cracks

Cleavage cracks are an important enough subset of intracrystalline cracks to be considered separately. They are separations along cleavage planes in minerals and as such are petrographically useful. They are often seen to occur in parallel sets of various lengths within one grain. More than one cleavage plane within a grain may be cracked. Unless the mineral has been kinked, the cleavage crack walls are extremely sharp and virtually parallel except where they gradually merge with uncracked material. In virgin rock, cleavage cracks may be decorated by alteration products (Simmons and Richter, 1976).

Frequently observed cleavage cracks occur in quartz along the r, z directions (e.g. Swolfs, 1972; Martin and Durham, 1975; Dunning, 1978), in feldspars along (001) and (010) directions (e.g. Bombolakis, 1973; Tullis and Yund, 1977; Marshall and McLaren, 1977; Kronenberg and Shelton, 1980), in micas along (001) (e.g. Wilson and Bell, 1979) and in pyroxene along (110) (e.g. Kronenberg and Shelton, 1980).

Generation of microcracks by mechanical stresses

As succinctly stated by Simmons and Richter (1976), "cracks in rock are produced when the local stress exceeds the local strength". Local stresses may be mechanically or thermally induced. In this section, we will consider various means of mechanically inducing crack nucleation and generation, saving for the next section thermally induced situations.

Dislocations are known to play a dominant role in producing cracks in metals. Dislocation pile-ups and interactions with grain boundaries and other dislocations can generate cracks (McClintock and Argon, 1966; Das and Marcinkowski, 1972; Lawn and Wilshaw, 1975). In silicate minerals, however, at low to moderate temperatures, dislocations are apparently not mobile or dense enough to directly cause crack nucleation (Carter and Kirby, 1978). Martin and Durham (1975) found few dislocations in cracked quartz specimens at temperatures up to 250°C. Those seen were random and unrelated to the cracks. Tullis and Yund (1977) found that dislocations are relatively unimportant in granite deformed below 500°C. Dunning et al. (1980) found no evidence of dislocation nucleation or mobility influencing crack growth in quartz in chemically active environments at room temperature. In calcite deformed at room temperature, dislocations were observed indirectly using etch pitting techniques, but dislocation pile-ups were not associated with cracks (W. Olsson, pers. commun., 1981).

At higher temperatures and pressures, and near the yield stress, dislocations have been observed to be associated with microcracks (Heard and Carter, 1968; Boland and Hobbs, 1973; Tullis and Yund, 1977; Marshall and McLaren, 1977; Carter and Kirby, 1978; Kronenberg and Shelton, 1980; Carter et al., 1981). The role of dislocations in nucleating microcracks in rock-forming minerals still merits, however, much further investigation.

Under the experimental and natural conditions where cracks are known to exist and proliferate, mechanical stresses can generate cracks through at least six mechanisms. These include microcracks produced as a result of (1) twin interactions with grain boundaries and other twins, (2) release of stored strain energy associated with kink bands and deformation lamellae, (3) cleavage separations, (4) stress concentrations in the neighborhood of grain boundaries, intracrystalline cavities and crack tips, (5) mismatches in elastic compliances of neighboring grains, and (6) grain translations and rotations. We will now briefly consider each of these.

Twin induced microcracking

Olsson and Peng (1976) identify and illustrate three types of microcrack nucleation involving twin lamellae (Fig. 1). These mechanisms are general and have been observed in calcite (Olsson and Peng, 1976; Rizer and Bombolakis, 1977; Koelsch, 1979), plagioclase (Marshall and McLaren, 1977; Mitra, 1978) and quartz (Martin and Durham, 1975). Twin lamellae concentrate stresses in much the same way as

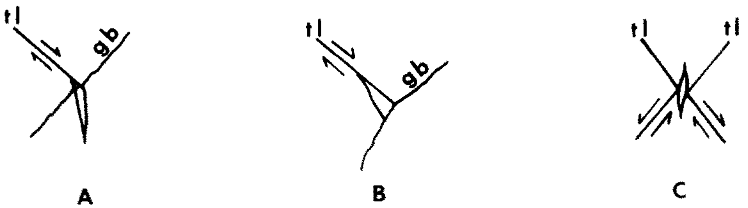

Fig. 1. Schematic diagram of microcrack nucleation mechanisms involving twin lamellae (after Olsson and Peng, 1976). *tl*: twin lamella, *gb*: grain boundary.

dislocations where they meet a grain boundary or a second lamellae. When these stresses exceed the local strength, a crack is nucleated. It is interesting to note, however, that Marshall and McLaren (1977) suggest that the reverse situation may occur. That is, the stress concentration around the crack may provide a site for the nucleation of twins.

Kink band and deformation lamellae associated microcracking

Accommodation problems can arise on one side of kink band boundaries and deformation lamellae which lead to either compressional or extensional strain concentrations. This strain may be relieved by the nucleation of microcracks approximately normal to the boundary (Carter and Kirby, 1978), Marshall and McLaren (1977) observed that slip lines approximately parallel to (010) and normal to kink bands in albite consisted of microcrack arrays. Microcracks approximately normal to kink bands in quartz have also been observed (Carter and Kirby, 1978, fig. 3b). Microcracks emanating from dislocation arrays (deformation lamellae) have been observed in quartz (Heard and Carter, 1968, plates 1 and 2), in microcline (Tullis and Yund, 1977, fig. 3c—note the captions for figs. 3 and 4 in this paper were interchanged) and in plagioclase (Fig. 2). Marshall and McLaren (1977) observed, however, that some optical deformation lamellae in plagioclase consist of "bundles of microcracks" as well as tangled dislocations.

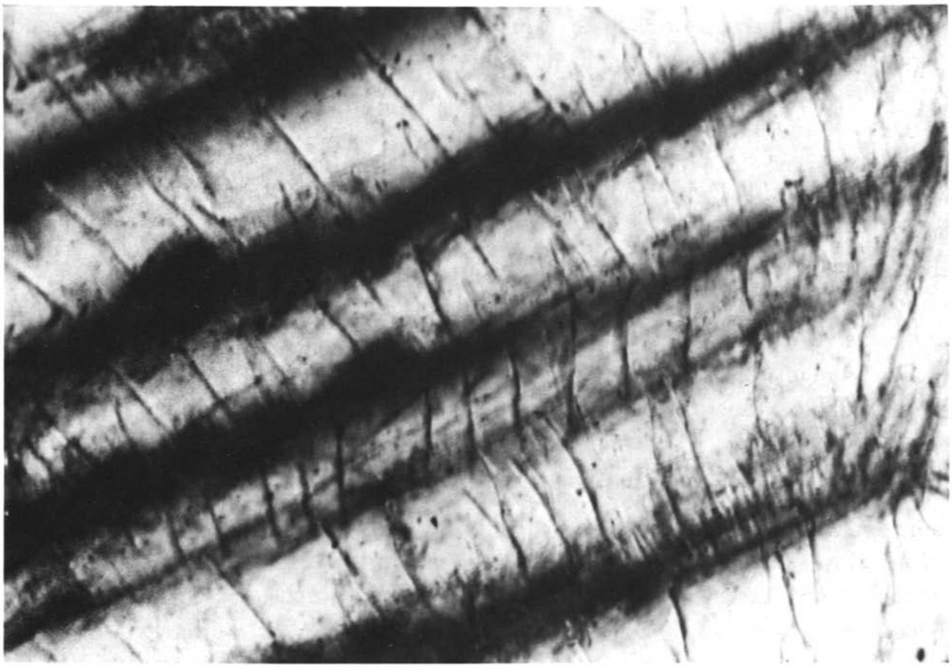

Fig. 2. Microcracks associated with deformation lamellae in naturally deformed plagioclase (courtesy of Robert Hooper). Scale: 575 μm top to bottom.

Cleavage separations

Cleavage planes in crystals are generally those planes with the lowest bond density or strength, and lowest surface energy (Brace and Walsh, 1962, supporting theory of Gilman, 1959). Thus in suitably oriented grains, strain energy stored in the grain as a result of applied stresses will tend to be relieved on those planes first. Published examples of cleavage cracking in rock-forming minerals have been previously noted.

Microcracking from stress concentrations at boundaries

Point and line contacts between portions of grain boundaries are sites of highly concentrated stresses. Tensile stresses exceeding local tensile strength are easily attained and microcracks nucleated in these regions are almost always extensional, Mode I cracks. Gallagher et al. (1974) analyzed and gave examples of microcracking initiated at grain boundaries in sandstones and for cemented and uncemented aggregates. Hallbauer et al. (1973) found that point loading of quartz grains by other grains was a frequent source of microcrack initiation in quartzite. Batzle et al. (1980) observed with the SEM new cracks being formed at points of crack wall contacts. The analyses of indentation fracture given by Swain and Lawn (1976) and crack wedging by Dey and Wang (1981) are also applicable here on the grain scale.

Microcracking from stress concentrations around cavities

Both pre-existing intracrystalline cracks and pores concentrate stresses within the surrounding solid material. The sign and magnitude of these stresses depends primarily on the orientation and geometry of the microcavity with respect to the externally applied stress field and, secondarily, on the tensor mechanical properties of the surrounding medium. Microcracks associated with intracrystalline pores were shown by Kranz (1979b, figs. 7–10) and Wong (1982, fig. 9). Bombolakis (1973) and Adams and Sines (1978b) have investigated the common occurrence of the initiation of secondary "daughter" cracks from pre-existing cracks. Such secondary cracks are apparently initiated at the point on the boundary of the pre-existing microcavity where the tensile stress concentration is greatest. This occurs near crack tips and acute angled pore boundaries. Bombolakis (1973) observed, however, that "as long as flaws remain open during loading, the geometry of crack walls away from the crack tips has little effect on stress concentrations around the crack tip".

Elastic mismatch induced microcracking

When two different minerals in effectively welded contact are subjected to the same externally applied stress, the stiffer or least compressible mineral will also be subjected to additional boundary tractions (Dey and Wang, 1981). These can exceed the local tensile strength of bonds at the boundary, leading to extensional cracks in the stiffer mineral (Fig. 3) [1]. The same situation can arise for two like, but

[1] Note that Brace (1976, fig. 8) shows cracks running from a quartz–feldpsar boundary into the feldspar

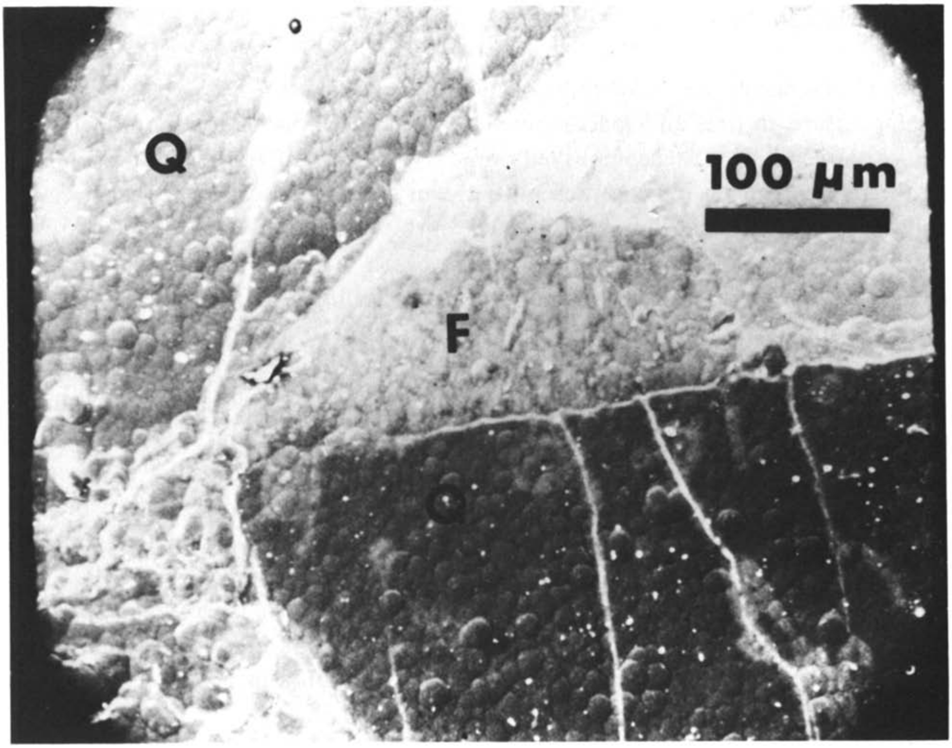

Fig. 3. Microcracks running from quartz–feldspar grain boundary into quartz. Compressive uniaxial stress on remote sample boundary was in vertical direction.

anisotropic minerals which are misaligned. Simple calculations will show that tensile stresses on the order of 100 MPa or more can arise in this way.

Grain rotations and translations

Under deviatoric stresses, grain boundary sliding in crystalline rock may occur, producing coincident grain boundary cracks. Cemented grains in sedimentary rock may be wedged apart and rotated by neighbor grains, producing cracks in the cement or along the grain boundary. Savanick and Johnson (1974) found, for example, that the pebble-matrix interface in a conglomerate had a unconfined tensile strength of only about 15 MPa. Quartz–feldspar interfaces separated under 10 MPa tensile stress. In crystalline rock grain rotations and translations may be more difficult, because of tighter interlocking, but can occur during cataclastic deformation.

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in contrast to Fig. 3 here. The inference is that the axis parallel to the ion milling mounds in his figure is stiffer in the microcline than the corresponding direction in the quartz.

Thermally induced microcracking

In the absence of sufficient mechanically induced deviatoric stresses, differences in temperature, in time and space, can cause microcrack nucleation by inducing the requisite stresses. This is accomplished primarily through differential and incompatible thermal expansion (or contraction) between grains with different thermoelastic moduli or between similar, but misaligned anisotropic grains. The induced stresses are on the order of the thermal expansion contrast across grain boundaries multiplied by the temperature change and Young's modulus. There may be, however, a confining pressure above which thermally induced stresses are not sufficient to cause microcracking by themselves (Wong and Brace, 1979; Van der Molen, 1981). Thermally induced microcracking may also be initiated within individual grains at internal boundaries which are sites of thermal gradients and attendant stress concentrations. Bruner (1979) argues that the direction of growth of a thermally induced intracrystalline crack depends on the relative magnitudes of the thermal expansion coefficients. In a confined and otherwise elastically isotropic grain, tensile cracks would tend to grow in the direction of minimum expansivity.

Experiments, mostly on crystalline rock, have shown that the absolute value of temperature, the heating or cooling rate and consequently induced thermal gradients, the thermal history and the mineralogy can all affect the numbers and characteristics of thermally induced microcracks.

Significant microcracking, as detected by acoustic emissions, seems to begin above a critical threshold temperature: about 70°–75°C for granite (Todd, 1973; Johnson et al., 1978; Bauer and Johnson, 1979; Yong and Wang, 1980) and 200°C for quartzite (Johnson et al., 1978). The threshold temperature is sensitive to the thermal history. New microcracking begins after the previous maximum temperature has been surpassed (Johnson et al., 1978; Yong and Wang, 1980). Slow, uniform heating above this temperature produces more open space in the rock. This is primarily due to separation of grain boundaries and secondarily a result of intragranular cracking (Sprunt and Brace, 1974a; Simmons and Richter, 1976; Friedman and Johnson, 1978; Potter, 1978; Bauer and Johnson, 1979). In Sioux quartzite, Friedman and Johnson (1978) found that most of the displacement was normal to, rather than parallel to crack surfaces. Intragranular cracks were subperpendicular to grain boundaries, tapering in from the boundary. Bauer and Johnson (1979) observed that most thermally-induced cracks in feldspars in granites were cleavage cracks. Pre-existing cracks or cracks formed at lower temperatures get larger at higher temperatures (Sprunt and Brace, 1974a; Simmons et al., 1975; Simmons and Cooper, 1978; Potter, 1978; Bauer and Johnson, 1979).

The amount of quartz in the rock has a significant effect on thermally induced microcracking because of its relatively large and anisotropic coefficients of thermal expansion. Below the α–β transition, differential thermal expansion between neighbor quartz and feldspar grains plays a dominant role in producing cracks in granite.

"Above $T_{\alpha\beta}$ new crack formation is minor; porosity increases by widening of earlier formed microcracks." (Johnson et al., 1978). The relative absence of crack porosity in diabase compared with granitic rock has been attributed to the lack of quartz (Nur and Simmons, 1970). Simmons and Cooper (1978) show, however, that crack porosity does increase with increasing temperature in diabase.

Differential contraction upon cooling can also produce cracks. An unusual example of thermal-cooling induced microcracking has been provided by Stout (1974). He showed that orthoamphiboles from metamorphic rocks developed helicoidal cracks about the *c* axis nucleated at quartz–amphibole boundaries during cooling. The quartz remained uncracked.

Thermal gradients can induce microcrack nucleation and aid microcrack growth. Heterogeneous temperatures in rock are a function of the heating rate, thermal conductivity of the constituent minerals and existing interior boundaries. Todd (1973) and Yong and Wang (1980) showed that microcracking rate, inferred from acoustic emissions, was strongly dependent on heating rate. Thermal gradients sufficient to cause microcracking by themselves may not develop below a threshold heating rate which will be a function of the conductivity. Johnson and Gangi (1980) demonstrated that the growth of a macrocrack from anastomosing grain boundary and transgranular microcracks was dependent on thermal gradients imposed in the rock. Local thermal gradients near any open air-filled internal cavities should be amplified by the two order of magnitude difference in thermal conductivity between air and minerals. Wong and Brace (1979) discuss the implications of this for crack growth.

MICROCRACK KINEMATICS AND DYNAMICS

Response to normal stresses

Given the existence of natural microcracks in rock and/or laboratory induced microcracks, it is important to understand how they respond to changes in the stress environment. For convenience, we can separate these changes into changes of deviatoric stress and changes of hydrostatic pressure. In this section, we will examine microcrack response to hydrostatic pressure and compressive uniaxial stress applied normal to the major crack axis, while in the next section, crack growth and interaction in rock as a result of general deviatoric stress states will be considered.

Microscopic observations have shown that most microcracks in rock have aspect ratios on the order of 10^{-3} to 10^{-5} (see the section on crack statistics). One must remember that these observations are made either in a vacuum or under atmospheric pressure, and it is fair to inquire about the change of shape and dimension of microcracks under pressure. Calculations of crack parameters such as aspect ratio using a particular crack model and measurements of velocity (Hadley, 1976; Warren, 1977; Cheng and Toksöz, 1979) or differential strain (Feves and Simmons, 1976;

Siegfried and Simmons, 1978; Wang and Simmons, 1978) are compatible with the microscopic observations and indicate that, as pressure increases, crack porosity decreases with most or all dry cracks effectively "closed" by pressures exceeding 100 to 200 MPa. Wang and Simmons (1978), and more recently Kowallis and Wang (1982), present data and observations which suggest that crystalline rock may be essentially crack-free in situ (existing microcracks closed or healed), with stress-relief cracking when samples are brought to the surface accounting for observed open microcracks. Extensive observations of partially filled or completely healed microcracks (Batzle and Simmons, 1976; Richter and Simmons, 1977; Wang and Simmons, 1978; Shirey et al., 1978; Sprunt and Nur, 1979; Padovani et al., 1982; Cox and Etheridge, 1983) indicate, however, that microcracks remain at least partially open and connected at depth until sealed by mineral growth from chemically saturated fluids.

The actual mechanics of closure depend on the crack model chosen; "crack spectra" derived from velocity or compression measurements are non-unique (Mavko and Nur, 1978). The pressure necessary to close a penny-shaped cavity, when applied normal to its axis, is on the order of the elastic modulus of the surrounding material multiplied by the cavity aspect ratio. "Closed", in this case, means reduction of the aspect ratio to zero. This model, or models which assume long, flat cracks which differ insignificantly from an ellipse in cross section, may be useful for modelling stress-induced cracks which are relatively straight and sharp-walled, but will be of less use for modelling grain boundary cracks or partially healed and bridged natural cracks or several intersecting cracks. Attempts to model the mechanics of rough-walled crack closure have been made by Gangi (1978, 1981), Mavko and Nur (1978) and Walsh and Grosenbaugh (1979). Asperities on the crack wall come into contact at various places as the crack closes under pressure. The overall closure rate dependency on pressure appears to be a function of the asperity height distribution chosen. The application of pressure often crushes asperities, leaving irreparable damage (Sprunt and Brace, 1974a), so that on subsequent pressure cycles, new asperity distributions may occur.

In a very interesting and illuminating series of experiments within an SEM, Batzle et al. (1980) were able to actually observe crack closure in response to a uniaxial normal stress. They found that thermally produced cracks with well matched but irregular walls tended to close smoothly, but cracks with rough, mismatched walls suffered only partial closure or closed only after debris lodged inside was crushed. They also found that "at some crack intersections, one fracture would open while another simultaneously closed, depending on their orientations". Shear motion along some intersecting cracks was sometimes necessary for closure of other cracks.

The application or removal of hydrostatic pressure could theoretically cause microcracking through differential compressibility of constituent minerals. One would expect these to be mostly coincident grain boundary cracks. Acoustic emissions during hydrostatic pressure cycling are usually one to two orders of magnitude less frequent than during the application of deviatoric stresses and occur principally on the first cycle (Todd, 1973; C. Sondergeld, pers. commun., 1981). These emissions, however, are most likely a result of crushing of grain boundaries and pre-existing crack asperities and not a result of new cracks.

Perhaps the most important effects of hydrostatic pressure are on microcrack growth and interaction. Briefly, the superposition of hydrostatic pressure upon an existing deviatoric stress field is likely to decrease the range and magnitude of deviatoric stresses concentrated near crack tips as well as increase frictional resistance to shear between crack surfaces in contact. For extensional and shear cracks, this increases energy or stress requirements for propagation and thereby makes crack interaction less probable. Additionally, since hydrostatic pressure may effectively reduce crack connectivity, the migration of chemically active fluids, and crack growth dependent upon such fluids, will be inhibited. The end result of all of this is that individual cracks and crack arrays, once formed, are more stable under pressure. For rock in the brittle regime, this leads to higher fracture strengths and greater dilatant (not necessarily total) strains prior to failure at higher pressures.

Few studies of the effect of pressure on crack growth in rock have been made. In a theoretical paper, François and Wilshaw (1968) indicate that hydrostatic pressure should have no significant effect on crack nucleation involving dislocation mechanisms, yet will inhibit cleavage crack propagation through grains and across grain boundaries. Perkins and Krech (1966) found that the strain energy release rate for cracks propagating in sandstone increases as pressure increased, indicating an increase of resistance to fracture at higher pressures. Hugman and Friedman (1979) noted that the density of microcracks which developed prior to failure in carbonate rocks increased with the confining pressure at which the tests were run. They also observed relatively fewer grain boundary cracks at higher confining pressures. Similar results were found by Wawersik and Brace (1971) for Westerly granite. From post-creep test observations using the SEM, Kranz (1980) noted that the number of cracks and average crack length was greater in granite tested under higher pressures. Wong (1981) found that "samples deformed under higher pressures and temperature tend to have more stress-induced cracks aligned at high angles to the maximum applied stress".

Response to deviatoric stress

The application of a deviatoric stress field resulting from differential loads across the boundaries of a rock mass produces a very complex stress system on a microscopic scale. Microcracks can open, close, grow or stop growing in response to locally-induced deviatoric stresses which can differ significantly in sign, magnitude and direction from the macroscopic stress field. Crack and pore interactions can give information about these local stresses (e.g. Kranz, 1979b). We assume that the principles of fracture mechanics (Lawn and Wilshaw, 1975; Paterson, 1978; Rudnicki,

1980) will apply to isolated, individual microcrack growth. The realities of crack propagation in rock deviate considerably, however, from the usual idealizations (Swan, 1975).

Without considering the atomistic details, a crack can be classified kinematically as extensional or shear depending on whether the crack walls very close to the crack tip or edge move, respectively, perpendicular or parallel to the instantaneous plane of propagation. The vast majority of observed microcracks appear to be extensional. This statement should be qualified by the fact that observations are made on essentially planar surfaces after all stresses have been removed. Additionally, it may be that any shear displacements which have been preserved are too small to be readily detected.

In an elastic continuum subjected to deviatoric stresses, the magnitude of the stress concentration around a crack is inversely proportional to the square root of the distance from the crack tip. The proportionality constant is a function of the azimuth to the crack plane and direction of the maximum principal stress. If the applied stresses are tensional, the crack extension force increases with crack length, accelerating the crack forward. Under compressive stresses, the crack extension force can diminish as the crack propagates so that the crack reaches a stable position and stops unless additional stresses are applied.

Either fracture toughness or strain energy release rate can be used to quantify resistance to further crack propagation through minerals or rock (Lawn and Wilshaw, 1975). Measurements of these by a large number of investigators using various methods are tabulated by Atkinson et al. (1979). Values of strain energy release rates for minerals are in the range 0.1 to 10 J m^{-2}, while for rocks, these values tend to be one to two orders of magnitude higher. The difference is most likely due to the fact that a small, but macroscopic crack propagating through rock has an apparent surface area much less than the real area of new surface created as it propagates through and around grains (Friedman et al., 1972). Additionally, new microscopic cracks adjacent to the main crack, but not part of the measured length, may be created increasing the real new surface area (Hoagland et al., 1973). Few of these important measurements exist at in-situ conditions.

Crack paths

Generally, the crack will tend to follow the local maximum stress trajectory, bending out of its initial plane, if necessary, to minimize shear stress effects and maximize the strain energy release rate. The presence of cleavage planes, pores, other cracks and inclusions will affect the crack propagation path by modifying the stress field in their vicinity or by introducing an anisotropy in local cohesive strength. Furthermore, the material discontinuity at a grain boundary can serve to arrest or deflect a propagating crack. In fine-grained ceramics, Wu et al. (1978) noticed "multiple twisting along the crack front due to interaction with the grain structure". Examples of these effects in rock and, in particular, the propagation by en echelon stepping from one cleavage plane to another, may be found in Bombolakis (1973, figs. 10 and 11), Gallagher et al. (1974, figs. 31 and 32), Montoto (1974, fig. 2), Martin and Durham (1975, fig. 3), Mosher et al. (1975, fig. 4), Hamil and Sriruang (1976, figs. 9 and 10), Simmons and Richter (1976, fig. 9), Kranz (1979b, figs. 7, 8 and 11), and Dunning et al. (1980, fig. 4).

On a somewhat larger scale (say several grain diameters), cracks may be partly transgranular and partly coincident with grain boundary segments. This seems to depend on the rock type and composition, with favorably oriented pre-existing planes of weakness a possible controlling factor. Hoagland et al. (1973) found the macrocrack path in oolitic limestone to be highly irregular with most of the path contained in the matrix, but some through the oolites. Similar results were found by Olsson (1974). In sandstone, Hoagland et al. (1973) found most of the crack path in the calcite cement. Gallagher et al. (1974) found mostly transgranular cracking connected at point contacts between poorly cemented grains in sandstone. Hamil and Sriruang (1976) found cracks propagated mostly along the grain boundaries in sedimentary rock but in crystalline rock transgranular paths were most frequent and sometimes dominant. In Westerly granite, Friedman et al. (1970) found "over 75% of the incipient shear fracture is parallel to cleavage planes in feldspar or to grain boundaries or both". Mosher et al. (1975) found, in granite under tension, most cracking was intergranular grain boundary with connecting cleavage or pre-existing cracks, while in granite under compression, discontinuous transgranular cracks dominated. In Carrara marble, Atkinson (1979b) observed that intragranular and transgranular cracks were strongly influenced by calcite cleavage directions. Swanson and Spetzler (1979) found a higher incidence of intergranular fractures in rock loaded at very slow rates compared with rock loaded at fast rates where transgranular fracturing was enhanced (P. Swanson, pers. commun., 1980). Finally, it seems the degree of intracrystalline crack bifurcation is influenced (at least in quartz) by the chemistry of the environment (Dunning and Petrovsky, 1982).

Since the rock and the local stress fields are generally heterogeneous, continuum approaches to crack propagation in rock on the scale of interest here may not meet with desired success, although Ingraffea and Heuze (1980) have had some success combining finite element techniques with fracture mechanics theory. Unless some sort of statistical approach is used, one may only get, at best, the trend of *macroscopic* crack paths.

Crack interaction and coalescence

Fracture in rock under compressive boundary stresses is a result of the growth, interaction and coalescence of many microcracks, not the growth of a single crack. Understanding crack interaction and coalescence is therefore essential to understanding the formation of fracture in rock. Early model studies of crack arrays (Bombolakis, 1964, 1968, 1973; Hoek and Bieniawski, 1965) indicated that certain array configurations and orientations are more favorable for coalescence than others

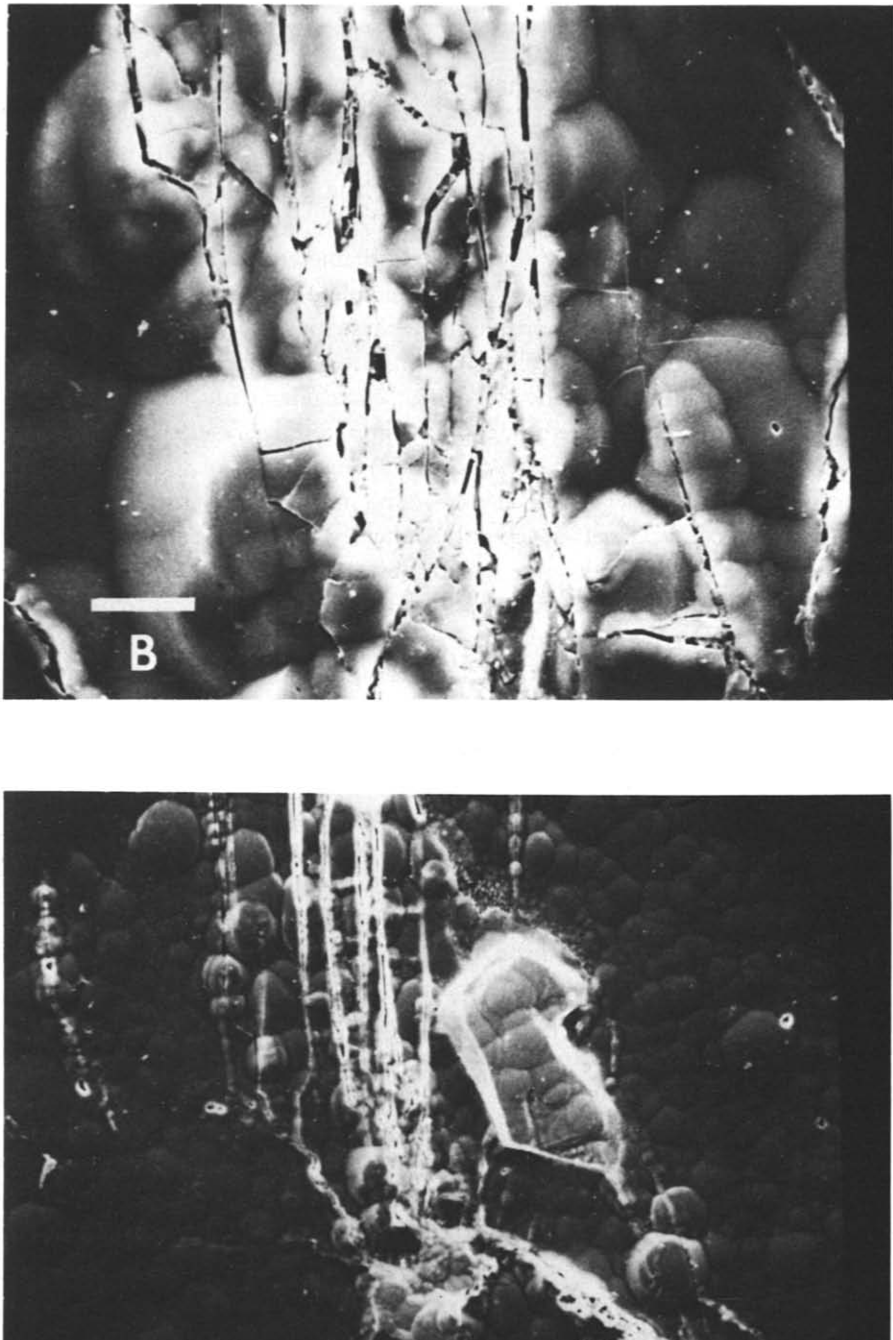

Fig. 4. Localization of microcracks and buckled columns in granite. Maximum applied stress on remote sample boundaries was in vertical direction. (a) Scale bar is 350 μm. (b) Photomicrograph from center of (a); scale bar is 25 μm.

and that coalescence will not occur when neighboring cracks are more than a few crack lengths apart. Interaction of pairs of cracks have been studied with models (e.g. Lange, 1968; Swain and Hagan, 1978) and have been observed microscopically (Kranz, 1979b). The amount of crack overlap with respect to the maximum principal stress seems to be a major factor in determining whether and how cracks link. These observations are not hard to understand in the light of previous discussions about the zone of stress concentrations around cracks.

More quantitative predictions of crack interactions are possible using the theoretical analysis of Segall and Pollard (1980) or Dey and Wang (1981). They have considered the mutual crack boundary stress perturbations of arbitrarily oriented cracks in an elastic continuum under deviatoric stresses. Large concentrated stresses in the regions between cracks are the source of extensional links, which may run from one crack tip to another or from a crack tip into a nearby crack wall.

Crack localization

A large body of observational evidence now exists which indicates that as deviatoric stress is raised on a rock sample, the microcrack population is initially random spatially, becoming more and more localized and intense as the failure stress is approached. Microcrack population studies in limestone (Olsson, 1974), sandstone (Hoshino and Koide, 1972; Sangha et al., 1974), granite (Peng and Johnson, 1972; Tapponnier and Brace, 1976; Kranz, 1979a), gabbro (Rong et al., 1979) and quartzite (Hallbauer et al., 1973) support this, as well as indirect measurements of surface strain (Liu and Livanos, 1976; Sobolev et al., 1978; Spetzler et al., 1981) and acoustic emissions (Sondergeld and Estey, 1981). The localization may accompany the formation of columns of rock material bounded by long subparallel cracks which buckle as stress is increased (Fig. 4) (Peng and Johnson, 1972; Wong, 1982).

Fault formation in compression is apparently accomplished by the linking up of locally dense crack regions, crack arrays and grain boundary cracks. Post-failure studies of microcracks associated with the fracture (Friedman et al., 1970; Hoagland et al., 1973; Dunn et al., 1973; Hamil and Sriruang, 1976; Tullis and Yund, 1977; Dengler, 1979; Wong, 1980, 1982) indicate the density is very high in the fault region and dies off rapidly to the background level a few grains away. There is some experimental evidence (Dunn et al., 1973; Tullis and Yund, 1977; Wong, 1980) that gouge found in the fault zone originated from shear movements within these pre-failure crack localizations.

Slow crack growth

Crack velocities as low as 10^{-7} cm sec^{-1} have been measured in rock or rock minerals (Atkinson, 1979a, 1980; P. Swanson, pers. commun., 1980). Below velocities of about 1 cm sec^{-1}, cracks can be significantly influenced by thermally activated processes which depend on the temperature and chemistry of the environment. In this range, crack propagation is called "subcritical", and the growth of cracks, primarily in silicates, is attributed to stress-aided corrosion processes at the crack tip (see review by Anderson and Grew, 1977).

There have been many studies demonstrating the effects of water (either in vapor or liquid phase) on the subcritical rate of propagation of individual microcracks or small, but macroscopic cracks. These include experiments with synthetic quartz (Atkinson, 1979a; Dunning, 1978; Dunning et al., 1980), natural quartz (Martin and Durham, 1975), micrite and marble (Henry et al., 1977), andesite and basalt (Waza et al., 1980), novaculite (Atkinson, 1980), granite (Swanson, 1981; Sano, 1981) and gabbro (Atkinson and Rawlings, 1981). In these studies, the crack velocity was shown to be either an exponential or power function of the applied stress and greater by one to several orders of magnitude in the presence of water than under dry conditions (Fig. 5). The crack velocity at a given applied stress has also been shown to be a strong function of the direction of propagation (Henry et al., 1977; Atkinson, 1979a) and as expected for a thermally activated process, a strong function of temperature (Meredith and Atkinson, 1981).

Current theory indicates that two of the most important factors controlling subcritical crack propagation velocities are the rate at which corrosion reactions proceed in the vicinity of the crack tip and the rate at which corrosive agents such as

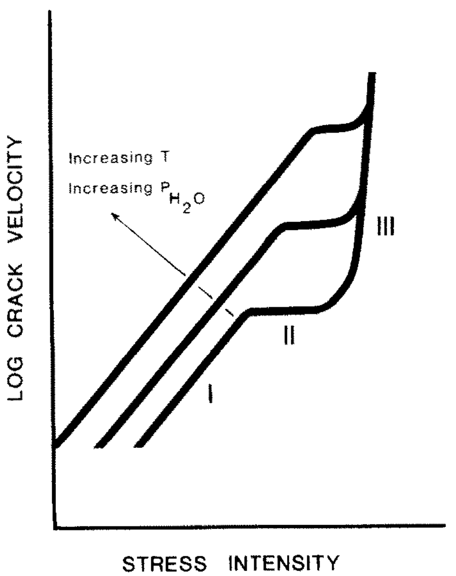

Fig. 5. Typical subcritical crack velocity vs. stress intensity factor diagram. Stress intensity axis is sometimes plotted in log scale, sometimes linear scale. Stress intensity is proportional to applied stress. Virtually all data for rock comes from Mode I, tensile stress loads. Region I is though to be dominated by chemical reaction-controlled kinetics, region II by diffusion-controlled kinetics and region III by a mixture of pure mechanical and chemical corrosion processes (Anderson and Grew, 1977).

water molecules can be brought to the crack tip area. Which of the two is the prime rate-controlling factor may depend on the pressure, temperature and stress environment, as well as rock mineralogy and porosity. Data collected by Henry et al. (1977), Atkinson (1979a, 1980), Waza et al. (1980) and Sano (1981) in region I (Fig. 5) indicate that at low crack velocities the chemical reaction rate is the prime factor. Work by Martin and Durham (1975), Dunning et al. (1980) and Henry et al. (1977) suggest that at higher velocities the availability of corrosive agents at the crack tip may limit the crack velocity. Dunning et al. (1980) indicate that slow growth may, in fact, be episodic with rapid advances made intermittently during the overall slow propagation. This might occur if the initial crack advance is faster than corrosive agents can diffuse back to the crack tip. This is a research topic that deserves much more attention.

MICROCRACK STATISTICS

Observational studies

The collection of data about microcrack populations is long and tedious work. Nevertheless, aside from observations on individual cracks, such data are of utmost importance for testing theories of fracture formation and understanding mechanical property changes in rock. Substantial differences in data collection methods, resolution, precision, data type and presentation exist in the literature and make all but generalizations difficult. Additionally, since cracks are three-dimensional, data from plane sections may be biased. Table I is a summary of recent observational studies, from which I present below the more significant generalizations possible.

Orientations

The traces of microcracks seen in plan view are more or less randomly oriented in most natural, unstressed rock. Some rocks have a significant nonrandom component of microcrack orientations which often accounts for anisotropic mechanical properties and from which it may be inferred they have been subjected to tectonic stress, since microcracks are less randomly oriented in stressed or mechanically deformed rock. Orientations are predominantly within 10° of the applied load in uniaxially stressed rock and oriented within 30° of the maximum compressive stress in rock deformed in a triaxial stress state. There appears to be a tendency for more of the observed trace orientations to be at higher angles to the maximum applied stress direction at higher confining pressures. Under stress, however, the dominant orientation of microcrack traces may vary from region to region within a sample, reflecting local variations in the stress field. Overall, a Gaussian distribution of orientations about some mean angle may approximate the observations, but the mean and standard deviation may change as failure is approached. Nonrandom, "preferred orientations", may also reflect initial fabric which often confuses inferences about applied stresses.

TABLE I

Summary of recent observational studies with microcrack statistical data

Reference	Rock type	Data type	Method
Wawersik and Brace (1971)	Westerly granite Frederick diabase	orientation histograms number, angles vs. confining pressure	thin and polished sections
Peng and Johnson (1972)	Chelmsford granite	density maps, length and orientation histograms vs. applied stress	thin sections
Swolfs (1972)	Coconino sandstone	histograms of angle between normals to cracks and c-axis of host grain	thin sections
Hallbauer et al. (1973)	Witwatersrand quartzite	density maps vs. applied stress	thin sections
Sangha et al. (1974)	Laurencekirk sandstone	length, density distributions vs. applied stress; numbers, orientations vs. strain rate	polished sections
Olsson (1974)	Crown Point limestone	orientation, frequency histograms; density maps	thin sections
Sprunt and Brace (1974b)	Westerly granite	histograms of length, aspect ratio, orientation; stressed, unstressed	SEM
Simmons et al. (1975)	Westerly granite Chelmsford granite Rutland quartzite	length, number, orientation distributions	test array over thin sections
Mosher et al. (1975)	Maine granite	histograms of crack numbers separated by mineral type, crack type; pre- and post-loading	photomicrographs of thin sections
Conrad and Friedman (1976)	Tennessee sandstone Coconino sandstone	microscopic feather fractures: orientations and numbers	thin sections
Tapponnier and Brace (1976)	Westerly granite	mean crack densities, lengths vs. applied stress	grid over SEM photomosaics
Hadley (1976)	Westerly granite	lengths, aspect ratios, orientations; stressed, unstressed	SEM photomicrographs
Brace (1977)	Westerly granite Sherman granite	length, width, number distributions	SEM photomicrographs
Rong et al. (1979)	Gabbro	length, histograms vs. stress, sample position	nitrocellulose peel of polished section

TABLE I (continued)

Reference	Rock type	Data type	Method
Hugman and Friedman (1979)	Various carbonates	crack type distributions	thin sections
Dengler (1979)	Graywacke sandstones	numbers, densities, and orientation distributions	thin sections, SEM
Kranz (1979a, 1980)	Barre granite	length, width, aspect ratio, orientation histograms	SEM
Wong (1980)	Westerly granite	length/area as a function of orientation	grid overlay on SEM photomicrographs
Spetzler et al. (1981)	Pyrophillite	length, orientation histograms	SEM
Hadizadeh and Rutter (1982)	Quartzite	length, orientation histograms	polished sections and SEM
Friedman et al. (1982)	Charcoal granodiorite	densities vs. temperature	thin sections

Lengths, widths and aspect ratios

Neglecting resolution and precision differences and neglecting the facts that cracks may obliquely intersect the section viewed or that the width (w) may vary along the length (l) of the crack, some observations are common and clear. The average crack length increases, and the average aspect ratio (w/l) decreases as the maximum applied stress to which the sample has been subject increases. Furthermore, the entire length spectrum shifts to longer cracks being more frequently observed in stressed versus unstressed rock. Significant average length changes do not appear until stresses greater than about 50% of the peak stress have been applied. Crack widths do not appear to change very much so that changes in aspect ratio are dominated by length changes. Aspect ratios observed, and inferred from velocity measurements, typically are 10^{-3} to 10^{-5}. It is important to remember, however, that almost all observational data are collected after stresses have been removed. Aspect ratios calculated from velocity measurements under pressure are model-dependent.

Length distributions for three unstressed granites are given in Fig. 6. It is not possible to say specifically what type of distribution function best approximates all the data, but it is this type of data which is now most needed to test the common assumption of lognormal or exponential size distributions. Differences in the mean lengths are probably a function of grain size, but may also reflect rock history and orientation of the section.

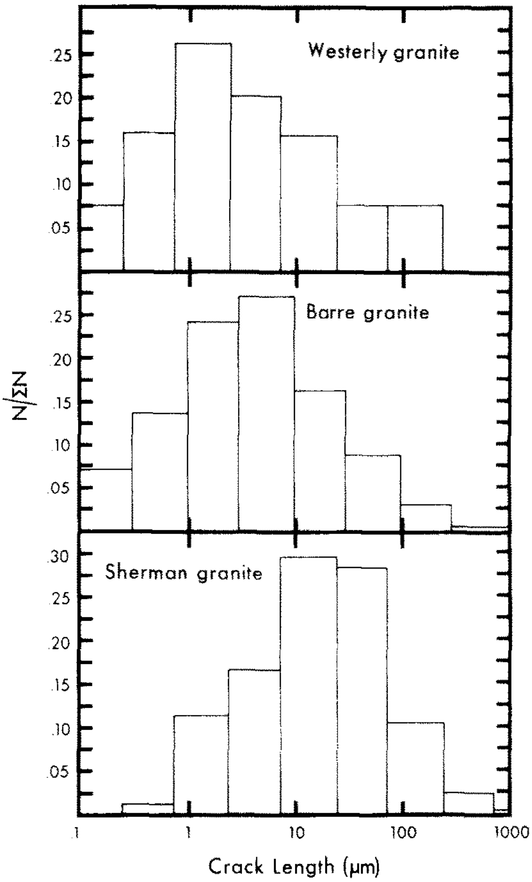

Fig. 6. Intracrystalline microcrack length probability density distributions for fine-grained (Westerly), medium-grained (Barre) and medium to coarse-grained (Sherman) granite. Westerly and Sherman data from Brace (1977).

Numbers and densities

Crack densities (either number of cracks per unit area or per grain) are more enlightening because of resolution constraints than the mere reporting of numbers of cracks. As noted previously, densities increase and appear to change from spatially random to nonrandom higher density regions as stress is increased and failure is approached. Significant high density localizations may not appear until within a few percent of the peak stress. The ratio of grain boundary cracks to intragranular cracks decreases at higher applied loads. The number of cracks generated prior to failure is greater at higher confining pressure, but for rock containing mostly calcite, cracking may be suppressed (in favor of twinning) at higher pressures. Densities decrease rapidly, reaching the background level within a few grains away from a fault. Some investigators find higher densities in particular mineral components; usually the stiffest mineral component contains a greater percentage of the total number of cracks, but susceptibility to cleavage cracking is also a factor. Densities, like orientations, may reflect a heterogeneous stress field in the sample resulting from non-uniform loading conditions.

Theoretical statistical studies

In this section, we touch briefly on theoretical statistical studies of crack nucleation and growth and the role of microcrack populations in the statistics of the fracture strength of rock. It is unfortunate that while a great deal of theory exists, there are not enough of the necessary observational data. Much of the theory has been formulated with metals or simple composites in mind, and the results obtained depend too heavily on assumptions not applicable to rock. The use of "weakest link" models, for example, would certainly not be appropriate for rock where a single "Griffith crack" will not cause failure in compression, even though such models may lead to predictions qualitatively in agreement with the data.

All statistical models have assumptions. For rock, acceptable assumptions include randomness in orientations and location of the initial population of cracks and an initial length or size frequency distribution which is either lognormal or exponential. Stochastic independence between cracks or volume partitions is less and less acceptable as failure is approached; it can be accepted below perhaps 90% of the peak stress or prior to the onset of tertiary creep. That is, assumptions of randomness in the microcracking process may be expedient, but we must acknowledge that there is a stress history dependence to rock failure.

Scholz (1968a) was one of the first to consider the failure of rock from a statistical viewpoint. His model assumes that local stresses in rock are random but constant over small volumetric partitions, which also have a specific strength. The distribution function for the local stress changes as the mean, applied stress changes. When the local strength is exceeded by the local stress, the region fails, a crack propagates and may be arrested if it enters a region of higher strength. In a subsequent paper (Scholz, 1968b), time and frequency of cracking were introduced into the model.

Vere-Jones (1976, 1977) considered the length distribution of pre-existing flaws and branches of daughter cracks emanating from them. He derived the probability of crack coalescence as stress increases and the length distribution changes. The probability of coalescence is taken to be a function of the number of cracks and their lengths. The rock approaches failure as this probability approaches unity. Vere-Jones assumes that the total length of cracks is proportional to the energy released during coalescence, and this leads him to conclude that the distribution function of lengths must follow a power law form to be compatible with the Gutenberg–Richter energy–magnitude relation for earthquakes.

Adams and Sines (1978a) proposed a model similar to Scholz's, but one which is more physically tangible. They considered the frequency distribution of initial flaw size to determine the distribution of local strengths through the application of Griffith's theory of cracking. The flaw orientation and changes in crack density are also explicitly taken into account.

Dienes (1978) assumes that the distribution function for crack size is exponential, that they are isotropically distributed in the rock, and that the population can be separated into growing and non-growing cracks. Growing cracks can be stopped by intersecting non-growing (grain boundary?) cracks. He works out the crack distribution as a function of time during rapid loading by assuming a constant average crack growth rate.

While not specifically directed at rock, two other models are of interest because they consider the relationship between crack linking and grain structure. Lindborg (1969) assumed individual cracks are nucleated independently at random sites and are transgranular. He determined that the fraction of cracked grains necessary for failure depends inversely on the total number of grains in the sample and directly on the average number of nearest neighbor grains (in plane section). For an average six nearest-neighbor grains, 20% of all grains must be cracked before enough cracks link to traverse the sample. McClintock and Mayson (1976) consider only hexagonal grains but with different orientations with respect to biaxial loads. They assume an exponential distribution of tensile grain boundary strengths and, in a numerical simulation, examine the linking of grain boundary cracks as loads are increased. Failure occurs when links bridge the sample.

Taking into account all that has been presented previously about microcrack growth and interaction, each of the above models contains some attractive and realistic features. It may be possible to combine these in a more sophisticated model.

FUTURE DIRECTIONS

The study of microcracks and fracture in rock will and should continue to benefit from the application of fracture mechanics concepts, theories and experimental techniques. Some caution should be exercised, however, in directly transferring this body of knowledge to rock which constitutes a vastly more complicated system than the underlying tenets of fracture mechanics. Awareness of the nature of the complexities will improve through two currently dynamic research areas: slow crack growth and acoustic emissions from stressed rock.

Most of the data on slow crack growth have been generated in the last 5 years. We have fast come to realize the importance of chemochemical processes in the Earth. Unfortunately, knowledge about the atomistic details of what goes on at the crack tip in a corrosive environment is still in its infancy. We can anticipate that increased interaction between geochemists and rock mechanicians will be stimulated by this problem which also has industrial significance. The United States Geological Survey sponsored a workshop in 1982 specifically dealing with the chemical role of water in crustal deformation. Papers from this workshop are to be published this year in the Journal of Geophysical Research.

When crack growth occurs, elastic waves are generated. There is still some controversy over whether or not slow growing cracks emit acoustically. We anticipate that if there is a threshold velocity below which emissions do not occur, or even if there is a threshold stress below which some crack growth does not occur, it will be dependent upon mineralogic and environmental parameters which need to be elucidated.

The study of acoustic emissions has kept pace with technological advances in electronics. It is now possible to locate thousands of acoustic emission "events" in a rock as it is being stressed. Location accuracy is at present limited to about a millimeter and it may not be possible to improve much on that (C. Sondergeld, pers. commun., 1981). More exciting is the current work on microseismic analysis of the acoustic emission signals (Granryd et al., 1980; Sondergeld and Estey, 1981, 1982; Maeda, 1981). Acoustic emissions from rock are being subjected to the same type of analyses as earthquakes have been in the hope of better understanding source mechanisms.

Finally, we anticipate many more studies of crack propagation and population statistics in rock naturally or experimentally subjected to mid-crustal conditions. These studies should increase our knowledge of the factors governing the brittle-ductile transition which will be intimately connected with poorly known processes such as crack tip blunting by dislocation mobility and the role of pressure on crack stability. As in the past, there will be no substitute for microscopic observations, yet we anticipate that computer-aided image analysis and pattern recognition techniques will also be employed. Such techniques will help bridge the gap between grain scale and outcrop scale statistics, thus increasing our understanding of the relationships between microcracking and joint and fault development.

APPENDIX—OBSERVATIONAL TECHNIQUES

The most common enhancement method for observing microcracks in rock is to inject the sample with dye penetrants prior to making a thin section for petrographic microscope work. The gentlest way to do this is to place the sample in a vacuum, then immerse it in the dye solution prior to releasing the vacuum. This, of course, will decorate only the interconnected cracks. To overcome fluid viscosity problems and shorten the time necessary for penetration, the sample could be placed unjacketed in a pressure vessel and subjected to a small hydrostatic pressure with the dye solution acting as the pressure medium. This may, however, damage some cracks. Simmons et al. (1975) describe a method for decorating cracks with the carbon residue from furfuryl acid. This method, however, involves heating the sample which may create new cracks. Simmons and Richter (1976) describe a method for decorating cracks with copper electrochemically.

Polished sections viewed obliquely have been used. Friedman and Johnson (1978) vacuum deposited a 100 Å layer of gold-palladium onto their surfaces prior to viewing in reflected fluorescent light. Kobayashi and Fourney (1978) used acetylcellulose replicating film on polished surfaces. The film can be removed intact from the surface for viewing in reflected light or subsequently coated for scanning electron microscopy.

Radiography (Potter, 1978; Wu et al., 1978) and X-rays (Nelson and Wang, 1977) have been used but with less than satisfactory resolution.

The scanning electron microscope (SEM) and transmission electron microscope (TEM) require more elaborate sample preparation, but the results are worth the trouble. "Thick sections" of 100 Å are also being used for study with either the petrographic microscope or the SEM. Sprunt and Brace (1974b) describe the now standard technique of grinding, polishing and then ion-milling sample surfaces prior to coating and viewing with the SEM. Dengler (1979) points out that gold-palladium coatings interfere with X-ray energy dispersive analysis in the SEM (the palladium peak is close to a potassium peak) so that if this analysis is anticipated, gold coatings are preferred. Carbon coating may tend to crack and deposit unevenly, so should not be used. Preparation of samples for the TEM by ion-milling is described by Barber (1970). These samples are the most delicate and good results require a good, stable ion-mill and some experience.

Crack measurements are often made from photomicrograph mosaics and statistics gathered with the aid of a superimposed grid. The grid should contain lines in at least two orthogonal directions to minimize azimuthal biases. One disadvantage to photomicrograph mosaic use is the large number of pictures necessary to cover a significant area with a sufficient resolution. I have worked directly off the SEM screen which has the advantage of being able to immediately check measurements and relationships at various magnifications, but the disadvantage of requiring impractically long periods of time on the microscope.

In the future we will likely see computer-aided image analysis routinely applied to the study of microcracks in rock. Hand-digitized crack traces (e.g. Warren, 1982) and, hopefully, automatically digitized images will greatly aid statistical studies.

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