DOE/ER/13228--T2

DE89 001780

A FINAL REPORT

FROM

TEXAS A & M
RESEARCH FOUNDATION

ON BEHALF OF

TEES

TEXAS ENGINEERING EXPERIMENT STATION
TEXAS A & M UNIVERSITY
COLLEGE STATION TEXAS 77843

MASTER

DISCLAIMER

This report was prepared as an account of work sponsored by an agency of the United States Government. Neither the United States Government nor any agency Thereof, nor any of their employees, makes any warranty, express or implied, or assumes any legal liability or responsibility for the accuracy, completeness, or usefulness of any information, apparatus, product, or process disclosed, or represents that its use would not infringe privately owned rights. Reference herein to any specific commercial product, process, or service by trade name, trademark, manufacturer, or otherwise does not necessarily constitute or imply its endorsement, recommendation, or favoring by the United States Government or any agency thereof. The views and opinions of authors expressed herein do not necessarily state or reflect those of the United States Government or any agency thereof.

DISCLAIMER

Portions of this document may be illegible in electronic image products. Images are produced from the best available original document.

MECHANICAL PROPERTIES OF ROCKS AT HIGH TEMPERATURES AND
PRESSURES

FINAL REPORT

BY

M. FRIEDMAN, P.I.

WITH

S. J. BAUER, F. M. CHESTER, J. HANDIN, T. W. HOPKINS, B. JOHNSON
A. K. KRONENBERG, D. MARDON AND J. E. RUSSELL

JULY 27, 1987

DOE OFFICE OF BASIC ENERGY SCIENCES
GRANT NO. DE-FG05-84ER13228

prepared by

CENTER FOR TECTONOPHYSICS
TEXAS A&M UNIVERSITY

for the

TEXAS A&M RESEARCH FOUNDATION
BOX 3578
COLLEGE STATION, TEXAS 77843

and

THE TEXAS ENGINEERING EXPERIMENT STATION

TABLE OF CONTENTS

Section No.

ABSTRACT

1. INTRODUCTION
2. STRENGTH AND MICROFRACTURING OF WESTERLY GRANITE
EXTENDED WET AND DRY AT TEMPERATURES TO 800°C AND
PRESSURES TO 200 MPA
3. EXTENSION EXPERIMENTS ON SIOUX QUARTZITE AND
WESTERLY GRANITE
4. FRACTOGRAPHY AND MICROCRACK DISTRIBUTION IN EXPERI-
MENTALLY EXTENDED SIOUX QUARTZITE: MECHANISM AND
ENERGETICS OF BRITTLE FRACTURE
5. ROCK FRICTION AND THE ANALYSIS OF BOREHOLE STABILITY
6. DEFORMATION EXPERIMENTS ON BIOTITE SINGLE CRYSTALS
7. SUMMARY AND CONCLUSIONS
8. ACKNOWLEDGEMENTS
9. REFERENCES

APPENDIX 1: *

MICROFRACTURING IN WESTERLY GRANITE EXPERIMENTALLY
EXTENDED WET AND DRY AT TEMPERATURES TO
800°C AND PRESSURES TO 200 MPA - MS THESIS, T. W. HOPKINS

** removed and cycled separately*

DISCLAIMER

This report was prepared as an account of work sponsored by an agency of the United States Government. Neither the United States Government nor any agency thereof, nor any of their employees, makes any warranty, express or implied, or assumes any legal liability or responsibility for the accuracy, completeness, or usefulness of any information, apparatus, product, or process disclosed, or represents that its use would not infringe privately owned rights. Reference herein to any specific commercial product, process, or service by trade name, trademark, manufacturer, or otherwise does not necessarily constitute or imply its endorsement, recommendation, or favoring by the United States Government or any agency thereof. The views and opinions of authors expressed herein do not necessarily state or reflect those of the United States Government or any agency thereof.

ABSTRACT

The recovery of geothermal energy requires stable boreholes at elevated temperatures, relatively modest effective pressures in the presence of hot aqueous fluids and differential stress and often in fractured rock masses. Similarly, there are a host of energy related problems such as underground storage of wastes for which a knowledge of rock properties at elevated temperatures is important. With these general problems in view, previously we investigated the influences of effective pressures to 200 MPa and temperatures to partial melting on shortened intact specimens of Charcoal granodiorite, Mt. Hood andesite, Cuerbio basalt, and Newberry rhyolite obsidian, and time effects at these P and T for the Westerly granite. In general, we found that boreholes in most crystalline rocks would be stable to at least 10 km provided the boreholes were pressured and kept cool and the ratio of the far-field horizontal to vertical stress is <1.0 . On the other hand, open holes in wet or dry andesite would likely fail at depths as shallow as 2 km. In addition, we learned that penetration rates during drilling would probably increase with depth (or as a magma body is approached by the drill) in that the rocks would remain brittle and their strengths would decrease as temperature increased. As for time effects, we learned they could be generalized with typical power-law flow-equations even though the rocks were brittle and failed primarily by brittle fracture. Accordingly, the predicted strength degradations due to time effects would be minor over the probable life of geothermal wells.

During this final year of the grant, we have investigated (a) why the strengths of rocks decrease with increasing temperature and in the

presence of water through study of the fracture process in Westerly granite and Sioux quartzite specimens deformed in extension (some in true tension), (b) frictional strengths of rocks at high temperatures, (c) the stability of boreholes in fractured rock, and (d) slip in biotite single crystals (in that biotite is probably the weakest and most ductile of the common constituents of crystalline rocks).

Principal Results 1986-87

Westerly granite in extension. Strength decreases markedly, wet and dry, between 600° and 800°C, with water much enhancing the temperature effect. Two distinctive patterns of intragranular and grain boundary cracks occur and systematically correlate with temperature and stress state at failure. Thermal-intragranular and grain-boundary cracking (the latter particularly conspicuous in wet specimens) are primarily responsible for the weakening. That is, the coalescence of precursive microfractures into macro-fractures (tensile or shear) is facilitated by thermal cracks.

Sioux quartzite in extension. Up to 500°C, strengths are insensitive to temperature and strain rate, as expected. Non-linear elastic behavior occurs at 500°C with a sharp increase in inelastic work from 0.22J at $T = 25^\circ - 412^\circ\text{C}$ to 1.15J at 500°C. Macroscopically, fracture surfaces are comprised of smooth (mirror) and rough (hackle) textures and indicate a consistent direction of overall fracture propagation across a specimen. Microscopically, however, propagation directions are randomly oriented as inferred from grain-scale fracture-surface markings.

Frictional strength of rock surfaces at temperature. Review of the literature and some new data indicate that although the effects of chemically reactive fluids are not yet well understood, it appears that Byerlee's empirical relation between frictional strength (shearing resistance) and normal stress is a good enough approximation for igneous rocks, under conditions of concern to the borehole stability problem, regardless of composition, temperature, and loading rate. The only certainly significant exception is that the phyllosilicates are much weaker, with μ as low as 0.15 in the swelling clays. Joints in typical igneous rocks may be filled with alteration products such as chlorite, and for natural fault gouges, μ may be less than 0.5.

Borehole stability in fractured rock. For a reasonable set of assumptions commonly made in engineering rock mechanics dealing with a vertical borehole cut by a fracture dipping β degrees, we find that for $\beta < 90^\circ$, stability depends on two parameters, the coefficient of sliding friction, μ , and the ratio, R , of mean normal stress, $(\sigma_1 + \sigma_3)/2$, to maximum shear stress, $(\sigma_1 - \sigma_3)/2$. Results include: (a) for a constant value of μ , the unstable range of β decreases and ultimately disappears as R increases; (b) R decreases as pore pressure increases, which leads to an increase in the size of the unstable range of the dip angle β ; conversely, decreases in pore pressure tend to stabilize the fracture and the borehole; and (c) increasing borehole pressure (above pore pressure) decreases R which in turn increases the size of the unstable range of β and thus has a destabilizing effect. The case for a vertical fracture at a vertical borehole also is explored.

Biotite single crystals. Single crystals oriented for basal slip deform by dislocation glide at confining pressures as low as 100 MPa.

Transitional behavior between frictional sliding and slip occurs at about 50 MPa. Within the ductile (high pressure) field, crystals oriented for slip along $t = [010]$ and $[310]$, while stronger, exhibit the same temperature dependence ($Q = 33\text{kJ/mol}$) as those oriented for $t = [100]$ and $[110]$ ($Q = 30\text{kJ/mol}$).

1. INTRODUCTION

This is the final report for the three-year, DOE/OBES Grant No. DE-FG05-84ER13228. It contains a summary of past work dating back to 1974 on the mechanical behavior of crystalline rocks at high temperatures and pressures with a view toward understanding borehole stability in geothermal regimes. In addition, it covers in detail activities on the project during the third and last year.

Following the introduction, which includes the historical summary, results of current work are presented proceeding from (a) investigation of rock strength and the fracture process in extension tests designed to gain better insight into why rock-strength decreases with increasing temperature and in the presence of water (Sections 2-4), through (b) evaluation of rock friction at temperature and a corresponding analysis of borehole stability involving holes in fractured rock (Section 5), to (c) study of the ductile strengths and mechanisms of biotite single crystals, (Section 6). Biotite is important to the prediction of rock behavior in that it is the most ductile of the mineral species in typical crystalline rocks. As such, its gliding mechanisms compete with brittle fracture in the other mineral species under P and T conditions likely to exist about boreholes in geothermal wells. These presentations of data and discussion are followed by sets of comprehensive Conclusions (Section 7), Acknowledgements (Section 8) and References (Section 9). Section 2 is now a published work based on the MS thesis of Mr. T. W. Hopkins. The latter contains data not utilized in the publication and is therefore included here as APPENDIX 1.

RESEARCH SCOPE AND HISTORICAL REVIEW

Research on the mechanical properties of rocks at moderate pressures (<200 MPa) and temperatures to partial melting (800° - 1000°C) has been carried out by Drs. M. Friedman and J. Handin and their associates since 1974, first through a low-level contract with Sandia National Laboratories (Contract No. 13-2242, 1974-79), and then through direct support by DOE/OBES (Contract No. DE-AS05-79ER10361, 03/01/80 to 02/29/84, Task 3), and Grant No. DE-FG05-84ER13228, 03/01/84 to 02/28/87). This research program was initiated as part of the effort to evaluate the scientific feasibility

of energy extraction from buried magma chambers. The main thrust of our work relative to this program has been and continues to be the assessment of borehole stability in the aqueous, high-temperature but relatively modest effective-confining-pressure environment above and at the interface with buried magma. However, our results are clearly applicable to other rock-mechanics problems within the mission of the Department of Energy; e.g., the siting of nuclear power plants, toxic waste storage and isolation, other geothermal projects, including hot-dry-rock, and continental scientific drilling program. Collectively our work has been reported in some 20 publications (Table 1). Because most of our experiments have been done by post-doctoral fellows or graduate research assistants whose theses or dissertations were based on other research, our project has directly supported but one Ph.D. dissertation and two M.S. theses (Table 1). However, it has afforded rare opportunities for about a dozen people, including international visiting scientists, to gain valuable experience with high-temperature triaxial-compression testing with the gas confining-medium.

While supported through Sandia National Laboratories our studies were of the short-term strengths and ductilities of both dry and water-saturated igneous rocks (Mt. Hood andesite, "Charcoal" granite, Cuerbio basalt, and Newberry rhyolite obsidian) to effective pressures of 100 MPa and temperatures to partial melting. This phase of the work was completed in 1980 during the first part of our contract with DOE/OBES. The results have strongly suggested that up to temperatures of melting, the crystalline rocks can be drilled by conventional methods (rocks will remain essentially brittle) and that boreholes in most igneous rocks will be stable to at least 10-km depth, so long as the tools can be cooled, the boreholes can be

TABLE 1
Chronological List of Publications Resulting from Research
Related to the Mechanical Behavior of Rocks at High
Temperatures and Pressures

Contract No. 13-2242, Sandia National Laboratories, 1974 - 1979

- Handin, J. and M. Friedman, High-temperature strengths of andesite and granodiorite as related to geothermal-energy recovery from magma (Abstract). Am. Nuclear Soc., Topical Meeting on Energy and Mineral Research Recovery, Colorado School of Mines, 42, 1977.
- Friedman, M., J. Handin, and N. C. Higgs, Deformation of dry andesite and granodiorite at temperatures to partial melting and pressures to 50 MPa (Abstract). 144th Nat. Meeting Am. Assoc. Adv. Sci., Washington, D.C., 110, 1978.
- Friedman, M. and B. Johnson, Thermal cracks in unconfined Sioux quartzite. Proc. 19th U.S. Symp. Rock Mechanics, U. Nevada-Reno, 1, 423-430, 1978.
- Bauer, S. J. and B. Johnson, Effect of slow uniform heating on the physical properties of Westerly and Charcoal Granite. Proc. 20th U. S. Symp. Rock Mechanics, U. Texas, Austin, 7-18, 1979.
- Friedman, M., J. Handin, N. G. Higgs, and J. R. Lantz, Strength and ductility of four dry igneous rocks at low pressures and temperatures to partial melting, Proc. 20th U. S. Symp. Rock Mechanics, U. Texas, Austin, 35-49, 1979.
- Bauer, S. J., The effects of unconfined slow uniform heating on the mechanical and transport properties of the Westerly and Charcoal granites, M.S. Thesis, 106 pp., Texas A&M U., 1980.
- Friedman, M., J. Handin, N. G. Higgs, J. R. Lantz, S. K. Bauer, Strength and ductility of room-dry and water-saturated igneous rocks at low pressures and temperatures to partial melting. Sandia National Laboratories, Report SANDSO-7159, 1980.

Contract No. DE-AS05-79ER10361 (Task 3), 03/01/80 - 02/29/84

- Handin, J. and N. L. Carter, Rheology of rocks at high temperature. 4th Cong., Intern. Soc. Rock Mechanics, Montreux, Switzerland, 3, 97-106, 1980.
- Johnson, B. and A. F. Gangi, Thermal cracking of nonuniformly heated, thick-walled cylinders of Westerly Granite. Proc. 21st U. S. Symp. Rock Mechanics, U. Missouri-Rolla, 197-206, 1980.
- Bauer, S. J., M. Friedman and J. Handin, Effects of water saturation on strength and ductility of three igneous rocks at effective pressures to 50 MPa and temperatures to partial melting, Proc. 22nd U. S. Symp. Rock Mechanics, MIT, 73-78, 1981.

- Bauer, S. J. and J. Handin, Thermal expansion of three water-saturated igneous rocks to 800°C at effective confining pressures of 5 and 50 MPa (Abstract), Am. Geophys. Union Trans., 62, 393, 1981.
- Bauer, S. J., N. L. Carter and J. M. Logan, Effects of temperature and pore-water pressure on static fatigue of Westerly Granite (Abstract), Am. Geophys. Union Trans., 62, 1025, 1981.
- Handin, J., M. Friedman and S. J. Bauer, Borehole stability and rock-drillability studies (Abstract), Am. Geophys. Union Trans., 62, 1057, 1981.
- Bauer, S. J., Brittle-ductile transition in dry and water-saturated Westerly granite (Abstract), Am. Geophys. Union Trans., 63, 437, 1982.
- Friedman, M., J. Handin and S. J. Bauer, Deformation mechanisms in granodiorite at effective pressures to 100 MPa and temperatures to partial melting, Proc. 23rd U. S. Symp. Rock Mechanics, U. California, Berkeley, 279-289, 1982.
- Bauer, S. J. and J. Handin, Thermal expansion and cracking of three, confined, water-saturated rocks to 800°C, Rocks Mechanics and Rock Engineering, 16, 181-198, 1983.
- Bauer, S. J., Semibrittle deformation of granite at upper crustal conditions, Ph.D. Dissertation, 186 pp., Texas A&M U., 1984.
- Bauer, S. J. and J. Handin, Mechanical properties of basalt cores from DSDP Hole 504B, Initial Reports of the Deep Sea Drilling Project, 83, 371-375, 1985.
- Grant No. DE-FG05-84ER13228 (03/01/84 - present)
- Handin, J. and N. L. Carter, Rheology of rocks, Encyclopedia of Structural Geology and Plate Tectonics, Dowden, Hutchison and Ross, Stroudsburg, Pa., in press.
- Shimamoto, T., The origin of large or great thrust-type earthquakes along subducting plate boundaries, Tectonophysics, 119, 37-65, 1985.
- Johnson, Brann, Mel Friedman, T. W. Hopkins, and S.J. Bauer, Strength and microfracturing of Westerly granite extended wet and dry at temperatures to 800°C and pressures to 200 MPa. Proc. 28th U. S. Symp. Rock Mechanics, University of Arizona, Tucson, 399-412, 1987.

pressurized by the drilling fluid, and the ratio of the far-field horizontal to vertical stress is < 1.0 (Friedman et al., 1979; Bauer et al., 1981; Handin et al., 1981; Bauer, 1982; Friedman et al., 1982). However, as the worst case, open boreholes in wet or dry andesite should fail shortly after drilling to < 2 km if the ratio of horizontal to vertical far-field stresses is 2.0. Ancillary studies by Bauer and Handin (1981, 1983) provide new data on the linear thermal expansions of these rocks to 800°C and 100 MPa effective confining pressure, and the work on the Newberry rhyolite obsidian (Friedman et al., 1979) shows dramatic thermal weakening between 600°C and 800°C (precluding natural glass as a suitable borehole liner) and equivalent viscosities in close agreement with those determined via beam-bending by Murase and McBirney (1973).

We recognized that interpretations of data from tests on intact rock might have to be modified to account for the effects of pre-existing fractures (joints) and transverse anisotropy (e.g., schistosity), and that all conclusions could be affected by duration of loading. Accordingly, during the period 1981-84 we looked into (1) time-dependent effects on strength and strain to failure of intact rock, and (2) temperature effects on the frictional strength of bare-rock surfaces. Furthermore, although expectable borehole stresses favor failure in shear under compression rather than in tension by hydraulic fracturing (Friedman et al., 1979, Table 6), we realized that all fracturing initiates at microscopic flaws (stress-risers) when the local tensile stress exceeds tensile strength, and therefore that the study of the effects of time and temperature on fracture strength in the brittle state should begin with well controlled extension tests in which macroscopic tensile fracturing can be induced.

Time dependencies were assessed by making time-to-failure (static fatigue) tests on Westerly granite at temperatures from 300° to 800°C and at effective confining pressures to 100 MPa (Bauer, 1984). Significantly, we found that the data could be generalized by "power-law" flow equations involving an exponent on the stress term and activation energy. Such flow "laws" were previously thought to be applicable only when the deformation mechanisms involved dislocation motions. In contrast fracture is the dominant deformation mechanism under our test conditions. The intact strength of the Westerly granite is about 20 percent weaker when deformed wet in creep tests than in our short-term experiments. To determine if the assessment of borehole stability needed revision, we extrapolated the times to failure for the most drastic case (800°C). Considering the expectable borehole stresses at 8 and 10 km depth and a horizontal to vertical far-field stress ratio of 1.0, we concluded that a borehole in intact granite would remain stable 10,000 and 1 year, respectively. Thus, our initial assessment based on the short-term tests does not change significantly, at least for intact granite over times of engineering concern.

At the same time we found that the frictional sliding strength of bare-rock surfaces is only about 50 percent of the intact strength. Moreover, the tensile strength of granite drops from about 10 MPa at 25°C to sensibly zero at 800°C. The full impact of these results on borehole stability will be evaluated during the latter part of the current year (see Progress Report on Current Research).

Over most of this time (1981-84), two other research tasks were joined under DOE Contract No. DE-AS05-79ER10361, namely: "The Physical Nature of Fracturing at Depth" (N.L. Carter, P.I.) and "Fracture Permeability of Crystalline Rocks as a Function of Temperature, Pressure and Hydrothermal

Alteration" (Brann Johnson, P.I.). We will not review their work, but in passing we note that Carter and his student F.D. Hansen found that flow equations of a similar power-law-form fit their wet and dry static-fatigue tests on Barre and Westerly granites at 100 MPa effective pressure and 200°C and at 1000 MPa and temperatures to nearly 800°C (Carter et al., 1981; Hansen and Carter, 1983; Hansen, 1982).

These empirical flow laws strongly suggest that fracture, the major deformation mechanism in the static-fatigue experiments, is itself thermally activated. This result, along with the dramatic decrease in ultimate strength with increasing temperature, particularly when water is present, led us to next evaluate the thermal-activation and stress-corrosion aspects of fracture in wet and dry extension tests on quartzite and granite (1983 to present). This work has resulted in the recognition of significant changes in grain-boundary and intragranular microcracking for changes in temperature and wet versus dry conditions (Hopkins, 1986). Most importantly we find that grain boundaries play an ever larger role in the microcracking that precedes macroscopic extension and shear fracturing as the temperature increases, especially in water-wet specimens.

Concurrently we are gathering more data on tensile and frictional strengths of rocks at high temperature. By the end of the current grant (02/28/87), we expect to be able to assess the effects on stability of boreholes of various conditions of far-field stress, temperature, and pore and borehole pressures, using strengths of joints, as well as extrapolated long-term intact-rock strengths.

The logical next step, proposed in detail elsewhere by Drs. Andreas K. Kronenberg and James E. Russell, deals with evaluation of the influence of transverse anisotropy on both the state of stress at the borehole and the

pressure-dependent and pressure-independent strengths of anisotropic rocks. In this effort Dr. Kronenberg will provide the leadership in the experimental work at elevated temperatures and pressures, and Dr. Russell will provide the expertise to generalize the results with the aid of numerical models of the possible failure modes of boreholes in the anisotropic rock mass.

Statements of summaries and conclusions will be presented chronologically in Section 7. The reader interested only in an executive summary might turn to Section 7 at this time.

2. STRENGTH AND MICROFRACTURING OF WESTERLY GRANITE EXTENDED WET AND DRY AT TEMPERATURES TO 800°C AND PRESSURES TO 200 MPa *

ABSTRACT

The strength of Westerly granite deformed in triaxial extension decreases significantly with increasing temperature with a disproportionately greater decrease occurring at the higher temperatures. The effect of temperature is equally important at tensile and compressive stress states. Water significantly enhances the effect of temperature on strength reduction. For a compressive stress state and 800°C (possibly as low as 600°C), strength in triaxial extension is significantly less than that for a comparable triaxial-compression test. Microstructural analyses indicate microfracturing is the dominant deformation at all temperatures. Two distinctive microcrack patterns occur. Microfracture statistics and the two microcrack patterns exhibit systematic relationships with temperature and stress state at failure. Observational data and comparisons with room-temperature strength data of thermally-cycled specimens indicate that the observed decrease of tensile and unconfined compressive strength at temperature primarily reflects the effects of thermal cracks. The decrease of confined compressive strength with increasing temperature correlates with a corresponding increased development of precursive intragranular microcracks. In this case, however, data are inadequate to differentiate uniquely the effects of thermal cracks from that of temperature-enhancement of growth of load-induced precursive microcracks, although an argument can be made that the former plays an important role. Strength and microstructural data indicate that a change of failure mechanism in compression occurs in dry and wet specimens at about 800°C and 700°C, respectively. The rock is markedly weaker and grain-boundary cracks appear to play an important role relative to intragranular cracks.

1 INTRODUCTION

Mechanical and microstructural analysis of igneous rocks deformed in short-term triaxial-compressive tests or in constant-stress creep tests at low effective confining pressure (<200 MPa) but high temperature (to partial melting) indicate that fracture strengths are temperature dependent (Friedman et al., 1979; Bauer et al., 1981; Friedman et al., 1982 and Bauer, 1983). The principal purpose of this work is to gain a better understanding of why, in the brittle regime, rock strength decreases with increasing temperature.

* Published as: Brann Johnson, Mel Friedman, T. W. Hopkins, and S. J. Bauer, 1987, (same title), Proc. 28th U. S. Symposium Rock Mechanics, Tucson, Arizona, p. 399-412.

Work is based on the MS thesis of T. W. Hopkins included here as APPENDIX I.

Previous work demonstrates that macroscopic shear fractures or faults in nature and in laboratory compression tests arise from the coalescence of precursive microscopic extension or tensile fractures (e.g. Brace, 1971; Friedman, 1975; and Kranz, 1983). These precursive microfractures arise from local stress concentrations, which are mechanically or thermomechanically induced, and are oriented perpendicular to the local tension (herein σ_3). At elevated temperatures thermal cracks and grain boundary partings, caused by differential thermal expansions of the grains, add to the precursive microfracture population.

The temperature state potentially may affect microcracking and failure of a crystalline rock in several direct and indirect ways. An increase of temperature can produce: 1) a generation of thermal stresses that may cause a fracture and modification of mechanically-induced differential stresses; 2) a reduction of fracture energy (or fracture toughness); 3) an enhancement of crack-growth rates as a consequence of brittle fracture being a thermally-activated process; and 4) an enhancement of chemically-assisted fracture, such as stress-corrosion fracture. The exact role, however, played by temperature in the decrease of rock strength remains unclear.

The current research effort primarily involves mechanical and microstructural analysis of Westerly granite deformed in triaxial-extension tests. Investigation of specimens deformed in extension provides insight into the fracturing processes at elevated temperatures without complications of post-shear-fracture microcrack contributions that characterize most of the previous work. These tests also allow study of temperature effects when σ_3 is tensile.

2 EXPERIMENTS

Specimens of Westerly granite were cored from a block provided by the U.S. Bureau of Mines from its standard rock suite (Krech et al, 1974). Bauer and Johnson (1979) and Bauer (1983) used the same starting material and provide a comprehensive petrofabric characterization of the undeformed material.

Annealed-copper jacketed, 2 by 4 cm, "dog-bone-shaped" specimens are deformed in triaxial extension ($\sigma_3 < \sigma_2 = \sigma_1 = p_c$), both dry and water-saturated at uniform temperatures between 25°C and 900°C and a confining pressure (p_c) between 20 and 200 MPa. The minimum principal stress, σ_3 , acts parallel to the long axis of the specimen. The dog-bone geometry allows generation of a macroscopic axial tensile stress without complications at the specimen-piston interface. As noted later, stress concentrations associated with the fillets do not appear to influence failure of the specimen.

A description of the apparatus, furnace design and the temperature distribution in the specimen is provided by Friedman et al (1979) and Higgs (1981); the latter also provides additional details on correction for seal friction, jacket strength and data-reduction procedures. Standard experimental procedures for extension tests at temperature are used and are outlined in Hopkins (1986).

Axial force and displacement are measured external to the pressure vessel and are used to calculate differential stress - axial strain curves with appropriate correction made for apparatus distortion, friction at seals and copper-jacket resistance. Correction is not made for deformation-induced change of the cross-sectional area of the

Figure 1. Strength data and associated failure envelopes in σ_1 versus σ_3 principal-stress space. Dashed lines represent failure envelopes determined in triaxial-compression tests. Letter B denotes which specimens exhibited Type-B microcrack pattern.

specimen. Uncertainty of the axial stress, σ_3 , is estimated to be less than 3% of p_c (e.g. 6 MPa at $p_c=200$ MPa) and primarily reflects uncertainty in the correction for seal friction. Temperature of the specimen is measured by a thermocouple contacting the lower end of the specimen. Temperature variations along the length of the specimen are approximately 3 to 4% of the temperature measured at the lower end.

Table 1 and Figure 1 summarize pertinent results of twenty-two extension tests. For each of three temperatures: 25°, 600°, and 800°C, a suite of tests (dry) was conducted at five different confining pressures (20, 50, 100, 150, 200 MPa). Two additional tests were conducted at 700° and 900°C for $p_c=150$ MPa. A preliminary assessment of the effect of water was provided by three tests at 600°, 700°, and 800°C and an effective confining pressure of 100 MPa ($p_p=100$ MPa and $p_c=200$ MPa). Some tests were taken to failure, but others were terminated at or near peak differential stress, $(\sigma_1 - \sigma_3)_p$, but prior to development of a macroscopic fracture.

Strength data for three suites of extension tests exhibit a smooth systematic relationship between σ_1 and σ_3 at failure (Figure 1). There is a significant weakening of the rock with increasing temperature with a disproportionately greater decrease of strength from 600° to 800°C than from 25° to 600°C. Intercepts of the failure envelopes with the σ_1 and σ_3 axes represent, respectively, the biaxial compressive strength ($\sigma_1 = \sigma_2 > \sigma_3 = 0$), C_b , and the uniaxial tensile strength ($\sigma_1 = \sigma_2 = 0 > \sigma_3$), T_0 . For 25°, 600° and 800°C the uniaxial tensile strengths are, respectively, 12, 8, and 3 MPa; similarly, the biaxial compressive strengths are, respectively, 195, 130, and 50 MPa. As indicated, the uniaxial tensile and biaxial compressive

Table 1. Results of Triaxial Extension Experiments for Westerly Granite as a function of temperature.

Exp. No.	T (C°)	$P_c = \sigma_1$ (MPa)	$\sigma_1 - \sigma_3$ (MPa)	σ_3 (MPa)	Axial Strain (%)	Fracture Angle# (°)	Fracture Type##	Micro-Fracture Pattern*
E1	25	20	32.0	-12.0	0.4	?	?	UF
E2	25	50	61.1	-11.1	0.4	3	T	UF
E3	25	100	111.0	-11.0	0.5	3	T	UF
E4	25	150	156.8	-6.8	0.6	3-5	T	UF
E5	25	200	198.4	1.6	0.9	3-5	T	UF
E6	600	20	28.0	-8.0	0.4	3	T	B
E7	600	50	56.3	-6.3	1.0	3	T	B
E8	600	100	102.5	-2.5	1.2	3	T	A
E9	600	150	148.1	1.9	1.0	---	***	A
E10+	600	150	147.1	2.9	1.2	10	S	-
E11	600	200	183.4	16.6	1.2	---	***	A
E12	600	200**	93.6	6.4	1.8	2	T	A
E13	700	150	144.2	5.8	1.3	9	S	A
E14	700	200**	86.8	13.2	2.4	---	T	A/B
E15	800	20	22.8	-2.8	0.9	2	T	B
E16	800	50	50.0	0.0	1.5	3	T	B
E17	800	100	84.1	15.9	1.2	5	?	A
E18	800	150	95.9	54.1	1.0	---	***	B
E19	800	150	87.0	63.0	1.5	9	S	B
E20	800	200	148.2	51.8	1.6	9	S	A
E21	800	200**	44.0	56.8	5.3	---	***	B
E22	900	150	46.9	103.1	1.1	---	***	B

Angle between cylinder axis and normal to fracture surface.

T = Tensile or extensile, S = shear, based on fracture angle.

* UF = few if any microfractures; A and B patterns (see text and Figures 3, 4)

** $P_c = 200$ MPa, P_p (water) = 100 MPa, $P_e = 100$ MPa

*** Test terminated at or near peak stress but prior to macroscopic fracture

+ Specimen not available for observational study

strengths decrease with increasing temperature. Furthermore, these two strengths decrease proportionately as indicated by the nearly constant C_p/T_0 ratio of 16.

The functional form of the failure envelope in the tensile field of σ_3 is best defined by the 25°C data, which compares favorably with that obtained by other workers (e.g. Amadei and Robison, 1986). The room-temperature uniaxial tensile strength of the Westerly granite as determined by these extension tests (12 MPa) compares favorably with tensile strengths determined by other workers (Schock and Louis, 1982; Bauer and Johnson, 1979; Krech et al, 1974). The biaxial compressive strength of 195 MPa is essentially the same as the uniaxial compressive strength (215 MPa) as determined for this starting material by Bauer and Johnson (1979). Furthermore, the failure

envelope for triaxial-compression tests merges with that of the triaxial-extension tests (Figure 1).

There are insufficient data to adequately constrain the slope of the failure envelope of the 600°C data within the compressive field of σ_3 . Thus the slope of the failure envelope (Figure 1) may not be necessarily as shown. This is less of a problem with the 800 °C data, but there are more scatter of the data at the higher compressive stresses. The cause and significance of this scatter are discussed later.

Comparison of results of the wet and dry tests (Figure 1 and Table 1) indicates that at these temperatures (600°-800°C) water promotes significant additional weakening.

Differential-stress versus axial-strain curves exhibit systematic differences with both increasing confining pressure and temperature (Hopkins, 1986). A typical stress-strain curve exhibits an initial linear region followed usually by a markedly nonlinear segment with an ever-decreasing slope with increasing differential stress. For all temperatures, if $\sigma_1 > C_b/2$, the onset of the nonlinearity occurs at about half the peak differential stress. If $\sigma_1 < C_b/2$, the onset of the nonlinearity occurs at a significantly greater relative differential-stress level. In addition, if $\sigma_1 > C_b/2$, there appears to be a characteristic axial strain at failure for each temperature. This strain is relatively insensitive to the peak differential stress (at least over the range of differential stresses of this study) but increases with increasing temperature.

Stress-strain data for wet tests are limited, but significantly larger axial strain at failure occurs relative to a comparable dry test. As with dry tests, the greater the temperature, the greater the strain at peak differential stress. The 800 °C wet test exhibits a marked ductility compared to its dry counterpart.

3 OBSERVATIONS

Observationally of interest are the type, length, density and configuration of microfractures induced by the deformation at temperature. Deformed specimens are impregnated with a low viscosity, room-temperature setting, blue-stained epoxy. Thin sections are made parallel to the long axis of the specimen. By considering only the epoxy-filled microfractures, we eliminate those possibly caused by specimen preparation post-impregnation.

Microfracture maps are constructed by mapping the traces of microcracks on photographic enlargements while observing a given field of view under the microscope. The mapped area of each sample is kept consistent and is determined by the width of the thin section and 4 mm to either side of the main macrofracture. For specimens without a macroscopic fracture, a similar area, representative of the overall deformation, is mapped.

Using these maps and the thin sections themselves, data are obtained on the linear density of intragranular (IGC) and grain-boundary cracks (GBC), GBC and IGC areal densities, and IGC length versus temperature for both wet and dry conditions. Details of the method are given in Hopkins (1986).

None of the macrofractures produced in the experiments occur at the fillets of the dog-boned specimen. Instead, the macrofractures occur within the central neck region at distances of 1 to 10 mm from the fillets; most are 5 to 7 mm distant from the nearest fillet.

In four of the specimens, the macrofractures are inclined to σ_1 at an angle of 9-10° and therefore are considered to be a shear (S) fracture. Other fractures are essentially perpendicular to σ_3 and are interpreted to be tensile or extension (T) fractures (Table 1).

In this study, grain-boundary cracks (GBC) are distinguished from intragranular cracks (IGC) (Figure 2). GBC are coincident with grain boundaries, whereas IGC cut across grains or are contained within a single grain. A crack of either type cannot be specifically attributed to mechanical or thermal stresses. However, Simmons and Richter (1976) among many others, point out that most cracks produced by a mechanical differential stress are parallel or subparallel to the direction of σ_1 and are likely to cross several grains. Bauer and Johnson (1979) reported both IGC and GBC produced in thermal cycling tests. The IGC produced from thermal stresses, however, are usually small cracks that are perpendicular to, and originate from, grain boundaries (Friedman and Johnson, 1978). In quartzites, these cracks tend to taper inward and tend to die out within a short distance from the grain boundary. In granites, thermal cracks also transect the grains. Thus, well developed IGC that continue across grain boundaries and are strongly oriented parallel to σ_1 are most likely to reflect the mechanical stresses. These may have grown, however, from initial thermal IGC, if suitably oriented, with the application of mechanical loads. GBC are more likely the result of thermal stresses (Johnson et al., 1978; and Friedman and Johnson, 1978), but grain boundaries that lie parallel to σ_1 may link up mechanically-induced cracks because the grain boundary may provide less resistance to crack propagation than the interior of a grain.

Maps of IGC and GBC (Figures 3, 4) provide visualization of these features. Additional maps are presented in the work of Hopkins (1986). Two distinct patterns (A and B) occur (Table 1).

Pattern A consists of well developed IGC with a high degree of preferred orientation, perpendicular to the σ_3 direction (e.g. Figure 3a, e, f). They often coalesce with nearby grain boundary cracks of nearly the same orientation to produce an array of parallel microfractures, often extending several grain diameters in length. The grain boundary cracks are developed less extensively and exhibit a weaker preferred orientation compared to the intragranular cracks, but they too are weakly aligned sub-perpendicular to the σ_3 direction (e.g. Figure 4a).

Pattern B contains both types of microfractures, but exhibits only a weak, if any, preferred orientation (e.g. Figures 3d and 4b, c). The grain boundary cracks are well developed, existing along several or all sides of many of the grains to form a random pattern (Figure 4b, c). Intragranular cracks show greater variation of orientation with respect to the σ_3 direction than in pattern A.

The microfracture population of a deformed specimen reflects the effects of both the temperature and stress state at failure. Table 2 and Figure 5 summarize microcrack statistics obtained from the maps (e.g. Figure 3 and 4). Specimens deformed at elevated temperatures have significantly greater numbers of IGC and GBC than those deformed at room temperature. Emphasis here is on microcrack data for 600° to 800°C, wet and dry. For wet or dry tests at a given temperature, there is little variation of the linear crack density (LCD) of GBC with differing peak differential stress. On the average, however, the

Figure 2. Photomicrographs of (a) coalesced intragranular microfractures (IGC) into a strongly oriented network in E12, and (b-c) well developed grain boundary cracks (GBC) in E21. The least principal stress (σ_3) is parallel to the short dimension of each photo. Plane polarized light. Scale line is 1.00mm for (a-b) and 0.20mm for (c).

Figure 3 Maps of intragranular microfractures (IGC) in specimens E12, E14, E21, E15, E23 and E20, (a-f), respectively. The least principal stress (σ_3) is parallel to the short dimension of each map. (a, e, and f) are characteristic IGC of Pattern A, and (b, c and d) of Pattern B (see text). All maps are at the same scale (see beneath map c).

LCD of GBC increases slightly from 600°C to 800°C (Table 2).

IGC exhibit greater variation with differing temperature and stress state at failure than do GBC. On the basis of microcrack pattern and linear and areal densities, the IGC in dry specimens are separable into two groups (Table 2 and Figure 5). One group occurs at low peak differential stresses and is characterized by a Type-B microcrack pattern and lowest linear and areal crack densities. The second group occurs at higher peak differential stresses and is characterized by a Type-A pattern and significantly greater crack densities. Differences of microcrack statistics as a function of temperature are greater for specimens in the higher peak differential-stress group. Also with increasing temperature the average length of IGC decreases (Table 2).

The two anomalously weaker 800°C (dry) tests and the 900°C test exhibit a Type-B microcrack pattern similar to that observed with small peak differential stresses. In this case, however, GBC are significantly wider and both GBC and IGC are visually more distinct and appear to be decorated with oxides(?).

Figure 4. Maps of grain boundary cracks (GBC) in specimens E12, E21 and E15, (a-c), respectively. The bolder tracings in (b) indicate the GBC are wider and more conspicuous than at the other conditions. The least principal stress (σ_3) is parallel to the short dimension of each map. GBC development in (a) is characteristic of Pattern A and that in (b and c) is typical of Pattern B (see text). All maps are at the same scale (see beneath map c).

The influence of water on the microcrack population varies with temperature and stress state at failure. The 600°C sample exhibits an exceptionally well developed Type-A microcrack pattern, and both the LCD and the areal density are increased relative to a comparable 600°C dry sample. With temperature increase to 700° and 800°C, the microcrack pattern is characterized by a mixed A/B-pattern and a B-pattern, respectively. This Type-B pattern is essentially the same as noted for the 900 °C and two weaker 800 °C tests. Concurrent with the change of microcrack pattern with increase of temperature, there is a decrease of the LCD and areal density of IGC, as well as a decrease of the average crack length (Table 2).

4 DISCUSSION AND CONCLUSIONS

The strength of the Westerly granite deformed in extension decreases significantly with increasing temperature. The effect of temperature is equally important for tensile and compressive stress states but the relative effect increases with increasing temperature. Furthermore, at least for temperatures of 600°C and higher, the presence of water significantly enhances the effect of temperature on strength reduction. Observational studies further indicate that microfracturing is the dominant deformation mechanism.

Insight into the relative roles of the various temperature effects on the strength degradation of Westerly granite is provided by a synthesis of the data on microcrack patterns, systematic variations of microcrack statistics with increasing temperature, and axial strain

Table 2. Microfracture statistics as a function of temperature and stress state at failure.

Temp (°C)	Stresses at Failure				IGC			GBC
	σ_1 (MPa)	σ_3 (MPa)	$(\sigma_1 - \sigma_3)_p$ $C_b/2$	mf pattern	LCD (mf/mm)	areal density (mm/mm ²)	Length* (mm)	LCD (mf/mm)
600	20	-8.0	0.43	B	0.73	1.3	.53	1.7
600	100	-2.5	1.70	A	1.80	2.0	.48	1.7
600	200	16.6	2.90	A	1.50	2.1	.67	1.9
800	20	-2.8	0.90	B	0.92	1.2	.32	1.9
800	100	16.0	3.30	A	2.20	2.9	.49	1.8
800	200	52.0	6.00	A	2.70	3.8	.40	2.1
600 ⁺	100	6.4	--	A	4.20	3.2	.51	1.7
700 ⁺	100	13.2	--	A/B	3.10	2.2	.37	2.1
800 ⁺	100	57.0	--	B	1.70	1.2	.33	1.8

*Average crack length

+Wet tests, $P_c=200$ MPa, $P_p=100$ MPa

Figure 5. Linear crack density (LCD) as a function of temperature and normalized peak differential stress. Letters A and B denote microcrack pattern exhibited by specimen. The peak differential stress is normalized with respect to half the biaxial compressive strength.

(i.e. strain parallel to minimum principal stress), which is especially sensitive to microcrack development. Our assessment is based largely on currently accepted models of rock failure at room temperature and the role of pre-existing and/or mechanically-induced precursive microcracks. Numerous prior studies reveal that the relative role of pre-existing microcracks (i.e. not generated by applied mechanical stress during the test) and mechanically-induced microcracks is a function of the imposed stress state; the value of σ_3 is especially important.

Experimental studies of uniaxial tensile strength and fracture-mechanics theory (e.g. Jaeger and Cook, 1976) reveal that tensile strength is dependent largely upon the density and length of pre-existing microcracks. The applied tensile stress generates few new microcracks. Instead, failure occurs by extension and coalescence of a few pre-existing cracks. Using this model, the reduction of the uniaxial tensile strength at 600° and 800°C is attributed to reflect primarily the significant increase in the number and length of microcracks as a consequence of thermal microfracturing. In addition to the more numerous, longer cracks (compared to rock not heated), thermal microcracking probably also reduces the fracture toughness of the polycrystalline aggregate. The dominant role of thermal microcracks versus temperature-enhanced growth of the incipient tensile macrofracture at temperature is indicated by the close coincidence of the uniaxial tensile strength measured at temperature with the room-temperature, tensile strength of samples thermally cycled to equivalent temperatures (Bauer and Johnson, 1979). It should be noted, however, that thermal cracking reflects the combined effects of thermal stresses and temperature enhancement of microcrack growth. Furthermore, both the microcrack pattern (Type-B) and the densities of GBC and IGC of the 600°C and 800°C specimens ($\sigma_1 < C_b/2$) are essentially identical to that observed in specimens of Westerly granite thermally-cycled, unconfined, to equivalent temperatures (Bauer and Johnson, 1979).

As indicated by the room-temperature data (Figure 1), tensile strength is nearly independent of σ_1 , provided σ_1 is less than about half the biaxial compressive strength. This interpretation is consistent with similar data of others (e.g. Schock and Louis, 1982; Amadei and Robison, 1986) and also appears to be valid at elevated temperature, but the strength data provide less constraint, especially the 800°C data. This interpretation also is consistent with the change of the microcrack pattern from Type-B to Type-A when $\sigma_1 > C_b/2$. If $\sigma_1 < C_b/2$, tensile failure is interpreted to occur by extension of only a few existing microcracks and significant load-induced precursive microcracking is minimal. The change to a Type-A microcrack pattern for $\sigma_1 > C_b/2$ is interpreted to reflect the importance of load-induced precursive microcracking on the failure process.

In contrast to the tensile strength, the manner in which temperature affects compressive strength ($\sigma_3 > 0$) is less clear. The marked similarity of the biaxial compressive strength at temperature with the room-temperature, uniaxial compressive strength of a specimen previously thermally cycled to an equivalent temperature suggests that thermal cracking may also be the dominant control on the decrease of compressive strength if σ_3 is small. At these stress conditions, other effects of temperature are of second-order (at least for the short durations typical of these experiments). An inference of this interpretation is that thermal microcracks cause load-induced precursive microcracks to form at smaller differential stress levels compared to rock containing no thermal cracks. The formation of precursive microcracks at a smaller differential stress level may be, in part, a consequence of modification of the grain-to-grain contact conditions, which modifies the intragranular stress field induced by the mechanical stresses. In contrast with unconfined compressive strength, the room-temperature, triaxial compressive strength of thermally-cycled Westerly granite is essentially unaffected by the presence of thermal cracks, provided $\sigma_3 > 10\text{--}20$ MPa (Bauer and Johnson, 1979). A similar effect also has been noted for other thermally-

cycled crystalline rocks (e.g. Rosengren and Jaeger, 1968; Friedman et al, 1979). This observation implies that only a relatively small confining pressure is required to significantly close the thermal cracks at room temperature and re-establish intimate grain-to-grain contacts. To assume, however, that a small confining pressure has the same effect when the rock is at temperature seems inappropriate, because the differential thermal expansions responsible for microcracking and modification of grain contacts persist until the temperature decreases. This view is supported by the pronounced effect of thermal cracks upon the p- and s-wave velocities of granitic rock at high temperature and pressure (e.g. Kern, 1978). Consequently, it is anticipated that thermal cracks also play a significant role in reducing the triaxial compressive strength when the rock is at temperature.

Microcrack statistics provide insight into the relationship between the microcrack population at failure and strength as a function of temperature. The microcrack populations resulting primarily from thermal stresses (i.e. Type-B patterns) do not differ greatly between 600°C and 800°C (Table 2 and Figure 5). At 800°C, there are slightly more GBC, and IGC are slightly more numerous but shorter on the average. When failure is influenced strongly by load-induced microcracking (Type-A microcrack pattern and $(\sigma_1 - \sigma_3)_p / (C_b/2) > 1$), both the linear and areal crack densities are greater at 800°C versus 600°C (Table 2 and Figure 5). The increase in precursive microcracking also correlates with the increase of axial strain at failure. Furthermore, for a given temperature, the characteristic precursive microcrack population and the axial strain at failure appear to be independent of the differential stress at failure. This observation is consistent with that of Wawersik and Brace (1971) and Tapponnier and Brace (1976) for Westerly granite deformed in triaxial compression at room temperature. Thus, the reduction of compressive strength with increasing temperature appears to correlate with an enhanced development of precursive intragranular microcracks. This inference is contrary to that of Friedman et al (1979) for the Charcoal granodiorite deformed at temperature in triaxial compression. They infer that the number of load-induced intragranular microcracks remains constant with increasing temperature at a constant confining pressure (50 MPa), even though the compressive strength decreases with increasing temperature. The difference in the load-induced precursive microcrack population noted in these two studies may be related to the difference of compressive strength in triaxial extension versus triaxial compression, which is discussed subsequently.

The cause of the enhanced precursive intragranular microcracking remains unclear, because the data do not allow isolation of the effect of thermal stress and thermal cracking from the effect of temperature on fracture initiation and crack growth via a reduction of fracture energy and the thermal activation of the fracture process. Comparable experiments with a crystalline rock that exhibits little, if any, thermal cracking should help differentiate between the different temperature effects.

A change of failure mechanism appears to occur in dry and wet specimens at about 800°C and 700°C, respectively. This change is suggested by the anomalously low strengths of two of the 800°C specimens and the associated occurrence of a Type-B microcrack pattern in these specimens as well as that at 900°C. The 700°C and 800°C wet tests also exhibit a Type-B pattern with exceptionally wide GBC. The difference in behavior amongst the 800°C dry tests is thought to reflect possible temperature differences between the tests (there is about a 30°C temperature uncertainty at 800°C) together with a strong

temperature dependence of the higher-temperature failure mechanism. Bauer (1983) also observed a significant change of deformation characteristics of Westerly granite at about 700°C and 800°C in wet and dry triaxial-compression tests, respectively. Furthermore, he noted that for the small temperature change from 800°C to 850°C (dry) a dramatic weakening occurred.

The presence of water, at least at temperatures of 600°C and greater, significantly enhances strength degradation and, as noted above, lowers the change of failure mechanism to about 700°C. In the 600°C wet specimen, water significantly enhanced load-induced precursive microcracking as indicated by the exceptionally well developed Type-A microcrack pattern. In addition, linear and areal crack densities and axial strain at peak stress are greater than for a comparable 600°C dry specimen. At higher temperatures, water appears to enhance the role of grain boundary cracks in the deformation, as suggested by the exceptionally wide GBC and a reduction of IGC.

For compressive σ_3 and a temperature of at least 800°C (possibly 600°C), Westerly granite deformed in extension is significantly weaker compared to triaxial compression for equal σ_3 (Figure 1). Interestingly, the single extension test of Griggs et al (1960) for Westerly granite at 800°C also was significantly weaker ($\sigma_1=500$ MPa, $\sigma_3=115$ MPa) than a comparable triaxial compression test. The reason for the strength difference is uncertain. The difference, however, does not appear to reflect the effect of the intermediate principal stress, σ_2 . Other studies (e.g. Mogi, 1967, Handin et al, 1967) indicate that the effect of an increase of σ_2 , if any, is to strengthen the rock. It is conjectured that the observed strength difference may reflect the stress-path difference of the two tests and differing contributions of thermal stresses on load-induced precursive microcracking. The difference in the compressive strength as a function of temperature for these two types of tests also may reflect the different temperature-dependence of the precursive microcrack population noted above.

5. ACKNOWLEDGMENTS

The writers wish to thank Professor John Handin for his many helpful suggestions during discussions of this work, and Mr. Jack Maquirk for his help and guidance with the experimental rock deformation. This work was supported by DOE/OBES Grant No: DE-FG05-86ER13228.

REFERENCES

- Amadei, B. & M.J. Robison. 1986. Strength of rock in multiaxial loading conditions. Proc. 27th U.S. Symp. Rock Mechanics:155-164.
- Bauer, S.J. 1983. Semi-brittle deformation of granite at upper crustal conditions. Ph.D. dissertation, Texas A&M University, College Station, TX.
- Bauer, S.J. and B. Johnson. 1979. Effects of slow uniform heating on the physical properties of the Westerly and Charcoal granites. Proc. 20th U.S. Symp. Rock Mech.:7-18.
- Bauer, S.J., M. Friedman and J. Handin. 1981. Effects of

- water-saturation on strength and ductility of three igneous rocks at effective pressures to 50 MPa and temperatures to partial melting. *Proc. 22nd U.S. Rock Mech.*:79-84.
- Brace, W.F. 1964. Brittle fracture of rocks. In *Proc. Int. Conf. of State of Stress in Earth's Crust*, pp. 111-174. New York: Elsevier.
- Brace, W.F. 1971. Recent laboratory studies of stick-slip and their application to earthquakes. *Phys. Earth*. 25:185-202.
- Friedman, M. 1975. Fracture in rock. *Rev. Geophys.* 13:352-358.
- Friedman, M., J. Handin, N.G. Higgs, and J.R. Lantz. 1979. Strength and ductility of four dry igneous rocks at low pressures and temperatures to partial melting. *Proc. 20th U.S. Symp. Rock Mech.*:35-43.
- Friedman, M., J. Handin and S.J. Bauer. 1982. Deformation mechanisms in granodiorite at effective pressures to 100 MPa and temperatures to partial melting. *Proc. 23rd U.S. Symp. Rock Mech.*:279-290.
- Friedman, M. and B. Johnson. 1978. Thermal cracks in unconfined Sioux quartzite. *Proc. 19th U.S. Symp. Rock Mech.*:423-430.
- Griggs, D., F.J. Turner and H.C. Heard. 1960. Deformation of rocks at 500°-800°C. In *Rock Deformation*, *Geol. Soc. Am. Mem.* 79:39-104.
- Handin, J., H.C. Heard and J.N. Magouirk. 1967. Effects of the intermediate principal stress on the failure of limestone, dolomite, and glass at different temperatures and strain rates. *J. Geop. Res.* 72:611-640.
- Higgs, N.G. 1981. Mechanical properties of ultrafine quartz, chlorite and bentonite in environments appropriate to upper-crustal earthquakes. Unpublished Ph.D. dissertation, Texas A&M University, 267 p.
- Hopkins, T.W. 1986. Microfracturing in Westerly granite experimentally extended wet and dry at temperatures to 800°C and pressures to 200 MPa. Unpublished M.S. thesis, Texas A&M University, 71p.
- Jaeger, J.C. and N.G.W. Cook. 1976. *Fundamentals of Rock Mechanics*. 2nd Ed. Chapman and Hall, London.
- Johnson, B., A.F. Gangi and J. Handin. 1978. Thermal cracking of rocks subjected to slow, uniform temperature changes. *Proc. 19th U.S. Symp. Rock Mech.*, Reno, Nev.:259-267.
- Kern, H. 1978. The effect of high temperature and confining pressure on compressional wave velocities in quartz-bearing and quartz free igneous and metamorphic rocks. *Tectonophysics* 44:185-203.
- Kranz, R.L. 1983. Microcracks in rocks: a review. *Tectonophysics* 100:449-480.
- Krech, W.W., F.A. Henderson and K. E. Hjelmstad. 1974. A standard rock suite for rapid excavation research. U.S. Bureau of Mines, RI 7865:29.
- Mogi, K. 1967. Effect of the intermediate principal stress on rock failure. *J. Geop. Res.* 72:5117-5131.
- Rosengren, K.J. and J.C. Jaeger. 1968. The mechanical properties of an interlocked low-porosity aggregate. *Geotechnique* 18:317-326.
- Schock, R.N. and H. Louis. 1982. Strain behavior of a granite and a graywacke sandstone in tension. *J. Geop. Res.* 87:7817-7823.
- Simmons, G. and D. Richter. 1976. Microcracks in rocks. In R.G.J. Strens, editor, *The Physics and Chemistry of Minerals and Rocks*. New York: Wiley.
- Tapponnier, P. and W.F. Brace. 1976. Development of stress-induced microcracks in Westerly granite. *Int. J. Rock Mech. Min. Sci.* 13:103-112.
- Wawersik, W.R. and W.F. Brace. 1971. Post-failure behavior of a granite and diabase. *Rock Mech.* 3:61-85.

3. EXTENSION EXPERIMENTS ON SIOUX QUARTZITE AND WESTERLY GRANITE

Extension experiments have been performed on notched samples of Sioux quartzite over the temperature T range 25°-500°C at a confining pressure P_c of 100 MPa and strain rates $\dot{\epsilon}$ from 5.2×10^{-6} to $6.7 \times 10^{-4} \text{ s}^{-1}$ (Table 3.1). Over this range of conditions, the maximum differential loads supported by samples prior to failure are insensitive to temperature and strain rate, and tensile fracture strengths (of ~ 11 MPa at $T = 25^\circ\text{C}$ and ~ 4.5 MPa at $T = 500^\circ\text{C}$) are comparable to unconfined tensile strengths (of 10-11 MPa at $T = 25^\circ\text{C}$) reported by Krech et al. (1974). While fracture strengths are only weakly dependent upon temperature, significant deviations from linear elasticity prior to tensile failure are observed for samples deformed at $T = 500^\circ\text{C}$ and calculated values of inelastic work (Table 3.2) done prior to failure show a sharp increase from $\sim 0.22\text{J}$ at $T = 25^\circ\text{C}$ to $\sim 1.15\text{J}$ at $T = 500^\circ\text{C}$.

Room temperature extension experiments have been repeated on notched samples of Westerly granite (Table 3.1) using methods identical to those employed for the quartzite extension tests, for the purpose of comparison with previous investigations of Westerly granite (Bauer, 1983; Hopkins, 1986; Johnson et al., 1987). Tensile fracture strengths of Westerly granite measured at $P_c = 100$ MPa and $\dot{\epsilon} = (4.6 \text{ to } 5.7) \times 10^{-5} \text{ s}^{-1}$ are of the order of 5 MPa, as compared with those reported by Bauer (1983, $\sigma_{\text{max}} = 11$ MPa at $P_c = 100$ MPa, $\dot{\epsilon} = 3 \times 10^{-4} \text{ s}^{-1}$) and Krech et al. (1974, $\sigma_{\text{max}} = 8.6$ to 10.5 MPa for unconfined specimens). Stress-strain records for granite samples extended at room temperature exhibit some nonlinearity prior to failure and may be contrasted with the nearly perfect linear elastic response of Sioux quartzite at room temperature.

In the following, we report on the results of experiments performed on Sioux quartzite and Westerly granite at temperatures of up to 500°C. Recognizing the need for a few more experiments, decisive to a full description and understanding of their tensile fracture, we still plan to run extension tests (using other sources of funding and time) on Westerly granite (to $T = 500^{\circ}\text{C}$) and Sioux quartzite (to $T = 700^{\circ}\text{C}$) and analyze their microcrack morphologies and distributions.

Starting Materials

Sioux quartzite and Westerly granite were chosen as our starting materials and are part of the standard rock suite for rapid excavation research assembled by the U.S. Bureau of Mines (Krech et al., 1974).

Sioux quartzite is a well-indurated sedimentary orthoquartzite made up of 99% quartz grains with minor phases consisting of hematite, magnetite, zircon and rutile. Its mean grain size, determined from the linear intercept method, is 160 μm , and it is remarkably free of initial microcracks with a density of 2.64 g/cm^3 (as compared with 2.651 g/cm^3 for single crystalline α -quartz, Sosman, 1965).

Westerly granite, with 43% plagioclase, 22% microcline, 25% quartz and lesser modal fractions of biotite, muscovite, zircon, and magnetite, is technically a granodiorite and has a mean grain size of 750 μm . While the feldspars may contain some microcracks, it is a relatively dense rock with a specific gravity of 2.64.

Many other studies have been done on these starting materials and a wide variety of physical properties measurements can be found in Krech et al. (1974).

Experimental Techniques

Improvements have been made in sample preparation, apparatus design and experimental methods in order to measure accurately tensile stresses that are small relative to total differential stresses. Sample preparation has been improved substantially by turning the diameters and notched sections of samples on a lathe with a cylindrical diamond grinding tool that is held in a controlled feed and turns on a perpendicular axis to ensure reproducible notch geometries to within 0.002 inches (0.05 mm). In addition, several modifications to the load column have been made (Figure 3.1) so that corrections can be made for frictional contributions to the load and for the strengths of the metal jackets. The modified load column allows for free travel of the lower load piston, making use of a gun-lock fitting between the lower piston and the lower pressure vessel closure. Friction measurements may thus be made prior to loading the samples by measuring the load while advancing and retracting the load piston. Jacket strength measurements have been made using dummy samples precut at the notch, and measuring the load as a function of piston displacement until jacket failure. The load cell has been recalibrated by a 30,000 lb. Tinius Olsen proving ring.

Notched quartzite and granite samples have been tested with both copper and lead jackets. Raw Load records (Figure 3.2) are characterized by an initial rise as the confining pressure is increased, followed by measurements of packing friction, and a drop in load corresponding to the deviatoric stress applied during the extension test. Before contact is made with the sample, load measurements reflect the confining pressure P_c over the cross-sectional area A of the piston and packing friction ($\pm f$). Once contact is made with the sample, load measurements also include the

differential stress applied to the sample, and nominal tensile stresses are achieved at the notch when the load equals $P_c \times (A-a)$ where a is the cross-sectional area of the notched portion of the sample. Extension tests on precut samples, performed at identical conditions to those of the notched specimens, provided load displacement curves for both Cu and Pb jackets (Figure 3.3) and have been subtracted to determine the tensile loads applied to the sample (Figure 3.4). While the jacket strengths and strains to jacket failure for Cu and Pb differ, the residual load measurements, which reflect the tensile stresses supported by the samples (Figure 3.5), are in agreement, independent of the jackets used.

Data Analysis

The results shown on Figures 3.2 through 3.5 have not been corrected for the stress concentration due to the notch. In order to estimate the stress concentration factor, we need first to define the geometry. The specimens used have typical diameters of $D_s = 0.720 (+0.000, -0.002)$ inches. Notch diameters, d , are typically $0.608 (\pm 0.002)$ inches. The circular notches have a radius, $r = 0.436$ inches leading to a ratio of $r/d = 0.716$. The piston has a diameter of 1 inch and an area, $A = 0.7854$ square inches, while the notch area, $a = 0.2900$ square inches.

If we define compression as positive and the force measured by the force transducer as F_p for the Sioux quartzite specimens, F_p' for the precut specimens, and let the force acting at the root of the notch be F_n , we have

$$F_n = F_p - P_c \times (A-a) + \text{Corr} \quad (3-1)$$

and, because the precut specimen can transmit no force across the notch,

$$F_n' = 0 = F_p' - P_c \times (A-a) + \text{Corr} \quad (3-2)$$

where Corr represents the corrections for packing friction and jacket strength. If we now subtract (2) from (1), we have

$$F_n = F_p - F_p'; \quad F_p' > F_p \quad (3-3)$$

and the nominal axial tensile stress at the notch is

$$\sigma_n = F_n/a < 0 \quad (3-4)$$

Note that by subtracting the force from the precut specimen from that for the quartzite specimens, we effectively remove the corrections for packing friction and jacket deformation, and calculate directly the nominal tensile stress at the notch.

It is important to note that

(1) as soon as contact is made at the end of the free-travel displacement and F_p begins to drop from its hydrostatic value, tensile stress develops at the periphery of the notch but the net axial force at the notch remains compressive until,

(2) F_p drops below the value of F_p' on the raw force record when the force at the notch becomes tensile, and

(3) as the force progresses from state (1) to state (2), as described above, the tensile zone grows, and if a crack develops, it may grow in a stable manner because the core of the specimen remains in compression.

A stress concentration factor must be applied to the nominal stress at the notch to obtain the maximum stress at the periphery of the notch, $\sigma_{\max}$.

$$\sigma_{\max} = K_t \times \sigma_n \quad (3-5)$$

where $K_t \sim 1.27$ is the stress concentration factor for our geometry as found by extrapolation from Peterson (1974) (Fig. 32, p. 51) for $D_s = 1.184$ and $r/d = 0.716$. Average and maximum tensile stresses at failure are shown on Table 3.1.

An elementary error analysis of the measurement system indicates that the maximum stresses given in Table 3.1 should be within about 3%. Thus we attribute the variation in the results to sample-to-sample variation and not to measurement error. The average value of $\sigma_{\max}$ at room temperature ($\bar{\sigma}_n = -10.9$ MPa) agrees well with those (10-11) reported by Krech et al. (1974).

Results

Using the same methods of experimentation and analysis, we have investigated the effects of temperature and strain rate (Table 3.1) on the tensile fracture of Sioux quartzite over the range $25^\circ \leq T \leq 500^\circ\text{C}$ and $5.2 \times 10^{-6}/\text{s} \leq \dot{\epsilon} \leq 6.7 \times 10^{-4}\text{s}^{-1}$ at $P_c = 100$ MPa, and compare these results with those for Westerly granite (Bauer, 1983). Representative records of extension tests performed on Sioux quartzite and Westerly granite are shown on Figures 3.6 and 3.7, respectively. While peak values of differential stress for the quartzite and granite at $P_c = 100$ MPa are similar in magnitude and show only weak dependencies upon temperature and strain rate,

the stress-strain behavior prior to failure differ, and appear to depend upon temperature.

Stress-strain records for Sioux quartzite specimens extended at temperatures from 25° to 412°C (Figure 3.6 A, B, and C) exhibit nearly perfect linear elastic slopes to a peak value somewhat larger than the magnitude of the confining pressure, followed by a small stress drop, at which time the sample fails, and then extends to failure of the metal jacket. Experiments performed at $T = 412^\circ$ and 500°C , however, capture a significant change in mechanical response (Figure 3.6 C, D, E) with the onset of nonlinear, inelastic deformation prior to failure. While further microstructural studies must yet be done, we suspect that this inelastic deformation prior to failure is related to opening along a much larger population of microcracks. These microcracks may be thermally-induced, as those which become significant at temperatures greater than 600°C for granite (Johnson et al., 1987), or they may be associated with the larger populations of microcracks necessary to propagate a macroscopic through-going fracture under conditions where crack blunting occurs. It should be emphasized, however, that this change in mechanical response occurs fully within the α -quartz field, unlike the reduction in strength of granite (Bauer, 1983) which occurred just above the α - β quartz transition at all confining pressures tested.

Stress-strain records for Westerly granite samples extended at $T = 25^\circ\text{C}$ and $P_c = 100\text{ MPa}$ (Figure 3.7) exhibit some nonlinear yielding prior to failure, unlike that observed for Sioux quartzite under identical conditions. We attribute this nonlinear response to a larger population of microcracks initially present within the granite starting material than in the quartzite, and to the differing behaviors of feldspar and quartz. With

the greater degree of nonlinear behavior of Westerly granite at room temperature, the transition in response observed for Sioux quartzite cannot be observed, and any changes in mechanical response due to thermal cracking and crack blunting at $T = 500^\circ\text{C}$ may be masked.

While changes in mechanical response with increasing temperature prior to tensile fracture occur, peak differential stresses supported by Sioux quartzite and Westerly granite are insensitive to temperature and strain rate over the range of conditions tested. The collected results of fracture strengths as a function of strain rate are shown in Figure 3.8 for Sioux quartzite and Westerly granite in terms of $\ln(\sigma_1 - \sigma_3)$ vs. $\ln \dot{\epsilon}$. The dependencies of peak differential stress upon temperature for Westerly granite and Sioux quartzite are shown in Figures 3.9 and 3.10, respectively, in terms of $\ln(\sigma_1 - \sigma_3)$ vs. $1/T$. The results of extension experiments performed at rates of $\dot{\epsilon} \geq 6 \times 10^{-4} \text{ s}^{-1}$ have been deleted from analysis since these data appear to have been affected by experimental problems in maintaining smooth and coaxial unloading (see Table 3.1, comments for experiment Sq-12).

With the exception of experiment Sq=22, performed on Sioux quartzite at $T = 500^\circ\text{C}$, $P_c = 100 \text{ MPa}$, and $\dot{\epsilon} = 5.2 \times 10^{-6} \text{ s}^{-1}$, differential stresses ($\sigma_1 - \sigma_3$) measured at failure exceeded the confining pressure P_c ($= \sigma_1$) over an order of magnitude in strain rate for both starting materials at room temperature and at temperatures up to 600°C for Westerly granite (Bauer, 1983) and 500°C for Sioux quartzite. Thus, as illustrated in Figure 3.5, values of σ_3 are evidently tensile and have been equated with nominal values of the tensile fracture strength σ_n . While further work is needed to ascertain the drop in strength at higher temperatures ($T \geq 500^\circ\text{C}$) and lower strain rates ($\dot{\epsilon} \leq 5 \times 10^{-6} \text{ s}^{-1}$) suggested by experiment Sq-22, the

reduction in strength under these conditions may be analogous to the dramatic drop in strength observed in previous extension experiments on Westerly granite (Bauer, 1983) at temperatures of $T > 600^{\circ}\text{C}$. As previously stated, we plan to run a few more decisive experiments to resolve this question.

Inelastic Work Prior to Failure of Sioux Quartzite

Associated with the transition from nearly perfect linear elastic behavior of Sioux quartzite at $T = 25^{\circ}\text{--}412^{\circ}\text{C}$ to its nonlinear response prior to failure at $T = 500^{\circ}\text{C}$ is a significant increase in inelastic mechanical work done in extension. Assuming that the elastic deformation of samples prior to fracture is continuous, homogeneous, and linear, we can express the stored elastic energy, pertaining to those portions of the sample devoid of microcracks or flaws, by

$$U = \frac{1}{2} \int_V \sigma_{ij} \epsilon_{ij} dV \quad (3-6)$$

where V is the total volume of the specimen. U cannot be calculated directly from this general relation since the full linear elastic stress and strain tensors σ_{ij} and ϵ_{ij} are not known. However, U can be expressed as just the work done on the sample in unloading it to failure by

$$-U = W = \frac{1}{2} F_n \Delta L \quad (3-7)$$

where F_n is defined as the force drop at failure ($F_n = \Delta F_p - \Delta F_p'$ and ΔL is the change in axial length (Figure 3.11) and the second equality in (3-7) follows from equilibrium of forces.

We have used this analysis, in combination with detailed microstructural observations, to determine the fracture energy G of Sioux quartzite at room temperature (see Section 4) and to contrast its behavior at temperatures of up to 412°C with that prior to failure at elevated temperatures (Figure 3-12). Calculated values of inelastic work done on notched samples of Sioux quartzite, as well as Westerly granite, are listed in Table 3.2. Taking mean values of inelastic work determinations, a significant increase in work done to fracture is observed from 0.22 to 1.15J; an increase which apparently occurs between $T = 412^{\circ}$ and $T = 500^{\circ}\text{C}$ (Figure 3.12). While further microstructural observations are required to understand this transition, we tentatively propose that the nonlinearity prior to macroscopic fracture is due to opening a larger population of microcracks at the higher temperatures, and giving rise thereby to larger values of inelastic work done on the specimen. Whether the larger microcrack population is due to differential stresses associated with the mismatch in thermal expansion anisotropies between neighboring quartz grains (Johnson et al., 1987) or to crack blunting at elevated temperatures so that larger crack populations are required for the development of a macroscopic, throughgoing fracture must await the microstructural observations we plan to complete.

Table 3.1. Results of Extension Tests on Sioux Quartzite and Westerly Granite

Experiment Number	Sample	Jacket Material	T (°C)	P _c [*] (MPa)	$\dot{\epsilon}$ (s ⁻¹)	$\sigma_1 - \sigma_3$ (MPa) at failure ^{**}	Average Tensile Stress σ_n (MPa) at failure [†]	Maximum Tensile Stress σ_{max} (MPa) at failure ^{††}	Strain at Failure (%)	Comments
Sq-1	Sioux quartzite	Cu	25	100	5.5×10^{-5}	124.5	- 8.6	-10.9	0.85	
Sq-2	Sioux quartzite	Pb	25	100	5.6×10^{-5}	112.8	-12.2	-15.5	0.84	
Cj-1	Precut sample	Cu	25	100	5.9×10^{-5}	115.9	---	---	1.24	
Pj-1	Precut sample	Pb	25	100	5.9×10^{-5}	100.6	---	---	0.86	
Sq-3	Sioux quartzite	Pb	25	100	5.7×10^{-5}	105.6	- 5.0	- 6.4	0.76	
Sq-4	Sioux quartzite	Cu	25	100	5.4×10^{-6}	116.2	- 0.3	- 0.4	0.92	
Sq-5	Sioux quartzite	Cu	25	100	5.4×10^{-6}	119.1	- 3.2	- 4.1	0.85	
Sq-6	Sioux quartzite	Pb	25	100	6.0×10^{-4}	96.4	---	---	0.93	Apparently did not enter tensile field
Sq-7	Sioux quartzite	Pb	25	100	6.3×10^{-4}	58.4	---	---	1.26	Apparently did not enter tensile field
Sq-8	Sioux quartzite	Pb	25	100	6.7×10^{-4}	79.2	---	---	0.73	Apparently did not enter tensile field

Table 3.1. Results of Extension Tests on Sioux Quartzite and Westerly Granite (continued)

Experiment Number	Sample	Jacket Material	T (°C)	P _c [*] (MPa)	$\dot{\epsilon}$ (s ⁻¹)	$\sigma_1 - \sigma_3$ (MPa) at failure ^{**}	Average Tensile Stress σ_n (MPa) at failure [†]	Maximum Tensile Stress σ_{max} (MPa) at failure ^{††}	Strain at Failure (%)	Comments
Sq-9	Westerly granite	Pb	25	100	4.6x10 ⁻⁵	109.2	- 8.6	-10.9	1.05	
Sq-10	Precut sample	Cu	25	100	1.0x10 ⁻³	64.8	---	---	1.22	Apparently did not enter tensile field
Sq-11	Precut sample	Cu	25	100	1.2x10 ⁻³	46.5	---	---	1.76	Apparently did not enter tensile field
Sq-12	Westerly granite	Pb	25	100	7.8x10 ⁻⁴	87.3	---	---	0.86	Apparently did not enter tensile field, loading platen observed to stick-slip at this rapid displacement rate ^{**}
Sq-13	Sioux quartzite	Pb	25	100	1.2x10 ⁻⁵	93.1	---	---	0.81	Shallow sawcut at at notch used to initiate fracture
Sq-14	Westerly granite	Pb	25	100	5.7x10 ⁻⁵	102.8	- 2.2	- 2.8	0.71	
Sq-15	Sioux quartzite	Cu	412	100	1.7x10 ⁻⁵	108.0	- 6.1	- 7.7	1.01	
Sq-16	Sioux quartzite	Cu	500	100	1.7x10 ⁻⁵	107.5	- 3.4	- 4.3	1.37	

Table 3.1. Results of Extension Tests on Sioux Quartzite and Westerly Granite (continued)

Experiment Number	Sample	Jacket Material	T (°C)	P _C [*] (MPa)	$\dot{\epsilon}$ (s ⁻¹)	$\sigma_1 - \sigma_3$ (MPa) at failure ^{**}	Average Tensile Stress σ_n (MPa) at failure [†]	Maximum Tensile Stress σ_{max} (MPa) at failure ^{††}	Strain at Failure (%)	Comments
Sq-17	Sioux quartzite	Cu	500	100	1.7x10 ⁻⁵	107.9	- 3.8	- 4.8	1.23	
Sq-18	Sioux quartzite	Cu	500	100	1.7x10 ⁻⁵	101.9	---	---	1.03	
Sq-19	Sioux quartzite	Cu	500	100	5.2x10 ⁻⁶	---	---	---	---	Furnace failure
Sq-20	Sioux quartzite	Cu	500	100	5.2x10 ⁻⁶	---	---	---	---	Jacket leak
Sq-21	Sioux quartzite	Cu	500	100	5.2x10 ⁻⁶	---	---	---	---	Jacket failure
Sq-22	Sioux quartzite	Cu	500	100	5.2x10 ⁻⁶	77.9	---	---	0.82	Apparently did not enter tensile field

* P_C = σ_1

** Nominal value at notch (uncorrected for jacket strength).

† Nominal value at notch corrected for friction and jacket strength.

†† Corrected for stress concentration at notch, friction and jacket strength.

*** Parallelism of loading platen apparently difficult to maintain at the rapid displacement rate of this test and may be a source of error in previous high strain rate experiments which did not enter tensile field.

Table 3.2. Inelastic Work Done in Extension Tests

Sioux Quartzite

 $T = 25^{\circ}\text{C}$ $P^* = 100 \text{ MPa}$ $\dot{\epsilon}^c = (5.5 \text{ to } 5.7) \times 10^{-5} \text{ s}^{-1}$

Experiment Number	$W^{**} \text{ (J)}$
Sq-1	0.21
Sq-2	0.34
Sq-3	0.12

Average Value = 0.22 J

Westerly Granite

 $T = 25^{\circ}\text{C}$ $P^* = 100 \text{ MPa}$ $\dot{\epsilon}^c = (4.6 \text{ to } 5.7) \times 10^{-5} \text{ s}^{-1}$

Experiment Number	$W^{**} \text{ (J)}$
Sq-9	1.22
Sq-14	0.31

Average Value = 0.77 J

Sioux Quartzite

 $T = 25^{\circ}\text{C}$ $P^* = 100 \text{ MPa}$ $\dot{\epsilon}^c = (5.4 \pm 0.1) \times 10^{-6} \text{ s}^{-1}$

Experiment Number	$W^{**} \text{ (J)}$
-------------------	----------------------

Sq-5

0.08

Table 3.2. Inelastic Work Done in Extension Tests (continued)

Sioux Quartzite

 $T = 412^{\circ}\text{C}$ $P^* = 100 \text{ MPa}$ $\dot{\epsilon}^C = (1.7 \pm 0.1) \times 10^{-5} \text{ s}^{-1}$

Experiment Number

 $W^{**} \text{ (J)}$

Sq-15

0.10

Sioux Quartzite

 $T = 500^{\circ}\text{C}$ $P^* = 100 \text{ MPa}$ $\dot{\epsilon}^C = (1.7 \pm 0.1) \times 10^{-5} \text{ s}^{-1}$

Experiment Number

 $W^{**} \text{ (J)}$

Sq-16

1.36

Sq-17

0.94

Average Value = 1.15

* $P_C = \sigma_1$

** Calculated from equation 3-7.

Figure 3.1. Lower piston assembly, incorporating hollow hardened steel tube threaded to pressure cell endplug, and lower piston which engages with hollow steel tube by way of gun-lock fitting. The piston assembly allows for free travel of the piston in order to make measurements of packing friction prior to loading the sample.

Figure 3.2. Raw load-ram displacement record for experiment Sq-2. Following pressurization to $P_c = 100$ MPa, the load piston was advanced and then retracted in order to measure the packing friction (measured load difference = $2f$). The extension experiment was then performed by retracting the load piston at a constant displacement rate. The negative deflection in the force record corresponds to the lower piston engaging with the hollow steel tube. Tensile stresses at the notched portion of the sample are nominally obtained at a load of $P_c \times (A-a)$ where P_c is the confining pressure, A is the cross sectional area of the load piston and a is the cross sectional area at the notch.

Figure 3.3. Raw differential load records for notched samples of Sioux quartzite (and precut dummy samples) jacketed in (A) Cu and (B) Pb jackets. Cu jackets are significantly stronger than Pb jackets with larger strains to failure. The load differences, corresponding to the strengths of the quartzite samples, however, appear to be nearly the same.

Figure 3.4. Load differences for experiments Sq-1 and Sq-2, subtracting the load records for Cu and Pb jacket strengths, respectively, and normalized by the cross-sectional area at the notch. Nominal tensile fracture strengths, measured at failure (shown as dashed lines), are 8.6 and 12.2 MPa, respectively, before correcting for stress concentrations at the notch.

Figure 3.5. Mohr diagram of experiments Sq-1, Sq-2, and Sq-3 performed at room temperature and a constant strain rate of $1.9-2.6 \times 10^{-5}$ /s, and $P_c = 100$ MPa with uncorrected tensile strengths of 8.6, 12.2 and 5.0 MPa, respectively.

Figure 3.6. Differential stress vs. strain for Sioux quartzite specimens extended at a confining pressure $P_c (= \sigma_1)$ of 100 MPa and strain rates of $(1.7 \text{ to } 5.6) \times 10^{-5} \text{ s}^{-1}$ and temperatures of $T = 25^\circ\text{C}$ (A and B), $T = 412^\circ\text{C}$ (C), and $T = 500^\circ\text{C}$ (D and E). Nominal differential stresses have been corrected for friction but not for jacket strength nor for the stress concentration at the notch (see Table 3.1).

Figure 3.7. Differential stress vs. strain for Westerly granite samples extended at room temperature, a confining pressure $P_c (= \sigma_1)$ of 100 MPa, and strain rates of $(4.6 \text{ to } 5.7) \times 10^{-5} \text{ s}^{-1}$ (A, B). Nominal differential stresses have been corrected for friction but not for jacket strength nor for the stress concentration at the notch (see Table 3.1)

Figure 3.8. Influence of deformation rate on differential stresses measured at failure. Ln-ln plots of peak differential stress vs. strain rate for Sioux quartzite at T = 25°C (A) and T = 500°C (B), and Westerly granite at T = 25°C (C) and T = 600°C (D) using data of Bauer (1983). Fields of tension and compression refer to the sign of σ_3 where we choose the convention that (+) refers to compressive stress, and (-) refers to tensile stress (= σ_1). Differential stresses have been corrected for friction and jacket strength, but not for the stress concentration at the notch.

Figure 3.9. Effect of temperature on differential stresses measured at failure for Westerly granite (data of Bauer, 1983, and recent experiments used for comparison). $\ln(\sigma_1 - \sigma_3)$ vs $1/T$ plot shows a dramatic transition from temperature-insensitive fracture strengths at $T < 600^\circ\text{C}$, and temperature-dependent strengths which remain within the macroscopically compressive field at $T > 600^\circ\text{C}$.

Figure 3.10. Effect of temperature on differential stresses measured at failure for Sioux quartzite. $\ln(\sigma_1 - \sigma_3)$ vs. $1/T$ plot for the quartzite shows the same behavior as that for Westerly granite at low temperatures with temperature-insensitive failure strengths, and the suggestion of temperature- and strain rate-dependent failure at $T \geq 500^\circ\text{C}$ and $\dot{\epsilon} \leq 5 \times 10^{-6} \text{ s}^{-1}$. Differential stresses have been corrected for friction and jacket strength, but not for the stress concentration at the notch.

Figure 3.11. Inelastic work determinations from differential load ΔF_p versus displacement ΔL records for extension tests Sq-5 (A) and Sq-2 (B). Measurement of force drop $F_n = \Delta F_p - \Delta F_p'$ at failure illustrated for high strength test Sq-2. Work W used to calculate fracture energy $\bar{G}$ corresponds to one-half the shaded area under each curve.

Figure 3.12. Inelastic work determinations from differential load versus displacement records for extension tests Sq-15 (A) and Sq-16 (B) performed at $T = 412^{\circ}\text{C}$ and $T = 500^{\circ}\text{C}$, respectively. A significant increase in inelastic work W done prior to failure is observed, between these two test temperatures, associated with the far more nonlinear response prior to failure at $T = 500^{\circ}\text{C}$.

4. FRACTOGRAPHY AND MICROCRACK DISTRIBUTION IN EXPERIMENTALLY EXTENDED SIOUX QUARTZITE: MECHANISM AND ENERGETICS OF BRITTLE FRACTURE

Introduction

Tensile fracture propagation in extended samples of Sioux quartzite has been studied by optical and SEM examination of surfaces and traces of fractures produced in extension experiments on notched cylindrical specimens at room temperature and confining pressures of 100 MPa and 200 MPa. In one test, fracture was initiated by an induced edge flaw (Sq-13, Table 3-1). Macroscopically, the resulting fracture surface comprised smooth (mirror) and rough (hackle) zones concentrically surrounding the flaw similar to those produced in experiments with glass and fine-grained ceramics. These zones and the presence of aligned macroscopic steps indicate a consistent direction of overall fracture growth, accelerating away from the edge flaw. In contrast, the direction of crack propagation at the microscopic scale inferred from surface features on individual quartz grains showed no such overall trend. Even within the zone of stress concentration associated with the induced fracture origin, the local direction of crack propagation is controlled by the location of critical flaws, especially grain boundaries, and the orientation of cleavage planes within individual grains. The results suggest a mechanism for catastrophic fracture propagation in quartzite not unlike that for stable, i.e., quasi-static, fracturing inferred from previous field and laboratory studies of various quartz-bearing crystalline rocks, i.e., propagation effected by localized growth and coalescence of parallel microcracks within a narrow zone about the primary fracture front. An estimate for the

fracture energy $G = 30 \text{ J/m}^2$ for another test obtained from force-displacement data and detailed mapping of fractures is smaller than previously reported values for quartzite and closer to G typical of quartz single crystals, 1 J/m^2 to 10 J/m^2 . The improved correspondence between single crystal data and the polycrystal value of G obtained here is due to a more accurate determination of the actual fracture surface area.

Methodology and Purpose of Detailed Microstructural Analysis

The rich variety of features developed on the surfaces of fractures in brittle solids provides a record of the variations in stress and velocity at the tips of propagating cracks. Detailed observation and interpretation of such features is thus a potentially powerful tool for studying the fracture process, especially when combined with experimental determination of fracture mechanics parameters, such as critical stress intensity factors or fracture surface energies. Microscopic examination of fracture surfaces may be used to determine the locations and dimensions of fracture-initiating flaws, the direction and velocity of crack propagation and details of the mechanism of crack nucleation and growth. Fractography is therefore routinely employed in failure analysis of engineering materials including metals, glass, ceramics and polymers (e.g. see Strauss and Cullen, eds., 1978), and has served to guide modern theoretical work on fracturing in these materials (e.g. Broek, 1974).

Although a large literature now exists on fracturing in rocks and minerals, relatively few studies have reported detailed observations of fracture surface features. Ryan and Sammis (1978) studied the mechanics of columnar jointing of basalt driven by thermal stresses during cooling from a melt. Combining evidence from seismic surveys of crystallizing basalt

flows with an analogy to fracture surface features produced in cyclic-loading experiments with aluminum and various polymers, they interpreted the parallel, evenly spaced markings on joint surfaces as fatigue striations, each corresponding to a discrete episode of crack extension. Smooth and rough bands within each striation were associated with elastic and inelastic deformation, respectively, of the host rock at crack tips. Bahat and Engelder (1984) described plumose markings on the surfaces of cross-fold joints cutting interbedded siltstones and shales, also attributing these to episodic crack growth.

Most experimental studies of fracturing in geologic materials which report detailed observations of fracture surfaces have been conducted with quartz single crystals. This is due mainly to its importance as a rock-forming and industrial mineral and its similarity to glass in which fracturing has been studied extensively. In fact, fracture surface features produced in tensile tests on quartz are quite similar to those in glass (Norton and Atkinson, 1981). In addition, quartz has a weak cleavage on its rhombohedral planes which effects the formation of stepped fracture surfaces (Martin and Durham, 1975; Ball and Payne, 1976) whose form provides further information on the details of its fracture behavior.

The primary purpose of this study is to investigate the mechanism of tensile crack growth in a quartz polycrystal by examining the direction of crack propagation inferred from various surface features produced in extension tests on Sioux quartzite. A second goal is to derive an estimate of the fracture energy from mechanical data and mapping of fracture traces in one extension test specimen. This study was undertaken as part of a larger project designed to assess the effects of temperature, strain rate and confining pressure on the deformation of brittle rock, particularly as

it pertains to the problem of borehole stability. While the results presented here are only for room temperature tests, they demonstrate the utility of making detailed observations of fractures and provide a standard for comparison of features produced at different conditions, especially higher temperature. The analysis of fracture formation in extension tests also provides a rational basis for investigating the more difficult problem of shear deformation of brittle rock, the likely mode of borehole failure at depth (Friedman et al., 1979), since such deformation involves cracking initiated at points of tensile stress concentration.

Characteristic Features on Fracture Surfaces in Brittle Solids

Glass and Ceramics

An immense body of experimental work on tensile crack propagation in glass and ceramics (e.g. see Rice et al., 1984) has established the existence of a canonical ensemble of four textural regions which are disposed in radially concentric fashion about the fracture origin (Figure 4-1). In most cases the origin in such experiments is an artificially induced surface flaw such as a saw-cut or indentation. Immediately surrounding the origin is a very smooth, almost featureless region known as the fracture mirror, the name being derived from its high reflectivity in glass specimens. Bordering the mirror is a zone of incipient roughening, usually consisting of very fine radial ridges, referred to as mist. This in turn gives way to a zone of larger scale ridges known as hackle beyond which macroscopic crack branching is observed to occur.

A variety of factors may influence the relative intensity of development of these textural zones, including microstructure, specimen

size and variations in macroscopic stresses. For example, all four zones are often well developed and mutually distinct on fracture surfaces in glass owing to its homogeneity and isotropy. In polycrystalline ceramics, however, some zones may be indistinct or even absent, especially in coarse-grained porous materials. The effect of specimen size is also important, since the distance from the fracture-initiating flaw to the boundary between a pair of zones is supposed to be a material property, i.e., a function of the applied stress required for fracture σ_f . This empirical relation is typically expressed in terms of the following equation:

$$\sigma_f \sqrt{r_i} = A_i, \quad [4-1]$$

where A_i is a constant for a particular zone boundary at radial distance r_i from the fracture origin (e.g., Mecholsky et al., 1978). For example, A_m is the so-called mirror constant corresponding to the boundary between the mirror and mist zones at a distance r_m from the origin. (Unfortunately, it is also common usage to designate the mirror and mist zones, as identified above, as the inner and outer mirror regions, respectively, in which case A_m refers to the boundary between mist and hackle areas.) All four textural zones may be present only if the characteristic dimension of the fracture surface, e.g. specimen diameter, is larger than r_b , the distance at which the onset of macroscopic crack branching occurs.

Equation 4-1 has the same form as the criterion for opening (mode I) crack propagation from linear elastic fracture mechanics (e.g. Lawn and Wilshaw, 1975):

$$K_{IC} = \sigma_f Y \sqrt{\pi c}, \quad [4-2]$$

where c is the dimension of the initiating flaw and Y is a constant that depends on the specimen geometry and loading conditions. K_{IC} , the critical stress intensity factor, expresses the strength of the tensile stress concentration at the tip of a critically loaded crack. The obvious implication from comparing equations 4-1 and 4-2 is that the various textural boundaries correspond to contours of stress intensity associated with the initial flaw, although Rice et al. (1984) discuss some evidence to the contrary. In fact, a number of studies have reported that A_i/K_{IC} is a constant (Mecholsky et al., 1978). A typical application of fractography in failure analysis is to determine the initial flaw size from measurements of r_i using known values for σ_f , Y and K_{IC} in an expression obtained by combining equations 4-1 and 4-2.

The development of zones of increasing surface roughness in the direction of crack propagation is a dynamic effect in which breakdown of a single crack front into numerous secondary cracks occurs (e.g., Mecholsky et al., 1978). The mirror region corresponds to the propagation of a single, primary crack that accelerates away from the initiating flaw. As its velocity v_c nears the terminal velocity for crack propagation of the medium v_c^T , sufficient energy becomes available to nucleate secondary microcracks about the primary crack tip, but not for extensive propagation of these secondary cracks, resulting in the formation of mist. The velocity of Rayleigh surface waves v_R is usually taken as an upper bound for v_c^T (e.g., Field, 1971). $v_c^T \approx 0.6 v_s$, where v_s is the s-wave velocity, is typically observed in experiments (Mecholsky et al., 1978). McColm (1983) gives the theoretical estimate $v_c^T \approx \sqrt{\sigma_y/4\rho}$ where σ_y is the yield

stress and ρ is the density. In any case, the larger energy release rate associated with a longer overall crack cannot affect further acceleration of the primary crack and instead produces increasingly extensive propagation of secondary cracks leading to hackle formation and, ultimately, crack branching. Typically, the presence of many more flaws in a polycrystal, especially grain boundaries, than in glass or single crystals tends to hasten the onset of breakdown of the accelerating primary crack front. Thus, for example, $r_m/c \approx 13$ is often observed for glass whereas $r_m/c \approx 6$ is commonly found for polycrystalline ceramics (Mecholsky et al., 1978).

A number of other characteristic fracture surface features have been identified in experiments with glass and ceramics whose form may be used to infer the direction of crack propagation. Straight steps indicate crack growth parallel to the step lines since this configuration minimizes the increase in surface area, and therefore energy, associated with a change in elevation of the fracture surface. River patterns are often formed on the surfaces of individual grains in which fracturing is cleavage-controlled. This can occur when a propagating crack crosses either a tilt or twist grain boundary or sub-grain boundary in the manner illustrated in Figure 4-2. The inferred direction of crack propagation is down-stream, i.e., in the direction of convergence of cleavage steps (see Friedel, 1964 for a theoretical treatment of river pattern formation). Wallner lines are arcuate striations formed by oscillatory, out-of-plane deflection of a propagating crack tip due to its interaction with an obliquely propagating s wave (McColm, 1983). These features have a gentle, low-amplitude profile and are usually only observed within the mirror zone. The source of the interfering stress wave is a flaw cut by the propagating crack tip and each

line marks the locus of intersection between the crack tip and an s wave front. Since $v_c < v_s$ each line is concave toward the fracture origin. In addition to indicating the direction of crack propagation, measurements of these features may be used to evaluate the ratio v_c/v_s . Thus, the local crack propagation velocity can be determined if the shear wave velocity for the material is known (Shand, 1954; Field, 1970).

Quartz

Essentially all the fracture surface features described above have been produced in tensile fracture experiments with quartz single crystals, with some modifications or additional features due to its cleavage on the rhombohedron planes $r(10\bar{1}1)$ and $z(01\bar{1}1)$ (Ball and Payne, 1976; Norton and Atkinson, 1981). A mirror region is always present about the fracture-initiating flaw. However, for most crystallographic orientations, mist and hackle are replaced by cleavage steps which are concave toward the fracture origin and increase in size in the direction of crack propagation (Figure 4-3). Macroscopic crack branching occurs on r and z . Ball and Payne (1976) made a comprehensive study of tensile crack growth in quartz from an edge slit in thin, single crystal slabs of five different orientations in a variety of chemical environments and temperatures ranging from 77K to 800K. Arcuate cleavage steps were produced on fracture surfaces of all orientations except for fracture on the $(11\bar{2}0)$ plane in the $[1\bar{1}00]$ direction. This last orientation exhibited mirror and hackle zones similar to glass with well developed Wallner lines on the mirror and no crack branching. These observations and considerations of the anisotropy of Si-O bond density in quartz lead to the interpretation that the applied stress direction was inclined at too great an angle to open cleavage cracks on r and z . Fracture velocities found from measurements of Wallner lines

agreed well with those determined by direct measurement using an electrical method. Norton and Atkinson (1981) studied crack propagation from a surface flaw in synthetic and natural Brazilian quartz crystal rods in bending and found the relation [4-1] for the mirror constant A_m to hold for fracturing on the basal (0001) and m (1010) planes. Arcuate cleavage steps on basal fracture surfaces exhibited an angular, saw-tooth profile and formed intersecting sets with a characteristic hexagonal symmetry. Two intersecting sets of arcuate steps with a gentler, low-amplitude form were produced on m-fractures. In both cases the direction of crack propagation was parallel to a line bisecting the angle of step intersection. The concavity of cleavage steps toward the fracture origin, especially those on m, suggests that their formation is influenced by interaction with elastic stress waves in a manner similar to that of Wallner lines.

Results and Discussion

The fractures examined in this study were produced in room temperature extension tests on nominally dry (no pore pressure) Sioux quartzite, described in Section 3. Because it is monomineralic, equigranular, has very low porosity and is relatively coarse-grained, this material is well suited for the purposes of this study. The physical properties and mechanical behavior of Sioux quartzite are known from previous studies (e.g., Krech et al., 1974), including fracture mechanics parameters (Peck and Gordon, 1982).

Direction of Crack Propagation

Macroscopically, all the fracture surfaces examined were quite flat. Flake-like steps are present in all cases but very few of these exhibit any noticeable preferred orientation. Compared to the surrounding rock, many

steps are much lighter in color, appearing almost translucent. This is due to scattering of light by parallel fractures just below the exposed fracture surface, a geometry which is shown more clearly by mapping of fracture traces in cross-section (see "Fracture Energy").

The fracture surface of one of the samples examined (not shown) contained three large steps the surfaces of which are parallel and inclined slightly to the fracture surface as a whole. These steps appear to radiate from a point on the margin of the sample, suggesting an overall direction of fracture growth outward from this point and parallel to the step lines. Although there is no macroscopically obvious flaw at the focus of the step lines, SEM did reveal what appeared to be a large pore in this area which may have initiated fracturing. Quartz grains exposed on the fracture surface exhibit numerous cleavage steps, including river pattern, from which the local crack propagation direction was "mapped" in two profiles of contiguous SEM photos trending outward from the apparent fracture origin. (A detailed description of surface features and the mapping procedure is given below for another sample.) No overall trend was discernable, i.e., the local direction of crack propagation appeared essentially random.

The lack of a radially inward trend in the inferred direction of local crack propagation is an expected result since the stress concentration associated with the notch is quite weak (see eqn. 3-5). However, it is sufficient to control the overall location of fracturing since it acts over a broad region comparable in dimension to the radius of curvature of the notch, $r = 0.44$ inch. Microstructural flaws having large ratios of length to width, e.g., grain boundary cracks, will produce much stronger local stress concentrations, especially if these are oriented normal to the maximum tensile stress direction. Such features will therefore be the

dominant sources of crack growth and, together with cleavage directions in individual grains, will control the local crack propagation direction. Thus, a consistent trend in local crack propagation direction over a region many grain diameters in extent is not expected for small flaws or weak stress concentrations. However, one might expect that a trend in macroscopic surface features indicating the overall direction of fracture growth would be preserved at the grain scale, especially if the origin was a strong stress concentrator whose dimensions were large relative to the grain size.

To address this question, an extension test was performed using a modified notched specimen in which a shallow saw cut was made into one side of the notch mid-plane to initiate fracturing (Figures 4-4a and b). The cut was made using a circular diamond wafering blade and has a maximum depth $c \approx 0.55 \text{ mm}$ (0.022 inch). The blade has a rounded edge and thickness t between 0.017 inch and 0.018 inch. Hence, in cross-section the cut has a blunt, sub-circular tip with a radius of curvature $r \approx t/2$ roughly 50 times smaller than that of the notch and width slightly greater than the rock grain size, 0.2 mm to 0.3 mm.

Because of the non-concentric geometry, an accurate estimate of the stress concentration at the tip of the saw cut is difficult to make. However, since r is small relative to all other specimen dimensions, a rough estimate of this stress intensity may be obtained by analogy to a cylinder with an all-around groove whose depth is equal to that of the saw cut plus the notch, $c + (D - d)/2$. For this simpler case, the stress concentration factor is $k \approx 13$ (eq. 3-5, Peterson, 1974, Figure 3-1, p. 50), i.e., the stress concentration at the deepest part of the saw cut may be approximately an order of magnitude stronger than that due to the notch.

The fracture surface of this sample comprises two regions which can be distinguished on the basis of surface roughness. Surrounding the saw cut is a relatively smooth zone containing a few, mostly very small steps that is interpreted as a fracture mirror. The remainder of the fracture surface contains much larger and more numerous steps and is interpreted as hackle. The edges of the two largest steps, near the center of the fracture surface and those of many smaller steps trend nearly normal to the saw cut, i.e., are parallel to the expected direction of fracture propagation. Note that nearly all steps have the same vergence, i.e., slope from left to right as viewed in Figures 4-4a and b. Viewed along the fracture plane away from the saw cut, these steps would appear as echelon cracks rotated slightly in a clockwise direction relative to the fracture plane as a whole. Presumably, these formed as secondary cracks associated with the breakdown of the primary crack front as it accelerated away from the saw cut. Mist is not apparent. Indeed, this is expected since the quartzite is so coarse-grained (Rice et al., 1984). The absence of a zone of macroscopic crack branching may be due to insufficient specimen size, $D < r_b$.

The boundary between the mirror and hackle regions is symmetric about and concave toward the saw cut and, despite the coarse grain size of the quartzite, is quite distinct (Figure 4.4a). The distance from the edge of the saw cut to this mirror boundary varies from a minimum of about 2.5 mm along the specimen edge to a maximum of 3.2 mm as measured along a diameter bisecting the saw cut (Figure 4.4b). Using this maximum value for the mirror radius r_m yields the ratio $r_m/c \approx 6$ equal to the typical value for polycrystalline ceramics given by Mecholsky et al. (1977). With this measurement for r_m , a corresponding mirror constant A_m could be calculated

from equation 4-1 if σ_f were known. Unfortunately, σ_f is not known for this test (Table 3-1) since the force F_n at failure was immeasurably small. This, of course, is not surprising because of the large stress concentration associated with the saw cut. Although K_{IC} has been measured for Sioux quartzite (Peck and Gordon, 1982), σ_f cannot be calculated from 4-2 since an appropriate stress analysis for the present sample configuration is not available, i.e., Y is also unknown. However, it is clear that, for a given geometry and set of test conditions, σ_f and r_m will both be greater for single crystals than for polycrystals of the same composition. Thus, we are at least assured that A_m for Sioux quartzite will be less than $3.7 \text{ MPa} \cdot \text{m}^{1/2}$, the value given for natural Brazilian quartz in Norton and Atkinson (1981).

The local direction of crack propagation inferred from surface features on individual quartz grains was mapped in a traverse of contiguous SEM photographs, each taken at 100x magnification, along a chord oriented normal to the saw cut (Figure 4-5; see Figure 4-4 for traverse location). The fracture surface consists almost entirely of parallel cracks through the interiors of grains, i.e., transgranular cracks, linked by small steps at grain boundaries inclined at large angles to the fracture surface. A very small proportion of the total fracture surface comprises grain boundary cracks (stippled, Figure 4-5). At the grain-scale, there is no marked difference in surface roughness between the mirror and hackle regions: surface roughness in SEM photos from opposite ends of the traverse appears essentially the same.

Each determination of local crack propagation direction has been assigned either a high-confidence (solid arrows, Figure 4-5) or low-confidence (open arrows) rating. The rating is subjective and reflects possible ambiguity in inferring crack propagation direction at the present

scale of mapping. High-confidence ratings were reserved for those features judged as entirely unambiguous indicators of local propagation direction. Examples include distinct river pattern (Figure 4-6) and closely spaced, parallel cleavage steps which are distinctly concave toward a common focus (Figures 4-7 and 4-8). In some cases, sets of intersecting arcuate steps are observed. In one example (Figure 4-9), these steps had a gentle, low-amplitude form highly reminiscent of Wallner lines. However, interpretation of the angles of intersection of these striations according to the mechanism of Wallner line formation yields an inconsistent direction of crack propagation. Norton and Atkinson (1981) interpreted similar intersecting arcuate striations on basal fracture surfaces in their experimental samples as cleavage steps. In general, the line of intersection of a planar crack and a cleavage plane should not be curved. We agree that these curved striations are cleavage steps but suggest that their form is modified due to interaction between a propagating cleavage crack tip and obliquely propagating stress waves in a manner similar to Wallner lines. This, in turn, suggests that measurements of these features could be used to infer crack propagation velocity if an appropriate analysis of the mechanics of interaction were available. Such an analysis is suggested as the subject of future research.

Low-confidence ratings were awarded in situations where an interpretation other than the one given was possible but still less likely. For instance, in some cases parallel, arcuate cleavage steps splay along their lengths to form fine-scale, yet distinct, river pattern (Figure 4-10). As a whole, such a feature was taken to indicate crack propagation normal to the step lines, from the concave to the convex side, although propagation parallel to the steps in the downstream direction of the rivers

is a plausible interpretation. Other examples include weakly curved or irregularly spaced steps (Figure 4-11) for which propagation either parallel or normal to the steps was defensible. No determination of crack propagation direction was made in cases where two or more alternative interpretations appeared equally likely. Low-confidence ratings were also given in cases where the inferred direction of crack propagation varied markedly within small areas of single grains, even if, taken separately, each direction could be regarded as a high-confidence determination (Figures 4-12 and 4-13). In such situations, the low-confidence rating is due to finite map resolution instead of ambiguity in inferring direction of propagation from a particular surface feature. Marked variation in crack propagation direction within individual grains can be due to the presence of subgrain structure (Figure 4-12) or multiple crack sources distributed along the grain boundary (Figure 4-13).

No overall trend in local crack propagation direction is evident in Figure 4-5. Although macroscopic surface features (Figure 4-4) indicate an overall direction of fracture growth outward from the saw cut, the direction of crack propagation at the grain scale again appears essentially random. As discussed above, this result indicates that microstructural heterogeneities, especially grain boundaries and cleavage, are the dominant controls on the local direction of crack growth. A striking illustration of this is shown in Figure 4-6 in which a prominent river pattern is developed on a transgranular crack surface adjacent to a grain boundary crack. The local direction of crack propagation is unambiguous and is at least 100° oblique (clockwise) to the overall direction of fracture extension. Note that this location is only $\sim 2r$ from the edge of the saw cut (Figure 4-6a) and therefore well within its associated zone of stress

concentration. The obvious interpretation here is that opening displacements across the grain boundary crack further concentrated tensile stresses along its perimeter thereby initiating crack growth. Note that this flaw is in the plane of the fracture surface and thus has the expected critical orientation, i.e., normal to the applied tension direction.

Taken together, the observations are consistent with a mechanism for catastrophic fracture propagation involving concentrated growth and coalescence of parallel microcracks about the primary crack tip. This is essentially the same process which has been inferred for stable fracture growth from observations of both naturally (e.g., Segall and Pollard, 1983) and experimentally (Hoagland et al., 1973; Peck and Gordon, 1982) deformed rocks. The present results indicate that while a consistent direction of crack propagation can be identified at the macroscopic scale, the microcrack growth which effects the propagation has no such preferred direction, even in the mirror region. Hence, the fracture front is never microscopically sharp but instead must be somewhat diffuse, being defined by a narrow zone of localized microcracks.

Fracture Energy

In one experiment (Sq-5, Table 3-1) using a standard notched specimen (no saw cut), the loading was stopped after the peak load was achieved but before separation of the specimen across the fracture occurred. The complete fracture trace for this run could thus be observed in a single thin section (Figure 4-14a). This suggested that a reasonably accurate estimate of the total fracture surface area could be obtained from which an estimate of the fracture energy could be derived.

The fracture energy G , also known as the energy release rate, is defined as the energy irreversibly consumed per unit area a of crack extension:

$$G = \frac{d}{da}(W-U), \quad [4-3]$$

where W is the work done by applied loads, U is the total strain energy of the cracked body and inertial forces are neglected. Thus, when fracture propagation occurs in a stable manner, G can be determined as a function of crack length from an experimental force versus displacement record. To calculate G in the present situation involving catastrophic fracturing, we assume that (a) deformation of the sample prior to fracture is continuous and linear elastic, e.g. no sub-critical crack growth occurs, (b) fracturing is instantaneous relative to the rate of unloading and (c) acceleration of the system (sample + apparatus) is negligible. The small force drops and absence of audible acoustic emissions at failure suggest that (c) is a reasonable assumption. Considering the strain rates involved, (b) is clearly reasonable and implies that we may assume a fixed displacement at the point of failure, i.e., the "fixed-grips" condition:

$$G = G|_{\text{fixed grips}} = \frac{-dU}{da}. \quad [4-4]$$

Recalling the relation for U from section 3, (a) in turn implies

$$U = \frac{1}{2} \int_V \sigma_{ij} \epsilon_{ij} dV, \quad [3-6]$$

or just

$$-U = W = \frac{1}{2} F_n \Delta L. \quad [3-7]$$

Since failure is catastrophic, incremental fracture growth cannot be

observed. We therefore calculate an average fracture energy G given by

$$\bar{G} = W/\alpha \quad [4-5]$$

where α is the total fracture surface area.

α has been estimated for test specimen Sq-5 by mapping fracture traces in a thin section which contains the cylinder axis (Figure 4-14b). The fracture consists of numerous parallel cracks most of which are linked. The vast majority of these are transgranular cracks concentrated within a narrow band whose width is comparable to the rock grain diameter, 0.2 mm to 0.3 mm. All mapped cracks occur within a zone 1.5 mm wide about the notch mid-plane. These observations indicate that all mapped cracks were produced during the extension test. At least three large overlapping echelon cracks can be distinguished. These are clearly composites, formed by the coalescence of numerous finer-scale cracks, which would appear as inclined steps on the fracture surface in plan view. Note that the specimen is essentially intact, i.e., only an incipient connection exists between the two large echelon cracks near the center of the mapped area where they overlap. The greater abundance of microcracks distributed over a broader region at the right side of the map may be due to the weak stress concentration associated with the notch. However, there is no corresponding concentration of microcracks at the left-hand side. This asymmetry may be due to slight non-parallelism of the specimen ends or misalignment of the loading column resulting in small bending moments applied to the specimen during extension.

The total line length of fracture traces measured on the map was found to be 58.75 mm. The horizontal length of the map, equal to the notch

diameter d , is 15.47 mm (0.609 inch). Thus, there is an average $\lambda \approx 58.75/15.47 \approx 3.8$ mm of mapped crack length per mm of horizontal distance on the map. Assuming λ is the same for any diameter, i.e., homogeneous fracturing, the total fracture surface area is

$$\alpha = \lambda^2 a \approx 2.7 \times 10^{-3} \text{ m}^2, \quad [4-6]$$

where a is the nominal cross-sectional area at the notch.

W has been calculated according to equations 3-7 and 4-5 from the differential load $\Delta F_p = F_p - F_p^0$ versus displacement ΔL records for this test (Sq-5, Figure 3-11a) and a second test (Sq-2, Figure 3-11b). These two tests represent the upper and lower extremes for strength in tension found for standard notched specimens of Sioux quartzite at room temperature. The measurements of ΔF_p and ΔL are quite accurate, probably to within at least 3%. However, errors in determining F_n , equal to the force drop shown as a small step at the peak of each record, are substantially larger. For test Sq-5 the step height on the strip chart record is comparable to the pen width so that F_n , and thus the corresponding value of $W = 0.08$ J, is probably accurate only to within a factor of 2 (Figure 3-11a). The value obtained for test Sq-2, $W = 0.34$ J, is probably accurate to within at least 10% because of the larger force drop at failure for this run (Figure 3-11b). Unfortunately, a useful estimate of α could not be obtained for this second test because the sample separated across the fracture resulting in substantial loss of fractured rock material.

Using the values of W and α found for test Sq-5 in 4-5 yields a value for the mean fracture energy $\bar{G} = 30 \text{ J/m}^2$. Assuming that the estimate of

α for this sample is representative of all the experiments, the measured value of W for the high strength test Sq-2 gives $\bar{G} = 125 \text{ J/m}^2$. This range of fracture energy values is comparable to those obtained by Friedman et al. (1972), 62 J/m^2 to 124 J/m^2 for Chilhowee quartzite by three-point bending of notched beams. However, they are smaller than those found for Sioux quartzite by Peck and Gordon (1982) using a wedge-loaded double cantilever beam configuration, 180 J/m^2 to 210 J/m^2 . Moreover, these values are all substantially greater than the fracture energies found for quartz single crystals most of which vary from 1 J/m^2 to 10 J/m^2 depending on orientation and the testing method used (e.g., see Atkinson, 1984, Table 3, p. 4006). That experimentally determined fracture energies for rocks are typically much larger than those found for their constituent single crystals is well known. Early attempts to explain this observation appealed to energy dissipation due to plastic deformation at crack tips but it is now generally accepted that this effect is due to distributed microcracking of the host rock surrounding a propagating fracture front which produces an actual fracture surface area that greatly exceeds the apparent, i.e., nominal, area (Friedman et al., 1972; Hoagland et al., 1973). Thus, for example, it is expected that G for quartzite would be roughly equal to the mean value for quartz single crystals if it were possible to measure the true total fracture surface area. In fact, the average fracture energy for single crystals should be an upper bound for polycrystals since G for grain boundary cracks should be smaller than for transgranular cracks.

Considering the effect of microcracking provides a rational basis for comparing the values of fracture energy for quartzite obtained here and in the previous studies cited above. In calculating G for Sioux quartzite

from their experimental data, Peck and Gordon (1982) used a standard compliance calibration formula for the stable propagation of a single, perfectly planar crack: inherent in its use is the assumption that the actual and nominal fracture surface areas are equal. No area correction for microcrack formation was made, although a zone of localized microcracking approximately 0.5 mm wide about the primary fracture was observed in every test. It is therefore not surprising that their values for G are about two orders of magnitude larger than single crystal values. The specific fracture surface energies $\gamma = G/2$ reported by Friedman et al. (1972) for Chilhowee quartzite include an area correction, corresponding to λ^2 in the range 1.2 to 1.7, obtained by measuring just the surface roughness of primary fractures: the correction did not include the surface area of microcracks shown to be present in the surrounding host. However, based on optical and x-ray studies of their samples, these authors suggested that the grain boundary area in a zone approximately 9 mm wide centered on primary fractures was fully cracked leading to an area correction which brought fracture energies for quartzite into fair agreement with single crystal data.

In both of these previous studies, the work W required to produce fracturing was accurately determined using measurements of applied loads and displacements. The principal source of error in determining G was the error in the fracture surface area α . For test Sq-5 in the present study, the converse situation exists. That is, grain-scale mapping of fracture traces has provided a much more accurate estimate for α but W is not as well constrained. Although the average value $\bar{G} = 30 \text{ J/m}^2$ obtained for this sample is probably accurate only to within a factor of 2, because of the error in measuring F_n , it is smaller than any previously reported value for

quartzite. The estimate of α from mapping is clearly a lower bound since some fractured rock material has been lost from the sample creating large holes at several places along the fracture trace (Figures 4-14a and b). Moreover, grain boundary cracks and microscopic surface roughness, e.g. cleavage steps, are not resolved on the present map. When these factors are considered the estimate [4-6] for α could easily be too low by a factor of 3, and perhaps by an order of magnitude, which would place $\bar{G}$ for this test well within the range of values typical of quartz single crystals. While it is not possible to critically evaluate the higher $\bar{G}$ found for test Sq-2, since α could not be reliably estimated for this sample, we speculate that α for test Sq-2 is several times that for Sq-5 which would then result in similar values of $\bar{G}$ for both.

Studies of crack propagation direction and the energetics of crack nucleation and growth are currently being pursued on Sioux quartzite samples extended at $T = 412^{\circ}\text{C}$ and $T = 500^{\circ}\text{C}$. In particular, we will investigate crack surface morphologies to determine whether two scales of crack propagation occur at high temperatures, estimate fracture surface energies at high temperatures and study the distributions and populations of microcracks which may give rise to the large amounts of inelastic work prior to failure.

Figure 4.1. Schematic illustration of mirror, mist, hackle and crack-branching zones on a fracture surface concentrically surrounding fracture origin formed by a surface flaw with radius R_f (after Rice et al., 1984).

Ⓐ TILT BOUNDARY

Ⓑ TWIST BOUNDARY

Figure 4.2. Schematic illustration of the formation of river pattern in cleavage fracturing. Local direction of crack propagation is "down-stream", i.e., in the direction of convergence of cleavage steps (after Henderson, 1975).

Figure 4.3. Surface of tensile fracture in quartz single crystal slab propagated from a surface notch on basal plane (0001) in $[11\bar{2}0]$ direction (after Ball and Payne, 1976).

Figure 4.4. Close-up photograph (a) and map (b) prepared therefrom of fracture surface in modified notched specimen of Sioux quartzite extended at room T and $P_c = 100$ MPa. Shallow saw cut was made at edge in notch mid-plane (bottom) to initiate fracturing. Edges of inclined steps in relief on fracture surface shown as solid lines on map. Dashed lines show perimeters of translucent areas associated with steps, apparently corresponding to the tips of echelon cracks just below main fracture surface. Note distinct boundary between smooth mirror region surrounding cut and outer hackle zone. Most steps are inclined from left to right and are aligned sub-normal to saw-cut indicating direction of macroscopic fracture growth. Local direction of crack propagation was mapped in contiguous series of SEM photos extending across fracture surface, location shown (see Figure 4.5).

Figure 4.5. Map of local crack propagation directions. See text for explanation.

Figure 4.6. Well developed river pattern on transgranular crack surface of quartz grain (close-up view in b). Adjacent grain boundary (pitted patch) parallel to fracture surface was crack nucleus. Local direction of crack propagation indicated by river pattern is unambiguous (large, solid arrow, a) and at least 100° clockwise oblique to overall direction of fracture growth, away from saw cut at left in a (small arrow).

Figure 4.7. Arcuate cleavage steps concave toward a common focus as an example of a high-confidence indicator of local crack propagation direction (arrow).

Figure 4.8. Arcuate cleavage steps concave toward a common focus as an example of a high-confidence indicator of local crack propagation direction (arrow).

Figure 4.9. Two well developed sets of intersecting arcuate steps provide a high-confidence indicator of local crack propagation direction (arrow). Close-up view in b. Interference pattern is highly similar in appearance to Wallner lines suggesting step formation was influenced by interaction of propagating crack tip and obliquely propagating *s*-waves.

Figure 4.10. Arcuate cleavage steps which splay along their lengths to form minute river pattern as an example of a low-confidence determination of local crack propagation direction (arrow). See text for explanation.

Figure 4.11. Weakly curved, irregularly spaced steps provide a low-confidence indicator of local crack propagation direction (arrow).

Figure 4.12. Isolated area, possibly a sub-grain, in which river pattern indicates local direction of crack propagation (small arrow) roughly 45° clockwise oblique to that of surrounding host grain (large arrow) indicated by arcuate cleavage steps. Latter direction was recorded as a low-confidence determination in Figure 4.5.

Figure 4.13. Surface features indicating strongly varying direction of crack propagation within a single quartz grain (small arrows). Patterns suggest multiple crack nucleation sites along grain boundary. Visual average of propagation directions for this grain (large arrow) recorded as a low-confidence determination in Figure 4.5.

Figure 4.14. Mosaic photomicrograph (a) and map (b) prepared therefrom of complete fracture trace in thin section containing cylinder axis of standard notched test specimen Sq-5.

5. ROCK FRICTION AND ANALYSIS OF BOREHOLE STABILITY

Introduction

To predict the drillability and stability of boreholes at the relatively low effective confining pressures but high temperatures that would be associated with the exploration for and production of geothermal energy from shallow magma chambers, we first measured the short-term strengths and ductilities of representative, intact, igneous rocks (andesite, basalt, granodiorite) at 200 MPa or less and up to 900° to 1000°C for partial melting. We concluded that these rocks would be brittle until melting began and holes could therefore be drilled by conventional methods. In basalt and granodiorite, holes would be stable at depths up to 10 km, provided that they could be filled with a fluid with an average density of about 10^3 kg/m^3 (Friedman et al., 1979, 1980, 1982; Bauer et al., 1981). We then showed that for engineering purposes the effects of time (creep rupture or static fatigue) could be neglected (Bauer, 1983; Handin et al., 1981; Kranz et al., 1982; Ross et al., 1983). Finally, we recognized that strength could be substantially degraded by such pre-existing flaws as fractures (or joints). To assess possible effects one must compare the intact strengths with the frictional strengths of these fractures, both open and filled with mineral matter of various kinds. Below we discuss the theory of friction, review previous work and our new data on slip on bare-rock surfaces in a typical rock (Westerly granite), and consider the influence of fractures on predictions of borehole stability.

General

For technical reasons most experimentation at high confining pressure is done on relatively small samples, and although a variety of testing modes has been used (e.g., Jaeger and Cook, 1979; Goodman and Sundaram, 1978), the triaxial-compression test is the most common. The test pieces consist of a cylinder of rock that is either prefractured or contains polished sawcut surfaces which may be separated by a layer of material simulating fault gouge. These configurations reflect an attempt to simulate the many kinds of surfaces present in nature, from freshly fractured surfaces upon which no slip has occurred (such as a tension joint), to old surfaces on which significant sliding has taken place and gouge has formed (Jaeger and Cook, 1979).

Early work primarily investigated the friction of bare-rock surfaces (see reviews by Brace, 1972; Barton, 1973; Logan, 1975, 1978, 1979; Byerlee, 1978a, b; Stesky, 1978b; Paterson, 1978; Jaeger and Cook, 1979). Similar to the frictional properties of other materials, rock friction is markedly dependent on the normal stress across the surfaces of slip.

Frictional strength at lower pressures is much more variable than at higher ones (Byerlee, 1978a; Stesky, 1978b). Rock surfaces move with respect to one another by lifting over the interlocked irregularities at low pressures, so the variation in strength is primarily a result of surface geometry (e.g., Barton, 1973, 1976; Stesky, 1978b). Friction laws that describe strength under these conditions are similar to those for shear of granular materials, and contain terms reflecting the frictional resistance between particles and the dilatancy during slip (Stesky, 1978b). Dilatancy depends on normal stress through relations involving measures of compressive or tensile strengths of the intact rock and empirical constants

that depend on the geometries of the surfaces (e.g., Barton, 1976; Schneider, 1976). This regime is not relevant to our investigation.

At high pressure, above 50 to 100 MPa, rock surfaces move with respect to one another by shearing interlocked asperities, and frictional strength depends much less on initial surface configuration (e.g., Vesic and Clough, 1968; Jaeger, and Cook, 1979; Byerlee, 1978a; Stesky, 1978b). Byerlee's compilation of early experimental data suggests that frictional strength is nearly independent of composition, temperature, and slip rate, and that maximum strength may be adequately predicted by the bilinear relations

$$\begin{aligned} \tau &= 0.85 \sigma_n & \sigma_n < 200 \text{ MPa} \\ \tau &= 50 \text{ MPa} + 0.6 \sigma_n & \sigma_n > 200 \text{ MPa} \end{aligned} \quad [5-1]$$

for all rock types except surfaces separated by layers of simulated gouge composed of certain clay minerals (Fig. 5-1).

The empirical relation 5-1 was used by Brace and Kohlstedt (1981) to place limits on the stresses in the crust. Since that time, this relation has become widely accepted in the geophysical community as a failure criterion rather than as an upper-limit on the strength of the crust. Because the data base from which equation 5-1 was derived has more than doubled, it is appropriate to re-evaluate the generality that is often assumed. More recent work has investigated the effects of (1) temperature, (2) displacement rate, (3) composition of gouge, particularly of phyllosilicates, (4) presence of water or other chemically reactive fluids, (5) and stress/displacement history (Table 5-1). One purpose of this report is to review recent findings, particularly the causes of deviations from the predictions of equation 5-1.

Even if friction laws adequately describe the frictional behaviors of small rock samples in the laboratory, how to apply them to the significantly larger, natural rock-masses important in engineering and geophysics is not clear. This problem has been recognized (e.g., Handin and Logan, 1981) but not received much study except at the lower pressures where sample size is less restricted (e.g., Pratt et al., 1974a, 1974b; Barton and Choubey, 1977; Barton and Bandis, 1978; Bieniawski, 1971, 1978; Heuze, 1980). The greatest value of laboratory testing of small samples at high pressures is to develop mechanistic models of microscopic processes. Such knowledge may provide the basis for valid predictions of macroscopic behavior.

Much of our knowledge of frictional phenomena derives from research on metals (e.g., Bowden and Tabor, 1954, 1964; Rabinowicz, 1965; Kragelskii, 1965). The adhesion theory, described in detail by Bowden and Tabor (1964), explains many of the observations of metallic friction. Because many results from the studies of deformation of metals apply directly to the deformation of rocks, particularly at high pressure and temperature (Carter, 1976; Poirer, 1985), the theory of metallic friction is a good starting point for our discussions. Although rock-forming minerals may deform by some of the same mechanisms as these in metals, they are significantly more brittle and fracture is an important mechanism at the relatively low pressure and high temperature of interest here. Several workers have claimed that theories for metals can not apply to friction in rock. However, as we shall show, many findings from rock-friction studies are at least qualitatively predictable by the adhesion theory. Moreover, the theory is being used more widely to explain time-dependent frictional phenomena of rock (e.g., Bowden and Tabor, 1964; Scholz and Engelder, 1976;

Teufel and Logan, 1978; Dieterich, 1972, 1978, 1979, 1981; Dieterich and Conrad, 1984; Logan and Teufel, 1986).

In this context, our review will briefly outline the adhesion theory, and its predictions of the frictional strengths of rock. With this background, experimental work relevant to characterizing frictional behavior of rock will be reviewed, and the validity of equation 5-1 will be assessed.

Theory of Friction

The shear force F necessary to initiate sliding of a body along an approximately planar surface with apparent contact area A under a normal force W is

$$F = \mu W, \quad [5-2]$$

where μ is the coefficient of friction. Some of the earliest results on friction, now known as Amonton's laws, are (1) frictional force F is related to the normal load W by the proportionality μ , and although μ depends on material properties and states of the contact surfaces, often $1/3 \leq \mu \leq 1$; and (2) friction is independent of the size of the body (Bowden and Tabor, 1964). Restated, μ is approximately independent of A and W . Hence dividing [5-2] by A gives

$$\tau = \mu \sigma_n, \quad [5-3]$$

where σ_n is the normal stress across the contact surface and τ is the shear stress necessary to initiate sliding.

Work on friction during the 18th and 19th centuries involved consideration of both the geometry of the contacting surface and the adhesion at points of contact. However, it was not until 30 or 40 years ago, some 250 years after the time of Amonton, that these ideas were integrated into a theory, able to provide an adequate explanation of Amonton's laws (for brief review of friction research see Bowden and Tabor, Chapter 24, 1964). Most modern work confirms the general idea that friction is largely due to two causes: (1) interfacial adhesion and (2) the energy expended in deforming the surfaces. Thus, the total observed frictional force may be expressed as

$$F = S + P, \quad [5-4]$$

where S is the force to shear adhesive junctions at regions of real contact, and P is the additional ploughing force that arises when asperities penetrate the opposing surfaces and deform the substrate (Bowden and Tabor, 1964). The term P may be further differentiated into (1) a component for strict ploughing, i.e., the penetration of hard asperities or wear particles into the substrate of one or both surfaces, or (2) the mutual deformation of interlocking asperities (see, for example, Suh and Sin, 1981).

The contribution of the ploughing force P to frictional strength may be significant in certain arrangements, however for relatively smooth, unlubricated metal surfaces, and possibly for rock surfaces at high pressure, P is small relative to S and may be neglected. Moreover, the friction based solely on the adhesive forces, i.e., the adhesion theory of friction, can provide a simple and satisfactory explanation of the approximate truth of Amonton's laws as expressed in equations 5-2 and 5-3.

The surfaces of bodies are not perfectly flat even with high polishing, so that when bodies are placed together they initially are only in contact at three points (Kragelskii, 1965). Because all the normal load is carried by only these three asperities, yield strength will commonly be exceeded, and they will deform and cause more asperities to come in contact. This proceeds until the true area of contact A_0 is just sufficient to support the normal load W , thus

$$A_0 = W/p_0, \quad [5-5]$$

where p_0 is the yield stress for a cylindrical asperity of a particular material. Adhesional bonding at the points of contact must be broken for sliding, and if the force to shear these junctions is s , then

$$F = A_0 s, \quad [5-6]$$

and thus

$$F = (s/p_0)W, \text{ where } s/p_0 = \mu. \quad [5-7]$$

This formulation explains the two basic laws of friction, that F is proportional to the load and independent of the apparent area of the surface. To first approximation the average force to shear the junctions is equivalent to the critical shear stress for substrate failure, s_m . For most metals, the critical shear stress for failure is proportional to the yield pressure p_0 . Thus, the absolute magnitude of the critical shear stress for different metals, which is reflected by relative hardness, has

little effect on the coefficient of friction, so for most metals μ is essentially the same (Bowden and Tabor, 1965).

For an axi-symmetric contact of perfectly plastic material, which should provide a good approximation for most metal contacts, p_0 is about five times s_m (Tabor, 1951) and equation 5-7 predicts a μ of 0.2. That this is significantly lower than observed for most metals is a result of the simplification that A_0 is only a function of normal forces W . For perfectly plastic material the application of shear force F causes continued yielding of the asperity and an increase in true contact area. There are no theoretical solutions for failure of a three-dimensional junction, however a relation based on two-dimensional modeling and supported by experiment of the form

$$p^2 + \alpha s^2 = K^2 \quad [5-8]$$

has been suggested by Bowden and Tabor (1964) as a failure criterion. In this case K is a constant that is comparable to the uniaxial yield strength. Empirically fitting this equation to experimental data suggests it may be written as

$$p^2 + \alpha s^2 = \alpha s_m^2, \quad [5-9]$$

where α is between 3 and 25. If the shear strength of the contact junction interface s_i (Fig. 5-2) is expressed in terms of the critical shear strength of the substrate

$$s_i = k s_m, \quad [5-10]$$

where k reflects the quality of the contact and is a numerical factor less than or equal to unity, then equation 5-9 may be expressed as

$$p^2 = \alpha s_i^2 (k^{-2} - 1), \quad [5-11]$$

or

$$\frac{s_i}{p} = \frac{1}{\alpha (k^{-2} - 1)^{\frac{1}{2}}}. \quad [5-12]$$

If at the initiation of sliding the true contact area is A_c , then similarly to equations 5-5 and 5-6 the tangential force $F = s_i A_c$ and the normal load $W = p A_c$, and by equation 5-2 the coefficient of friction is

$$\mu = \frac{F}{W} = \frac{s_i}{p} = \frac{1}{\alpha (k^{-2} - 1)^{\frac{1}{2}}}. \quad [5-13]$$

The variation of μ with k for empirically derived values of α indicates that as interfacial strength approaches substrate strength ($k \rightarrow 1$), $\mu \rightarrow \infty$ (Fig. 5-3). However, as interfacial strength is diminished by only 5% (i.e., $k = 0.95$), μ falls to about unity. If s_i is sufficiently less than s_m (say 20%) then equation 5-13 reduces to 5-7. This formulation retains the ability to explain the basic laws of friction and predicts frictional strengths similar to those observed. For example, the lubrication of metal surfaces inhibits the adhesion between contact junctions, s_i is decreased without affecting s_m , and μ is significantly decreased. The opposite occurs for very clean surfaces, such as those produced in vacuo. Under these conditions surfaces created during sliding

are not oxidized and the quality of the adhesive junctions is very high ($k \approx 1$); thus the interfacial strength s_i approaches that of the substrate s_m (eq. 5-10). Experiments show that in vacuo μ is approximately an order of magnitude greater than under atmospheric conditions as would be predicted by the theory.

As a first approximation the basic assumptions of the adhesion theory are valid for most rock-forming minerals. However, in detail the theory should be modified to accommodate the more brittle behavior of rock as compared to metals. Certainly in rock the true area of contact between surfaces is less than the apparent area, and deformation at points of contact occurs from the normal and shear loading. Direct measurements of surface geometry of natural and artificial sliding surfaces documents roughness at all scales (Scholz and Avilee, 1986). Size, density, and total area of points of contact have been measured on sliding surfaces of limestone and quartzose sandstone in experiments by use of thermodyes (Teufel and Logan, 1978; Logan and Teufel, 1986). True area of contact is found to be in the range of several percent to 40% of the apparent area, and to increase with an increase in normal stress and/or contact time. These observations indicate that some inelastic deformations of the asperities do occur. Stresses at points of contact are calculated to be in the range of the unconfined compressive strength of the material in accord with the predictions of the adhesion theory.

Some mechanisms of deformation in rocks differ from those in most metals. Ductile metals deform by intracrystalline slip, whereas rock deformation may involve ductile and brittle mechanisms. The choice of plastic rheology for modeling asperity deformation in the theory is logical for metals, and may even be a good enough approximation for rock. The

stress conditions around asperities consist of a high mean compressive stress so as to inhibit fracture and promote ductility. These conditions are much like those produced in indenter tests. Study of indentation of rock minerals show that both brittle and ductile mechanisms operate (Hockey and Lawn, 1975; Evans and Goetze, 1979; Evans, 1984), yet the bulk response can be modeled fairly well as plastic (Tabor, 1970; Bowden and Tabor, 1964). Moreover, the flow strengths determined from such tests are comparable to the creep strength of polycrystalline aggregates under high mean-stress conditions to suppress fracture (Rice, 1971; Evans and Goetze, 1979). The critical feature of the model is that asperity failure depends on both normal and shear tractions (eq. 5-8). Other rheologies chosen better to represent rock materials should also show this feature.

One major difference is that high-quality contact junctions in metals grow indefinitely leading to infinitely high friction. For rock however, work-hardening is more important, and asperity growth is limited. Thus for extremely high-quality junctions in rock, asperity growth will occur only until catastrophic failure occurs. For this reason there is an upper limit to frictional strength. This is supported by experiments in vacuo and for extremely clean surfaces (Bowden and Tabor, 1964; Horne and Deere, 1962).

One basic tenet of the theory is that adhesional bonding occurs at points of contact. Direct measurements of adhesional bonding in rock materials is difficult for the same reasons as for hard metals (Bowden and Tabor, 1964). The bonding that occurs at high loads is broken by the elastic recovery of the asperities upon the decrease in load, so it cannot be measured. Adhesional bonding can be measured in more ductile metals because behavior can accommodate the elastic distortion. Adhesional bonding is expected to occur in all solids when brought into intimate

contact by the same atomistic forces that produce the cohesion in the solids proper. The most important factor to affect adhesional bonding is the state of the contacting surfaces. Surface films or oxidation lower surface energy and inhibit adhesion. With some sliding, however, films are destroyed and new surfaces are produced, so high-quality adhesional bonding is both expected and observed.

The most significant prediction of the adhesion theory is that a large component of the coefficient of friction is equal to the ratio of the shear strength of the contact junctions to the strength of the substrate. If applied to rock, the adhesion theory would predict little variation in μ for rocks with different compositions. The variation that does exist should reflect the effects of environmental conditions on the quality of adhesive junctions. Furthermore, the theory would predict that changing extrinsic parameters will not dramatically affect μ if the change affects contact-junction and substrate strengths equivalently. For example, increasing temperature can activate mechanisms of dislocation glide and appreciably lower the intrinsic shear strength of the substrate, but it is likely to affect contact-junction strength in similar ways so that the ratio defining friction will be little changed. Changing extrinsic variables that lower contact-junction strength more than the substrate strength, however, would lower μ significantly. Chemically reactive environments that lower surface energy without significantly weakening the substrate are expected to decrease μ .

The frictional behavior of lamellar solids, such as phyllosilicates, has also received attention. It may also be adequately explained by the adhesion theory (Bowden and Tabor, 1964). During sliding the small platelets in polycrystalline lamellar solids tend to become aligned. The

naturally low surface energy of the cleavage surfaces of the platelets produces low-strength contact junctions, whereas junction growth is controlled by the compressive strength of the polycrystal. The anisotropic properties of the lamellar solids result in a smaller ratio of junction to substrate strengths, and μ is often lower than for non-lamellar solids. Much of the strength of these materials arises from the bonding strength of the edges of the small platelets (Fig. 5-4) in edge-to-face and edge-to-edge contacts (Deacon and Goodman, 1958). Atmospheric conditions have been shown to play an important role in the frictional behavior of these materials because the oxygen and water vapor tends to react with the edges of the platelets, reducing their surface energies (Bowden and Tabor, 1964). This analysis may equally apply to the phyllosilicate minerals of rocks, particularly the clays, which may be reactive with water.

Effects of Composition

After early investigations of the frictional behavior of sawcut or fractured-rock surfaces, as the basis of equation 5-1, studies shifted to the effect of fault gouge on frictional behavior.

In a comprehensive study of monomineralic, simulated gouges of different composition and hardness, Shimamoto and Logan (1981a) showed that although hardness does influence sliding mode, it has little effect on maximum frictional strength (Fig. 5-5). Gouges composed of minerals ranging in hardness from 3 to 9 on the Mohs scale show essentially the same maximum strength, slightly greater than that predicted by equation (5-1). Other studies with quartz and granite gouges also agree with these findings (Engelder et al., 1975; Logan et al. 1979, 1981; Dieterich, 1981; Higgs, 1981; Soldberg and Byerlee, 1984; Dula, 1985). That frictional

strength of rock-forming minerals does not change with different indentation hardness is consistent with the adhesion theory. The only exceptions are the minerals halite, barite, and galena that have hardnesses less than 3 and which are significantly weaker (Shimamoto and Logan, 1981a). The lower strengths in these soft minerals reflect different primary mechanisms of deformation. Frictional slip between grains, along fractures, or along surfaces in these gouges is subordinate to intracrystalline dislocation glide. It is not clear whether these deformations should even be regarded as frictional in nature.

Maximum strengths of mono-mineralic gouges of phyllosilicate minerals do deviate significantly from equation 5-1, as was originally noted by Byerlee (1978a). Phyllosilicate gouges, particularly those composed of clay minerals, have been much investigated (Shimamoto and Logan, 1981b; Summers and Byerlee, 1977; Wang and Mao, 1979; Higgs, 1981; Logan et al., 1981; Morrow et al., 1981; Moore et al., 1983) because these minerals are often present in fractured and faulted rock (Wallace and Morris, 1979; Flinn, 1977; Wu et al., 1975; Wu, 1978; Anderson et al., 1983; Steirman, 1984).

At room temperature and a variety of normal stresses, clay gouges (Fig. 5-6) are significantly weaker than non-clay gouges or bare-rock surfaces (Summers and Byerlee, 1977; Wang and Mao, 1979; Shimamoto and Logan, 1981b; Higgs, 1981; Logan et al., 1981; Morrow et al., 1982; Dengo and Logan, 1981). In general, frictional strength is related to the degree of hydration of the phyllosilicate. The water swelling clays (e.g., smectite) are the weakest, with coefficients of friction of 0.15 to 0.25. Other phyllosilicates with low pressure and temperature stability fields are intermediate in strength, and micas tend to be strongest. The relative

strengths of the phyllosilicates may be correlated with types of bonding at faces and edges of the crystals and/or exchange capacity (e.g., Wu, 1978). That the phyllosilicates have lower frictional strengths than non-phyllosilicates, and that relative frictional strengths may be correlated with atomic structure and water content is consistent with the adhesion theory.

Because most natural fault-zone materials are polymineralic, the effect of mixtures on frictional strength has been investigated (Shimamoto and Logan, 1981; Logan et al., 1981; Higgs, 1981; Chang, 1983; Bauer and Ross, 1985; Rutter et al., 1986; Logan and Rauenzahn, 1986). Studies of bi-mineralic gouges show that only a small amount (<10%) of a stabilizing mineral is necessary to affect frictional sliding mode (Shimamoto and Logan, 1981a, b; Logan and Rauenzahn, 1986), whereas in general, greater amounts (20 to 30%) of a weak or strong component is necessary significantly to affect frictional strength (Shimamoto and Logan, 1981b; Bauer and Ross, 1985; Rutter et al., 1986; Logan and Rauenzahn, 1986). The latter effect (Fig. 5-7) is well illustrated by mixtures of quartz and montmorillonite (Logan and Rauenzahn, 1986).

Testing of more complex polymineralic mixtures has been in the form of natural fault-zone materials collected from along the San Andreas (Logan et al., 1981; Chu et al., 1981; Morrow et al., 1983; Chester and Logan, 1986). These natural gouges contain large portions of clays (>50%). Their maximum strengths are greater than those of pure smectite, but often considerably less than those predicted by equation 5-1 (Fig. 5-8). The variation in strengths observed are difficult to correlate with composition or texture in detail (Morrow et al., 1982; Chu et al., 1981; Chester and Logan, 1986). However, comparisons of all available data do show an inverse relation

between clay content and strength. Logan et al. (1981) test the mineral components of a natural gouge individually and conclude that sliding mode may be controlled by one component while strength may be controlled by another, indicating complex interactions, not yet delineated.

Effects of Temperature

We now review previous work and our own unpublished data. The effects of temperature on frictional properties of most dry rock-forming minerals are relatively small. Sawcut and prefractured surfaces in granite and gabbro have maximum frictional strengths well predicted by equation 5-1 up to confining pressures of 500 MPa and temperatures of 500°C at moderate displacement rates (Byerlee, 1967, 1968; Stesky et al., 1974). The samples are a little weaker at 500° to 700° C (Fig. 5-9), but this effect actually reflects the temperature dependence of shear fracture strengths of the intact samples. Near 500°C the fracture and frictional strengths become about equal, and above 500°C fracture strength is the lesser and deformation is accommodated by widening of the fractured region; i.e., "planes of weakness" become essentially passive and the rock acts like a continuum.

At sufficiently high homologous temperatures thermally activated mechanisms of dislocation motion become important so that lowering of strength can occur. Friedman and Higgs (1981) have demonstrated this effect in calcite gouge (Fig. 5-10). Textures show that the onset of weakening correlates with significant dislocation glide in the gouge; that is, the weakening is associated with a change in mechanism from frictional slip to ductile flow. Similar transitions have been seen in halite gouge by Shimamoto and Logan (1986) and Chester (1986). As with the experiments

of Stesky et al. (1974), the weakening associated with a change in deformation mode is not necessarily associated with a temperature effect on frictional strength per se, but on the competing ductile flow mechanisms in gouge or intact rock.

The lack of a major temperature effect on dry friction is demonstrated by recent experiments by Dula (1985) on quartz gouge between sawcut surfaces of a stronger dioritic rock (Fig 5-11). Frictional strength is essentially independent of temperature to 900°C and normal pressures to 400 MPa. These data agree well with other results on quartz gouge (e.g., Engelder et al., 1975; Higgs, 1981; Shimamoto and Logan, 1981a).

For wet quartz gouge, granite, and sandstone, significant reductions in frictional strength, do occur at either high temperature and moderate displacement rates (Fig. 5-11), or at moderate temperatures and very low displacement rates (Rutter and Mainprice, 1978; Higgs, 1981; Dennis and Atkinson, 1982; Dula, 1985). Figure 5-12 shows that significant weakening does not occur in dry specimens (Rutter and Mainprice, 1978; Higgs, 1981; Dula, 1985), so the effect must be related to the presence of water as discussed below.

Temperature does affect the frictional properties of phyllosilicates, particularly the swelling clays. Raising the temperature appears to increase the coefficient of friction from 0.2 to 0.6 to 0.5 to 0.8 (Fig. 5-13). In drained tests on montmorillonite by Higgs (1981) and Logan et al. (1981) strength increases with time of heating prior to sliding, suggesting that strengthening results from driving off excess water, presumably increasing the surface energy of the platelets and hence bonding strength. These phenomena would be consistent with the adhesion theory. Presumably in undrained tests or in nature where permeability is low, the

water released into the pore space may decrease the effective normal stress, and hence heating may well decrease, not increase frictional resistance.

We have studied the slip on bare-rock surfaces in Westerly granite at the high temperatures (to 900°C), but relatively low confining pressures (200 MPa or less) of interest to the borehole-stability problem to discover whether yield stress still drops more than frictional strength as Stesky et al. (1974) report for higher confining pressures.

Our specimens, 2 by 4-cm, right-circular cylinders, are cut at 35° to the cylinder axis. The precut is then hand-polished with 400 grit. Thus far 24 experiments have been done at temperatures of 200°, 400°, 600°, 700°, 800°, and 900°C, and confining pressures (P_c) of 25, 50, 100, 150, and 200 MPa (Table 5-2). Because the experimental matrix is large, we can evaluate the effects of both confining pressure and temperature at constant axial strain-rate.

At both 200° and 400°C, frictional strength increases with increasing confining pressure. At 600°C, increasing confining pressure from 25 to 100 MPa increases the frictional strength from 145 MPa to 400 MPa. However, further increases in confining pressure to 200 MPa, do not cause further changes. At 700°C, our measurements are poorly reproducible and range from 300 to 500 MPa for 100 to 200-MPa confining pressure. This behavior is probably associated with a brittle-ductile transition. At 800°C, strength increases from 185 MPa ($P_c = 25$ MPa) to 250 MPa ($P_c = 50$ MPa) to 310 MPa ($P_c = 100$ MPa), then decreases to 280 MPa ($P_c = 150$ MPa) to 230 MPa ($P_c = 200$ MPa). The apparent lack of pressure dependence on frictional strength at 800°C may be signaling a predominance of ductile deformation mechanisms. Frictional resistance may well increase with temperature because asperities

flow, and real area of contact grows (cf. Logan, 1975; Logan and Teufel, 1976).

At each of the five confining pressures, increasing the temperature from 200° to 400°C causes an increase in frictional strength. For $P_c = 25$ MPa, further increases to 600° and 800°C result in systematic increases in strength, whereas at $P_c + 50$ MPa, although the strength increases from 400°C to 600°C, a decrease is observed from 600° to 900°C. Frictional strengths at 100 MPa display a slight systematic weakening from 400° to 800°C. At 150 MPa the strength hovers around 450 MPa for tests at 400°, 600°, and 700°C, whereas the strength at 800° drops to 290 MPa. Finally, at $P_c = 200$ MPa, the strength at 400°C (520 MPa) drops to 370 MPa at 600° to 700°C and 230 MPa at 800°C.

Few rock-masses are free of macrofractures (joints). For predictive modeling of the deformations of fractural rocks, for example, of borehole stability to be discussed later, we must know the relative values of frictional strength and fracture or yield strength of the intact material, as well as orientations of fractures relative to the local stress field (Handin, 1967). Comparisons of the data in Table 5-2 with data on intact strengths of a comparable granodiorite (Friedman et al., 1979, 1980; Bauer et al., 1981) reveals that the frictional strength of pre-cut specimens is always less than the strength of intact specimens at $P_c = 50$ and 100 MPa. The difference in strengths does decrease as temperature is increased. For example, at 400°C the frictional strengths are 30 to 40 percent of intact strengths, whereas at 800°C they are 60 to 70 percent. Thus, we have not yet found a regime where slip along pre-existing fractures in dry rock is not potentially important, i.e., where the fractures can be ignored as

passive, at least for the confining pressures of 50 to 200 MPa that are significant to the analysis of borehole stability.

Effects of Fluids

Little systematic work has been done on the effects of fluids and fluid chemistry on rock friction. However, preliminary work suggests such effects may well be significant. The presence of only small amounts of H_2O in the form of atmospheric humidity has been shown to lower the strength of frictional sliding for quartzose sandstone (Dieterich and Conrad, 1984). Sliding of specimens in a dry argon environment (< 1 ppm H_2O) has coefficients of friction of 0.85 to 1.00. Sliding of samples exposed to laboratory humidity or to gas bubbled through H_2O show coefficients of 0.55 to 0.65. Dieterich and Conrad (1984) interpret the strengthening by drying as due to (1) the increased efficiency of adhesion at points of contact by removing contaminate boundary layers that partially interfere with adhesion, or (2) the higher strength of the substrate that supports the shear stress at points of contact, making more difficult the shearing off of irregularities. In addition, drying inhibits time-dependent increases in frictional strength, supporting the idea that increasing friction with time is caused by water-assisted creep of asperities that increases the real area of contact. The time-dependent effects are the more consistent with (1) above, and hence consistent with the adhesion theory.

Saturation with H_2O also substantially weakens quartz gouge and quartz-rich rocks at moderate temperatures and low displacement rates (Fig. 5-12). This is best illustrated by the experiments at 600° to 900°C by Dula (1985) and by the stress-relaxation experiments of Higgs (1981), on quartzose sandstone by Rutter and Mainprice (1978), or granite by Dennis

and Atkinson (1982). The observed weakening must be caused by H_2O because it is not observed in nominally dry samples. The weakening is strongly rate dependent, occurring only at slip rates below 10^{-6} /s at temperatures below $500^{\circ}C$. Although the mechanisms are not yet well known, they are thought to be related to either stress corrosion cracking (e.g., Dennis and Atkinson, 1982; Atkinson and Dennis, 1982; Atkinson, 1984) or diffusive mass transfer processes (Rutter and Mainprice, 1978; Rutter and White, 1979; Rutter, 1983). A model for pressure-solution slip by Rutter and Mainprice (1979) predicts strengths similar to those observed at the lowest rates, and would predict even more dramatic weakening at geologic slip rates.

The deformation mechanisms in the high-temperature experiments of Dula (1985) are also poorly understood. Textures in the wet quartz gouge show distributed deformation with little grain-size reduction, as compared to the highly localized and dramatic grain-size reductions in the dry samples. Moreover, grain boundary migration apparently occurs in the wet gouge. Anyhow, these experiments also appear to fit the category of weakening associated with the onset of some temperature-sensitive, probably thermally activated, mechanism in the host and not with frictional sliding.

If pressure-solution slip is regarded as a frictional process, the weakening is consistent with the adhesion theory because lower frictional strength is achieved by weakening contacts between grains without affecting the strength of the interior of the grains. However, at present the experiments should not be taken to imply real reductions of frictional strength.

Preliminary work on the effect of chemically reactive fluids yields mixed results. Early work by Blackwell (1973), as reported by Logan and

Blackwell (1983), suggests that reactive fluids can reduce the frictional strength of sandstone as much as 25% and that friction is most affected for rocks with low porosity, sandstones with high porosity and permeability being unaffected. This suggestion would appear to be supported by two recent studies by Dunning and Miller (1986) on Berea sandstone and Feucht (1986) on Tennessee sandstone. The former investigated frictional sliding of a high-porosity quartzose sandstone, saturated with a variety of surfactants that have known chemical reactivity with quartz and been shown to affect fracture propagation in quartz (e.g., Dunning et al., 1980, 1984), however, the frictional strength of Berea sandstone was unaffected (Dunning and Miller, 1986).

Feucht (1986) investigated frictional sliding of Tennessee sandstone saturated with fluids with different ionic strengths and pH values. The ionic strength of the pore fluid is controlling under most conditions; the influence of pH is secondary. A maximum weakening of about 45% occurred in samples with low or high ionic strengths for both monovalent and divalent cations. Interestingly, the same solution may weaken a friction surface at low concentrations, but strengthen it at higher concentrations. Even pure H₂O weakens the surfaces by as much as 33%.

At present the effect of chemically reactive fluids on frictional strengths of rocks cannot be predicted in terms of any single parameter, such as zeta potential (Westwood, 1974; Ishido and Nishizawa, 1984) or surface energy (Barenblatt, 1962; Dunning et al., 1984).

Effects of Displacement Rates

Recently studied is the velocity dependence of frictional strength, particularly with the goal of understanding instabilities (e.g., Dieterich,

1972, 1979, 1986; Gu et al., 1984; Rice, 1983; Rice and Gu, 1983; Rice and Ruina, 1980, 1983; Tse and Rice, 1986; Okubo and Dieterich, 1986; Weeks and Tullis, 1985). In general the maximum variation in strength is only about 1% change per decade in sliding velocity (e.g., Dieterich, 1978, 1979a, 1981; Soldberg and Byerlee, 1984; Shimamoto and Logan, 1986; Lockner and Byerlee, 1986). Thus the effect of displacement rate should be negligible for geotechnical applications.

Several studies have demonstrated that considerable sliding is necessary before a consistent, long-term velocity dependent behavior is observed (e.g., Byerlee, 1978; Dieterich, 1979, 1981; Shimamoto and Logan, 1984, 1986). This characteristic is in fact seen of most friction experiments, and is exemplified by the differences in initial and maximum frictional strengths reported in Byerlee (1978) and in Figure 5-5 herein. The effect of load path has been investigated by Shimamoto and Logan (1984, 1986), Logan and Rauenzahn (1986) and Chester (1985). All find that the initial behavior is associated with the compactions of the highly porous gouge-layers. Maximum strengths and long-term sliding behavior are observed only after significant compaction and grain-size reduction has occurred. It is not known what texture or how much displacement is needed most realistically to simulate natural faults, so that the wide ranges of strengths observed in individual experiments cannot be ignored. However, it seems probable that the variations in the rates of loading to initiate slip and of following displacements that are associated with geotechnical problems such as borehole stability may significantly affect the mode (stable sliding or stick-slip), but not the maximum strength (coefficient of "static" friction).

Summary of Frictional Properties

Although the effects of chemically reactive fluids are not yet well understood, it appears that Byerlee's empirical relation (equation 5-1) between frictional strength (shearing resistance) and normal stress is a good enough approximation for igneous rocks under conditions of concern to the borehole-stability problem, regardless of composition, temperature, and loading rate. The only certainly significant exception is that the phyllosilicates are much weaker with μ as low as 0.15 in the swelling clays. Joints in typical igneous rocks may be filled with alteration products like chlorite, and for natural fault gouges, μ may be less than 0.5 (Fig. 5.8).

Analysis of Borehole Stability

Objective and Background. The objective of this section is to examine the influence of a pre-existing fracture on the stability of a deep borehole in hard, crystalline rock. The stability of boreholes in intact rock has been reported previously for this project (Friedman et al., 1979, 1980). Borehole stability has also received considerable recent attention from other investigators who are relating borehole breakouts to in-situ stress (Zoback et al., 1985; Hansen and Purcell, 1986; Kerr 1987; Paillet and Kim, 1987). We note that these previous analyses have considered the rock to be intact and have not accounted for the possible presence of pre-existing fractures. However, Paillet and Kim discuss the possible role of fractures in providing local stress relief and present three examples of televiewer logs showing terminations of continuous breakouts at oblique fractures in the Columbia River Basalt Group.

The important factors in borehole stability are

- (1) the downhole environment, including the in-situ stress state, temperature, pore fluid pressure and chemistry, and excess pressure in the borehole resulting from high density and viscosity drilling muds;
- (2) geometric factors such as the orientation of the borehole with respect to the principal stresses, the cross-sectional shape and straightness of the borehole, and the orientation of the borehole with respect to rock fabric, including anisotropy, and any fractures that may be present; and
- (3) the deformational and strength properties of the intact rock, and the deformational and frictional strength properties of any fractures that may intersect the borehole or be in its neighborhood.

The general in-situ stress state in the upper crust has been studied by many investigators and summarized by Brown and Hoek (1978) who place upper and lower bounds on k , the ratio of average horizontal stress to vertical stress. Thus

$$\sigma_v = 27z, k_{\min} = 0.10/z + 0.3, \text{ and } k_{\max} = 1.5/z + 0.5, \quad [5-14]$$

where σ_v is the vertical stress in MPa, z is depth in km, and $k_{\min}$ and $k_{\max}$ are respectively the minimum and maximum values of $\sigma_{h \text{ av}}/\sigma_v$. Eqs. 5-14 are valid to depths of about 3 km (9100 ft) and indicate that high horizontal stresses typically occur near the surface. If the values k are multiplied by the average vertical stress, one finds that

$$\begin{aligned} \text{Min } \sigma_{h \text{ av}} &= 2.7 + 8.1z, \text{ and} \\ \text{Max } \sigma_{h \text{ av}} &= 40.5 + 13.5 z, \end{aligned} \quad [5-15]$$

which indicate that the average horizontal stress varies linearly with depth at least for depths less than about 3 km. Also we note that the relative influence of the "zero depth" average horizontal stresses declines with depth and the values of k equal to 0.5, 1, and 2 used by Friedman et al. (1979, 1980) are reasonable for depths greater than 1 km and will also be used in this analysis.

Jamison and Cook (1980) have considered measured values for the state of stress in the earth's crust using equations 5-14 and found that the data group according to whether the vertical stress is the maximum, minimum, or intermediate principal stress (corresponding to normal, thrust, and strike-slip faulting), and allowing a coefficient of sliding friction to be estimated for each type of faulting. The combined data lead to a coefficient of friction of 0.522 which is slightly more than the 0.5 lower bound used in our analysis below. Jamison and Cook also discuss whether the average horizontal stress are limited by the actual applied stress or simply by the strength of the rock. The analysis of Jamison and Cook suggests that the cohesion of the crust is negligible (an assumption used in our analysis below) but that the uniaxial strength of intact rock is at least 37.5 MPa.

McGarr (1980) has also considered some constraints on levels of shear stress in the crust from observations and theory down to about 10 km. Stress components are again assumed to vary linearly with depth and have the form $A+Bz$ where A and B are constants that differ for the different components of stress, which implies that deviatoric stress varies with depth. This form is consistent with equations 5-14, and McGarr notes that the maximum shear stress increases with depth and in the upper 3-5 km and there is no indication of a reduction in the gradient with depth.

More recently, Herget (1987) considers the in-situ stress state in the Canadian Shield and also finds that stress components vary linearly with

depth. Herget shows that the upper limit of Brown and Hoek (1978) does not bound data from the Canadian Shield below about 1 km. However, the linear variation of horizontal stress components with depth remains a reasonable representation; only the slope and intercept values are different. Herget suggests a cohesion intercept of 4 MPa and a coefficient of friction of only 0.18 for the shield rocks, significantly lower than the 0.5 used as a lower bound below.

Assumptions and Methodology. We seek a first-order analysis of the influence of a single fracture and, therefore, a number of assumptions are needed to make the analysis tractable. We assume that:

- (1) the state of stress around the borehole is not altered by the presence of the fracture (a common assumption for preliminary analysis in engineering rock mechanics);
- (2) the far-field horizontal principal stresses are equal;
- (3) the Kirsch solution for the stress field around a circular hole in a linear elastic, isotropic, homogeneous material is applicable;
- (4) the influence of possible thermal gradients around the borehole is negligible;
- (5) both vertical and horizontal stresses are principal stress components with magnitudes proportional to depth, pore pressures may exist in the rock mass and the pressure in the borehole may exceed the pore pressure;
- (6) the frictional strength of the fracture plane is proportional to the normal stress acting on the fracture plane, independent of temperature, and the coefficient of sliding friction, μ , ranges between 0.5 and 0.85;

- (7) slip is imminent when shear stress on the fracture exceeds shear strength of the fracture at some point on the borehole periphery; and
- (8) the borehole is vertical and the fracture is dipping with an arbitrary dip angle, $\beta \leq 90^\circ$.

With these assumptions in mind, we first consider a vertical borehole intersected by a fracture dipping with angle β , where $0 < \beta < 90^\circ$. Two cases exist. In Case 1 the maximum principal stress is vertical, and in Case 2 it is horizontal.

In Case 1, the effective stresses σ_1' and σ_2' are given by the following expressions with the Kirsch solution to obtain the tangential stress at the borehole surface;

$$\sigma_1' = 27z - P_p$$

and

[5-16]

$$\begin{aligned}\sigma_2' &= 2 \sigma_h - P_b - P_p \\ &= 2 \sigma_h' - \delta P,\end{aligned}$$

where $\sigma_h = k \sigma_v$, $\sigma_v = 27 z$, P_p is the pore pressure, P_b is the pressure in the borehole, $\delta P = P_b - P_p$ is the excess of borehole pressure over pore pressure, and the prime on a stress component indicates that it is an effective stress. Note that σ_2 is not necessarily the intermediate principal stress but is simply the lesser normal stress component acting in the vertical plane which is tangent to the borehole.

The effective normal stress component and the shear stress acting on the fracture plane with dip angle β are

$$\sigma_n' = (\sigma_1' + \sigma_2')/2 + (\sigma_1 - \sigma_2)/2 \cos(2\beta)$$

and

[5-17]

$$\tau = (\sigma_1' - \sigma_2')/2 \sin(2\beta).$$

In keeping with our assumption of no cohesion, the shear strength of the fracture plane is

$$\tau_o = \mu \sigma_n. \quad [5-18]$$

If we divide equations 5-17 and 5-18 by the maximum shear stress in the vertical plane,

$$\tau_{\max} = (\sigma_1 - \sigma_2)/2, \quad [5-19]$$

we have for normalized shear stress

$$\tau/\tau_{\max} = \sin(2\beta), \quad [5-20]$$

and for normalized shear strength

$$\tau_o/\tau_{\max} = \mu [R + \cos(2\beta)], \quad [5-21]$$

where R , the ratio of mean normal stress to maximum shear stress in the vertical plane, is given by

$$R = (\sigma_1 + \sigma_2) / (\sigma_1 - \sigma_2) \quad [5-22]$$

where $1 \leq R < \infty$. Note that for any dip angle β , the normalized stress and shear strength depend only on the coefficient of friction, μ , and the stress ratio, R , and significantly, not on depth.

Slip can occur on the fracture plane only when the shear stress exceeds the shear strength. Figure 5-14 shows normalized strength and stress versus fracture plane dip angle for values of μ equal to 0.5 (our assumed lower bound) and 0.85 (Byerlee's upper bound), and values of R equal to 2.4 and 4.4 corresponding to the stress conditions shown on Table 5-3. For these parameters, stress does not exceed strength for any value of β and slip is not possible.

In order to determine the maximum value of R for which slip is possible for a particular value of μ in Case 1, we set normalized strength (Eq. 5-21), equal to normalized stress (Eq. 5-20) and solve for R to get

$$R = \sin(2\beta)/\mu - \cos(2\beta), \quad [5-23]$$

which may be differentiated and set equal to zero to find the value of β corresponding to $R_{\max}$. For $\mu = 0.5$, we find $\beta = 58.3^\circ$ and $R_{\max} = 2.24$ while for $\mu = 0.85$, we find $\beta = 65.2^\circ$ corresponding to $R_{\max} = 1.54$. Note that for $\mu = 0.5$, the curve corresponding to $R = 2.4$ nearly touches the strength curve at about 58° because R is just slightly greater than the maximum value of R for slip, 2.24.

In Case 2, the maximum principal stress is horizontal and equal to the tangential stress from the Kirsch solution, Then,

$$\sigma_1' = 2 \sigma_h' - \delta P$$

and

$$\sigma_2' = 27z - P_p.$$

[5-24]

Now the angle between the maximum normal stress and the normal stress on the fracture plane is $90^\circ - \beta$ and equation 5-21 for the normalized shear strength becomes

$$\tau_0/\tau_{\max} = \mu [R - \cos(2\beta)], \quad [5-25]$$

and the remaining equations stay the same.

Figures 5-15 and 5-16 show the normalized strength and stress curves for the Case 2 stress conditions from Table 5-3 for $\mu = 0.5$ and 0.85 , respectively. As expected, for both values of μ , slip occurs for smaller values of the stress ratio R , but is suppressed for larger values of R . Again we calculate the maximum value of R for which slip is possible and the corresponding value of β : for $\mu = 0.5$, $R_{\max} = 2.23$ at $\beta = 31.7^\circ$, while for $\mu = 0.85$, $R_{\max} = 1.54$ at $\beta = 24.8^\circ$.

For a vertical fracture ($\beta = 90^\circ$), slip can occur on the fracture, and we consider the case where the tangential stress is σ_1 and the radial stress is σ_2 , noting again that σ_2 is the lesser of the two principal stresses in the horizontal plane and not necessarily the intermediate principal stress. For the geometry shown in Figure 5-17, which defines the position of the fracture relative to the center of the borehole using the angle α , the effective normal and shear stresses on the fracture plane are

$$\sigma_n' = [(\sigma_1' + \sigma_2')/2 + (\sigma_1' - \sigma_2')/2 \cos(2\alpha) - \delta P]$$

and [5-26]

$$\tau = (\sigma_1' - \sigma_2')/2 \sin(2\alpha),$$

but now $\delta P = \sigma_2'$ which allows us to simplify the first of equations 5-26 to

$$\sigma_n = \tau_{\max} [1 + \cos(2\alpha)]. \quad [5-27]$$

The normalized shear stress and strength equations are then

$$\tau/\tau_{\max} = \sin(2\alpha)$$

and

[5-28]

$$\tau_0/\tau_{\max} = \mu [1 + \cos(2\alpha)].$$

Figure 5-18 shows the curves for normalized shear stress and strength as a function of α for $\mu = 0.5$ and 0.85 . Note that for the vertical fracture, a failure zone always exists if $\alpha > \alpha_{cr}$ which defines the beginning of the slip zone. Note that α_{cr} is independent of the stress state and depends only on μ . By setting the ratio of strength-to-stress equal to unity, and using standard trigonometric identities, we show that

$$\tan(\alpha_{cr}) = \mu.$$

[5-29]

Equation 5-29 allows us to calculate $\alpha_{cr} = 26.6^\circ$ for $\mu = 0.5$ and $\alpha_{cr} = 40.4^\circ$ for $\mu = 0.85$. Figure 5-19 confirms that these values of α represent the beginning of the zone where shear stress on the fracture exceeds shear strength.

Discussion and Conclusions. (1) For a fracture dip angle $\beta < 90^\circ$, stability depends on two parameters, the coefficient of sliding friction, μ , and the ratio, R , of mean normal stress, $(\sigma_1 + \sigma_2)/2$, to maximum shear stress, $(\sigma_1 - \sigma_2)/2$, but stability does not depend on depth. (2) As expected, for a constant value of μ , the unstable range of dip angle β decreases and ultimately disappears as R increases. (3) For a constant value of R , the unstable range of dip angles decreases and finally disappears for larger values of μ . (4) The stress ratio, R , decreases as pore pressure increases, which leads to an increase in the size of the unstable range of the dip angle, β , conversely, decreases in pore pressure tend to stabilize the fracture and lead to a more stable borehole. (5)

Increasing borehole pressure above pore pressure by using denser drilling muds, decreases the stress ratio, R , which leads to an increase in the size of the unstable range of the dip angle, β , and thus has a destabilizing effect on borehole stability, which is in contrast to the case of a borehole in intact rock where increasing the mud density leads to a more stable wellbore as long as the excess borehole pressure does not cause hydraulic fracturing. (6) For a vertical fracture, $\beta = 90^\circ$, an unstable region exists for all stress states as long as the fracture does not pass through a region near the center of the borehole which is defined by the angle, α_{cr} . The angle, α is defined as the angle between a line which is parallel to the vertical fracture plane and passing through the center of the borehole and the intersection between the borehole periphery and the fracture plane. The angle, α_{cr} , depends only on the coefficient of sliding friction, μ , according to the relationship

$$\tan(\alpha_{cr}) = \mu.$$

Again, increasing the density of the mud does not stabilize the fracture.

Table 5-1. Experimental Rock-Friction (1978-1986)

Reference	Type of Test		σ_n	$\dot{d}$	Temp	Parameters Investigated			Fluids	Final Fabric
	Bare Rock	Gouge				Composition	Mixtures	Texture		
Stesky (1978a)	X			X	X					X
Teufel & Logan (1978)	X			X						
Rutter & Mainprice (1978)	X		X	X	X				X	
Dieterich (1978)	X		X	X				X		
Logan et al (1979)		X								X
Wang & Mao (1979)		X	X			X				
Dieterich (1979)	X			X				X		
Ruina (1980)	X			X						
Chu et al (1981)		X	X			X		X		
Morrow et al (1981)		X							X	
Dengo & Logan (1981)	X		X					X		X
Johnson (1981)	X			X						
Friedman & Higgs (1981)		X			X					X
Logan et al (1981)		X			X	X	X			X
Higgs (1981)		X		X	X	X	X	X	X	X
Dieterich (1981)		X		X				X		
Shimamoto & Logan (1981a)		X	X			X	X			
Shimamoto & Logan (1981b)		X	X			X	X			
Morrow et al (1982)		X	X			X		X	X	
Chang (1982)		X		X		X	X			X
Dennis & Atkinson (1982)	X		X	X	X				X	
Logan & Blackwell (1983)	X					X			X	
Moore et al (1983)		X	X		X	X				
Shimamoto & Logan (1984)		X	X							
Dieterich & Conrad (1984)	X			X					X	
Byerlee & Vaughn (1984)		X		X					X	
Hungr & Morgenstern (1984)		X		X				X		
Solborg & Byerlee (1984)		X	X	X						
Weeks & Tullis (1985)	X			X		X				
Shimamoto (1985a)	X	X	X							
Shimamoto (1985b)	X	X	X	X	X				X	
Feucht (1985)	X								X	X

Reference	Type of Test		σ_n	$\dot{\epsilon}$	Temp	Parameters Investigated			Fluids	Final Fabric
	Bare Rock	Gouge				Composition	Mixtures	Texture		
Dula (1985)		X	X	X	X				X	X
Chester (1985,86)		X	X	X						X
Chester & Logan (1986)		X				X			X	
Dunning & Miller (1986)	X								X	
Lockner et al. (1986)		X	X	X	X					
Logan & Teufel (1986)	X		X			X				
Logan & Rauenzahn (1986)		X		X		X	X	X		X
Moore et al. (1986)		X		X	X	X				X
Rutter et al (1968)		X				X	X			X
Shimamoto & Logan (1986)		X	X	X						X
Tullis & Weeks (1986)	X		X	X						
Shimamoto (1986)		X	X	X						

Table 5-2. Results of Friction Experiments on Bare-Rock Surfaces in Dry Westerly Granite at Different Temperatures (T) and Confining Pressures (P_c) at a Constant Axial Strain-Rate of 10^{-4} s^{-1}

Exp. No.	T (°C)	$P_c = \sigma_3$ (MPa)	At slip	At termination		At slip
			$\sigma_1 - \sigma_3$ (MPa)	$\sigma_1 - \sigma_3$ (MPa)	Strain (%)	τ/σ
F1	200	25	40	115	1.4	0.49
F2	200	50	130	220	2.0	0.66
F3	200	100	355	395	1.5	0.77
F4	200	200	465	500	1.7	0.62
F5	400	25	75	120	0.8	0.71
F6*	400	50	180	190	0.6	0.77
F7	400	100	325	425	0.8	0.74
F8	400	150	465	465	0.6	0.72
F9	400	200	525	525	0.4	0.66
F10*	600	25	145	145	1.0	0.94
F11*	600	50	230	200	0.7	0.86
F12	600	100	325	400	1.5	0.74
F13	600	150	445	450	2.0	0.71
F14	600	200	390	330	1.5	0.56
F15	600	200	355	325	1.5	0.53
F16	700	100	260	350	1.6	0.66
F17	700	150	495	490	1.7	0.74
F18	700	200	370	325	1.7	0.54
F19	800	25	160	185	0.9	0.97
F20	800	50	240	255	1.2	0.87
F21	800	100	320	320	1.3	0.73
F22	800	150	280	215	1.2	0.54
F23	800	200	235	165	1.7	0.40
F24	900	50	120	130	0.9	0.63

*Stick-slip was observed in these three specimens only.

Table 5-3. Borehole wall effective stresses and approximate slip region for fracture dip angle, $\beta < 90^\circ$.

Case	$k=\sigma_h/\sigma_v$	P_p/z	$\delta P/z$	σ_1'/z	σ_2'/z	Approx. Slip R	Slip Region, β	
							(-)	(+)
		(MPa/km)	(MPa/km)	(MPa/km)	(MPa/km)	(-)	$\mu=.5$	$\mu=.85$
1	0.5	0	0	27	27	∞	None	
	0.5	10	0	17	7	2.4	None	
	0.5	0	10	27	17	4.4	None	
	0.5	10	10	17	-3	N/A	None	
2	1	0	0	54	27	3.00	None	
	1	10	0	34	17	3.00	None	
	1	0	10	44	27	4.18	None	
	1	10	10	24	17	5.86	None	
	2	0	0	108	27	1.67	10.8-52.5	None
	2	10	0	88	17	1.48	7.2-55.8	None
	2	0	10	98	27	1.76	12.6-51.3	None
	2	10	10	78	17	1.56	9.0-54.9	None

*Case 1 : $\sigma_1' = \sigma_v' = 27z - P_p$

Case 2 : $\sigma_1' = \sigma_\theta' = 2\sigma_h' - \delta P$

**R = $(\sigma_1' - \sigma_2')/(\sigma_1' + \sigma_2')$

INITIAL FRICTION

MAXIMUM FRICTION

Figure 5-1. Shear stress versus normal stress for initial (A) and maximum (B) friction for a variety of rock types (taken from Byerlee, 1978a).

Figure 5-2. Model of asperity junction (taken from Bowden and Tabor, 1964).

Figure 5-3. Variation of μ with k for several values of α (taken from Bowden and Tabor, 1964).

Figure 5-4. Schematic diagram of interactions of edges and cleavage faces in the sliding of lamellar solids (taken from Bowden and Tabor, 1964).

Figure 5-5. Shear stress versus normal stress for monomineralic gouges of hardness between 3 and 9. (A) Initial frictional strength taken at 2.5% axial shortening. (B) Maximum frictional strength for same experiments. Data from Shimamoto and Logan, 1981a.

Figure 5-6. Shear stress versus normal stress for maximum frictional strength of monomineralic clay gouges. Gouge in all experiments is nominally dry. Source of data indicated.

Figure 5-7. Coefficient of friction versus quartz content in mixed quartz-montmorillonite gouge at 50-MPa confining pressure (taken from Logan and Rauenzahn, 1986).

Figure 5-8. Shear stress versus normal stress for polymineralic, natural fault gouges. All experiments are room dry. Source of data indicated.

Figure 5-9. Frictional and fracture strength of San Marcos gabbro and Westerly granite at elevated temperatures. (A) Shear stress versus normal stress for experiments at 25° to 700°C. (B) Comparison of frictional strength and fracture strength at 400 and 500 MPa confining pressure (taken from Stesky et al., 1974).

Figure 5-10. Ratio of shear stress and normal stress versus temperature for calcite gouge. Data from Friedman and Higgs (1981).

Figure 5-11. Shear stress versus normal stress for quartz gouge, wet and dry, at elevated temperatures. Source of data indicated.

Figure 5-12. Coefficient of friction versus strain rate for prefRACTURED Tennessee sandstone, wet and dry, from stress relaxation tests by Rutter and Mainprice, 1978.

Figure 5-13. Coefficient of friction versus temperature for clay-bearing gouge. Source of data indicated.

Figure 5-14. Normalized shear stress and strength of fracture plane versus dip angle β for values of μ and R from Table 5-3, Case 1 - $\sigma_1' = \sigma_v'$.

Figure 5-15. Normalized shear stress and strength of fracture plane versus dip angle β for $\mu = 0.5$ and values of R from Table 5-3, Case 2 - $\sigma_1' = 2\sigma_h' - \delta P$.

Figure 5-16. Normalized shear stress and strength of fracture plane versus dip angle β for $\mu = 0.85$ and values of R from Table 5-3, Case 2 - $\sigma_1' = 2\sigma_h' - \delta P$.

Figure 5-17. Horizontal cross-section of a vertical fracture plane intersecting a borehole and definition of angle α .

Figure 5-18. Normalized shear stress and strength of fracture plane versus location angle α for two values of μ .

6. DEFORMATION EXPERIMENTS ON BIOTITE SINGLE CRYSTALS

Oriented biotite single crystals have been deformed at temperatures of 19° and 400°C, and confining pressures of 100 to 300 MPa at strain rates of $2.3 - 2.5 \times 10^{-5}$ /s in an effort to complete our studies of basal slip as a function of temperature within the ductile field and delineate the transition from brittle deformation and frictional sliding to ductile deformation by dislocation slip. Biotite deforms by dislocation glide at confining pressures as low as 100 MPa, and the results of experiments, currently in hand, are consistent with previous studies at higher confining pressures (Kronenberg et al., 1985a, b).

Prismatic single crystals have been shortened at 45 degrees to their basal plane (Figure 6.1) using Ag jackets with cylindrical outer diameters and interior dimensions machined square to fit the samples. Jacket strengths were determined by running duplicate experiments on single crystals of NaCl using the same jacketing arrangement and comparing the measured loads with previously published mechanical results for NaCl (Carter and Heard, 1970). The results of deformation experiments on biotite, and the jacket strength experiments are listed in Table 6.1. The load records for biotite samples have been corrected for jacket strength, for apparatus distortion, and for the increase in cross-sectional area with strain. Corrected stress-strain curves (Figure 6.2) are characterized by a steep, initial, elastic slope, a sharply-defined yield point, and continued deformation at nearly constant stress at strains greater than 1 to 4%.

The combined results of this study at 400°C (Figure 6.3), the frictional sliding measurements of Horn and Deere (1962), and previous

studies of biotite deformation at higher confining pressures (Kronenberg et al., 1985a, b) suggest that the transition between frictional sliding and dislocation slip occurs at confining pressures of approximately 50 MPa. Due to experimental difficulties with the apparatus used for these experiments, however, and particularly with the furnace feed-throughs and ceramic insulation, we have not been able to complete the experiments needed to further refine this transition.

In addition to the pressure dependence of biotite deformation, we have defined further the temperature dependence of basal slip within the high pressure, ductile field. Combining results of this study at $T = 19^\circ\text{C}$ and $T = 400^\circ\text{C}$ (Figure 6.4), we have determined the activation energy for basal dislocation glide in the crystallographically nearly equivalent (Figure 6.5) strong directions $[010]$ and $[310]$, and compare this with that determined previously (Kronenberg et al., 1985a) for the weak slip directions $[100]$ and $[110]$. Within experimental error, the activation energy for these stronger shear directions of $\sim 33 \text{ kJ/mol}$ is in excellent agreement with the activation energy of $\sim 30 \text{ kJ/mol}$ reported previously for $[100]$ and $[110]$ slip.

On the basis of these and previous experiments performed on biotite single crystals, we have made advances in our understanding of the slip mechanisms in biotite by applying micromechanical theories of dislocation glide to the mechanical results of experiments performed in different orientations. The weakest directions in shear on the basal plane (Figure 6.5) are $[100]$ and $[110]$. However, strain energy considerations suggest that $[100]$ and $\frac{1}{2}[110]$ dislocations may split up into pairs of partial

dislocations, separated by narrow stacking faults, by such dislocation reactions as

$$[100] \rightarrow \frac{1}{6}[310] + \frac{1}{6}[310] \quad (6-1)$$

and

$$\frac{1}{2}[110] \rightarrow \frac{1}{3}[010] + \frac{1}{6}[310] \quad (6-2)$$

In addition, dislocations may interact to produce dislocation nodes leading to hardening effects for basal shear in the $[310]$ and the $[010]$ directions.

We have attempted to test these interpretations using a Phillips 400T transmission electron microscopy (TEM) at an accelerating voltage to 120 KV. However, attempts to image biotite thin foils have not met with success due to rapid electron beam damage. Damage rates have been so rapid (within times of 3 sec) that observations have been extremely difficult and image contrast experiments impossible. While the use of our newly-acquired cryogenic TEM specimen holder may cut down substantially on electron beam damage rates, we have only recently received this unit and have not had the opportunity to use it on these samples. While we cannot, at this time, report on TEM of the experimentally-deformed biotite samples, we plan to investigate them in the near future. Finally, if damage rates are still too rapid at liquid N_2 temperature, we may need to investigate the slip systems of muscovite, which apparently has slower damage rates (Wilson, personal communication), as an analogue material to biotite.

Table 6.1. Results of Biotite Deformation Experiments

Experiment Number	Sample and Orientation	Jacket Material	T (°C)	P _c * (MPa)	$\dot{\epsilon}$ (s ⁻¹)	$\sigma_1 - \sigma_3$ [†] (MPa)	ϵ ^{††} (%)	Comments
TP-1	Westerly granite with cored center	Ag	517°-532°C at nominal 525°C	20	0	0	0	Temperature gradient along full length of 0.600" sample
BI-1	Biotite single crystal C.A.** <45°{001} <45°[310]	Ag	400	200	2.37x10 ⁻⁵	60	11	
NA-1	NaCl single crystal C.A.** $\perp$ {100}	Ag	400	200	2.46x 10 ⁻⁵	40	11	Jacket strengths determined using NaCl data of Carter and Heard (1970)
BI-2	Biotite single crystal C.A.** <45°{001} <45°[310]	Ag	400	100	2.34x10 ⁻⁵	16	7	
BI-3	Biotite single crystal C.A.** <45°{001} <45°[010]	Ag	19	300	2.34x10 ⁻⁵	120	8	
NA-2	NaCl single crystal C.A.** $\perp$ {100}	Ag	19	300	2.32x 10 ⁻⁵	140	12	
BI-4	Biotite single crystal C.A.** <45°{001} <45°[010]	Ag	400	50	2.47x10 ⁻⁵	0	15	Jacket leak

* P_c = σ_3

** C.A. refers to maximum compression axis

† Measured at $\epsilon = 5\%$ (BI# experiments corrected for jacket strength)

†† Total strain at end of experiment

Figure 6.1. Sample assembly and loading configuration for biotite deformation experiments. Prismatic single crystals have been prepared using techniques developed in previous studies (Kronenberg et al., 1985a) and were sealed in fitted Ag jackets (with cylindrical outer diameters and machined square inside).

SAMPLE PREPARATION

- 1) Impregnate Biotite XI w/ Lakeside Cement
- 2) Oriented rectangular Samples cut w/ Wire Saw
- 3) Assemble w/o sealing
- 4) Heat Treatment $T = 400\text{ C}$ $t = 24\text{ hrs}$
 $(\text{Lakeside} \rightarrow \text{C} \Rightarrow \text{C} + \text{O}_2 \rightarrow \text{CO}_2 \uparrow)$
- 5) Seal Ag jacket w/ WC rings

Figure 6.2. Corrected stress-strain curves for biotite samples #BI-1, BI-2, and BI-3. The compression axis (C.A.) was oriented at 45 degrees to $\{001\}$ in all experiments to maximize the resolved shear stress for basal slip. Samples BI-1 and BI-2 were furthermore oriented with C.A. at 45° to the $[310]$ direction, whereas C.A. was at 45° to $[010]$ (crystallographically nearly identical) for sample #BI-3.

Figure 6.4. Arrhenius plot of differential stress at $\epsilon = 5\%$ for biotite single crystals shortened at 45° to {001} where $1/T$ refers to the reciprocal of the absolute temperature in K. Best fit lines to the combined results of crystals oriented for shear in the directions [010] and [310], and the combined results for shear directions [100] and [110] have remarkably similar slopes corresponding to values of Q/nR . Taking $n = 11$ from previous studies (Kronenberg et al., 1985a), activation energies Q ($\sim 30 \text{ kJ/mol}$) for slip in the strong ([010] and [310]) and weak ([100] and [110]) orientations are in excellent agreement.

Figure 6.5. Principal slip directions in biotite include $[100]$, and $\frac{1}{2}[110]$, and symmetrically equivalent directions. Samples sheared in intermediate directions $[310]$ (between $[100]$ and $\frac{1}{2}[110]$ shown) and $[010]$ (between $\frac{1}{2}[110]$ and $\frac{1}{2}[\bar{1}10]$ shown) are significantly stronger and are likely to require the operation of more than one slip system.

7. SUMMARY AND CONCLUSIONS

CHRONOLOGICAL SUMMARY OF RESULTS

1974-1979 - Contract No. 13-2242, Sandia National Laboratories

Energy extraction from magma requires stable boreholes at relatively shallow depths (< 10 km) in rocks at temperatures approaching partial melting. Accordingly, the failure strengths, strains at failure, and associated deformation mechanisms were determined at temperatures to partial melting (> 1050 °C) and a strain rate of 10^{-4} s^{-1} for (a) room-dry Mt. Hood andesite, Cuerbio basalt, and Charcoal granodiorite at confining pressures of 0, 50, and 100 MPa, (b) H_2O -saturated specimens of these same crystalline rocks at zero effective confining pressure (pore and confining pressures both of 50 MPa), and (c) room-dry Newberry rhyolite obsidian at 0 and 50 MPa confining pressure.

The major conclusions from the experimental and petrofabric results as are as follows:

(1) The ultimate compressive strengths of the dry crystalline rocks decrease gradually with increasing temperature to about 900°C , and then more rapidly as partial melting is approached at between 1000° and 1100°C . The strengthening effect of increasing effective confining pressure diminishes with increasing temperature. The obsidian is even stronger than the

crystalline rocks up to 600°C, above which, however, the glass softens and its strength vanishes at 800°C.

(2) There is pronounced water-weakening only of the Mt. Hood andesite and that occurs between 850° and 880°C (zero effective pressure). At 870°-880°C its ultimate strength is only 5 to 25 MPa compared to an average room-dry strength of 75 MPa at 1000°C. The Cuerbio basalt may be weakened slightly between 800° and 900°C, but the Charcoal granodiorite is unaffected by the presence of water to 1000°C.

(3) The crystalline rocks are essentially brittle throughout the pressure and temperature ranges explored until partial melting occurs. They tend to deform primarily by precursive microscopic extension fracturing and its coalescence into macroscopic faults. The abundance of load-induced fractures in dry specimens remains about constant, but thermal cracking increases with increasing temperature. Results from tests at 25°C on specimens that previously had been heated to 900-1000°C clearly show that the weakening of the unconfined dry specimens is due to the thermal cracking. Weakening of confined dry specimens is due to an inherent temperature effect on the load-induced fracturing process. Each wet specimen also fails along multiple macroscopic shear fractures or along a discrete fault. Weakening of the andesite is inferred to be due to hydration of Si-O bonds which causes stress-corrosion cracking or a lowering of the melting point. Significant water-weakening of the andesite occurs in contrast to the other two crystalline rocks, apparently because its greater porosity (10-12 percent) and assumed corresponding permeability permits greater access of the H₂O to internal surfaces.

(4) There is some evidence of small ductile flow, particularly adjacent to faults. We see some gliding flow (kinking) in the biotite and

thermal alteration of biotite, hornblende and pyroxene contributing to the strength reduction by an unknown, but probably minor amount. Shortenings at failure are ≤ 3 percent in room-dry tests and ≤ 1 percent in wet ones. Virtually all the permanent shortening is ascribable to macroscopic faulting.

(5) At confining pressures to 100 MPa the degradation of strength and continued brittle behavior with increasing temperature suggest that the crystalline rocks will be drillable by conventional methods up to temperatures of partial melting. In fact, rapid decreases in strength at the highest temperatures suggest that the penetration rate ought to increase as the roof rock-magma interface is approached.

(6) Time-dependent flow characteristics were obtained for the dry obsidian through stress-relaxation tests at 0 and 50 MPa, 10^{-4} to 10^{-6} s^{-1} , and between 700° and 800°C (i.e., in its glass-softening phase). Unconfined equivalent viscosities decrease from 3.4×10^{11} at 700°C to about $3 \times 10^{10} \text{ N}\cdot\text{s}/\text{m}^2$ at 800°C. At 50 MPa, they decrease with increasing temperature along a slightly steeper slope, but the effect of the pressure increment is insignificant. A flow law of the type $\dot{\epsilon} = A \exp(-Q/RT) (\sigma_1 - \sigma_3)^n$ fit the data between 700° and 800°C with $A = 2 \times 10^{13}$ (when σ is in MPa and $\dot{\epsilon}$ is in s^{-1}), $Q = 380 \text{ kJ/mol}$, $n \approx 1.5$, and R is the gas constant. Our data on viscosities and Q are in very good agreement with those of Murase and McBirney (1973) on the same obsidian. The dramatic decrease in strength between 600° and 800°C precludes natural glass as a possible borehole liner to enhance stability, at least for boreholes at temperatures above 600°C.

(7) Several facts strongly suggest that the temperature-weakening of the room-dry crystalline rocks is due to intrinsic effects of temperature on the processes leading to macroscopic shear fracturing (or faulting), and

not merely to degradation of strength by thermal cracking, at least when the rocks are subjected to confining pressure. (a) Brittle fracture accounts for virtually all the permanent strain. (b) The degree of microfracturing associated with the mechanical loading, as distinguished from the thermoeleastic stressing, is about the same at all temperatures from 25° to 1000°C. (c) Specimens heated to 900°C, cooled, and then deformed at 25°C are at least as strong as those that have not been preheated.

On the other hand, similar thermal cycling greatly degrades the strengths of the unconfined rocks, presumably because the effect of the thermal cracks on subsequent failure is not suppressed by confining pressure; that is, heating significantly modifies the mechanical state of the material. Thus, data from unconfined tests alone cannot be used for valid predictions of rock properties under in situ conditions, particularly for geological or geophysical applications. However, such tests may be useful to estimate strengths during drilling where essentially unconfined conditions and rapid cooling may obtain.

Several workers have regarded fracturing as a thermally-activated process, and they have derived an apparent activation energy (see Carter and Kirby, 1978). Owing to the lack of data on time effects, we cannot do so. However, even if brittle fracturing is indeed rate-controlled, how could we, or the previous workers for that matter, associate the purely empirical, apparent activation energy with a specific process? Is it the initiation and/or propagation of intragranular microfracturing? Is it grain-boundary sliding? Is it the condensation of microfracturing in preferred zones and/or the precursory coalescence that appears to cause the instability manifest in the macroscopic fracturing, stress drop, and work-softening that characterize the ultimate failure of these rocks? We

felt that a lot more research on fracture had to be done before a meaningful constitutive "law" could be written for inelastic deformations in the brittle state.

(8) Water-weakening in the andesite, in light of its dominantly brittle behavior, probably is caused by hydrolytic weakening of silanol bonds in the silicate minerals which promotes stress-corrosion cracking and reduces melting temperatures. The andesite, rather than the basalt or granodiorite, is water-weakened probably because its greater porosity (10 to 12 percent) permits more pervasive contact between water and solid internal surfaces than in the other two rocks.

(9) Because the deformations of the rocks are essentially cataclastic at all temperatures, we suppose that boreholes in these rocks will fail in shear when the differential stress reaches the ultimate compressive strength for a given mean stress. In general, the maximum depth for stable boreholes increases with decreasing temperature and decreasing σ_h/σ_v ; and for dry rocks, filled boreholes are more stable than open ones. Further, for similar conditions boreholes in granodiorite will be stable to greater depths than those in basalt or andesite, with the latter providing the least stable ones. Assessment of stability for a given rock type is, therefore, primarily a function of assumed or known borehole stress and temperature conditions. For examples: for σ_h/σ_v of 1.0, filled boreholes (medium with specific gravity of 1.0) in dry impermeable rock are apt to be stable to > 10 km even at temperatures of 900°C or higher. On the other hand, in open boreholes and the same stress ratio, boreholes in dry andesite are likely to fail at 2, 3, 4, or 5 km depth for temperatures of 900°, 700°, 400°, and 25°C, respectively. For dry rocks, the σ_h/σ_v ratio of 0.5 provides the greatest stability of the three ratios considered, but for water-saturated

rocks and filled boreholes this same ratio provides a "worst-case" condition because the tangential borehole stress becomes tensile. Accordingly, stability of filled boreholes in saturated rocks would require smooth walls and temperatures no hotter than 600°C. On the other hand, even though the andesite is weakened by water at high temperature, it should sustain stable boreholes in open holes to depths > 10 km and temperatures to 880°C if the effective pressure is zero and σ_h/σ_v is 1.0.

Our predictions of borehole stability are optimistic because we had not accounted for stress concentrations due to irregularities in the borehole walls, the frictional sliding strength of pre-existing fractures intersecting the borehole and dangerously inclined to the borehole stresses, or the kinetics of water-weakening. Moreover, we focussed initially on the instantaneous strengths and ductilities of the rocks with borehole stability predictions pertinent only to relatively short periods of time immediately after penetration by the drill. Long-term stability of the boreholes seems probable, at least in light of the few time-dependent strength data available in the literature. However, data under appropriate conditions were meager and pertinent testing had just begun.

1980 - Contract No. DE-A505-79ER10361, Task 3 DOE/OBES

During this first year of work on Task 3 of our three-year contract with DOE/OBES, we (1) completed and later published our petrographic studies of dry and wet specimens of crystalline rocks deformed in short-time experiments at temperatures to partial melting (Friedman et al., 1982). (2) We were asked by Matt Salisbury of the NSF Deep Sea Drilling Project at Scripps Institution of Oceanography to test a few samples of basalt from a drillhole in crystalline rock. He was concerned about the

marked decline in drilling rates with increasing depth subsea and wondered whether this parameter might correlate with the triaxial-compressive strength, expected to increase with increasing overburden pressure. Knowing of the interest of DOE/OBES in continental deep-drilling and believing that results might be generally applicable to engineering problems of common interest, we tested 15 specimens. We did indeed find a good correlation between drilling rate and strength, and we understand that our data have been used in discussions with drillbit manufacturers on how to improve drilling rates in dense crystalline rocks like basalt (Bauer and Handin, 1985). (3) We concluded from tests on water-saturated specimens that (a) the granodiorite does not water-weaken at $P_e = 0$ and 50 MPa; (b) the more porous and finer-grained andesite begins to water-weaken at 600°C and between 870° and 920°C its strength is reduced by factors of 5 and 4 at $P_e = 0$ and 50 MPa, respectively; (c) both rocks are at least as brittle when deformed wet as when dry, with the coalescence of microfractures into one or more faults the major mechanism of deformation; and (d) the melting temperature for the andesite is decreased $\approx 150^\circ\text{C}$ in the presence of water. Water-weakening is probably caused by the hydration of silanol bonds that lowers fracture strengths (stress corrosion) and melting temperatures.

Measurements of relative linear thermal expansions of the rocks revealed that the granodiorite, andesite, and basalt at $P_e = 0$ and 50 MPa and over the temperature range of 25° to 900°C the granodiorite undergoes more expansion at the lower P_e and more total expansion at both pressures than the finer-grained andesite and basalt. Moreover, the expansion in the granodiorite accelerates at the $\alpha - \beta$ transition of quartz (573°C). These results are consistent with a greater degree of thermal cracking in the

granodiorite compared to that in the much finer-grained andesite and basalt (Bauer and Handin, 1983).

Extrapolations of strength and ductility data for the intact specimens indicate: (1) crystalline rocks should be drillable because they remain brittle until partial melting occurs, and penetration rates should increase as temperatures increase because there is a corresponding decrease in brittle fracture strength; (2) instantaneous failure strengths are apt to be more important to the assessment of borehole stability than transient creep of intact rock; (3) boreholes in "water-filled" holes will be stable to >10 km at temperatures less than T_m ; (4) if temperatures are kept to $\leq 700^\circ\text{C}$, even unfilled (open) boreholes in granodiorite should be stable to ≥ 10 km; and (5) open boreholes in the andesite are expected to be much less stable, and at similar temperatures would fail at 1 to 3 km depth.

The deformation in these crystalline-rock specimens in multimechanistic; microfracturing of apparent extensile and shear origin, glide in quartz and biotite, microfracture healing, dissolution, and mineral alteration are mechanisms observed and evaluated as functions of temperature and strain. We attempt to partition the bulk specimen deformation into that contributed by the individual mineral species in the analysis. The systematic change in micromechanisms observed with increasing temperature is compatible with the origin of the gradational succession of macroscopic deformation (Friedman et al., 1982).

Work to date on intact specimens supported the scientific feasibility of drilling to buried magma chambers at depths to 10 km.

1981

During the reporting period, work on short-term strengths and ductilities of water-saturated Mt. Hood andesite and Charcoal granodiorite was extended to higher effective pressures of 100 MPa. Measurements of thermal expansions of these rocks and the Cuerbio basalt at effective pressures of 5 and 50 MPa and temperatures to 800°C were completed to provide basic data for thermomechanical modeling. Times to failure of wet and dry Westerly granite at an effective confining pressure of 100 MPa and temperatures of 300°, 400°, 600°, and 800°C were determined to assess the stability of boreholes and other underground openings for long periods of time.

Further consideration of data from water-saturated specimens of the granodiorite and andesite compared to room-dry counterparts show that the pore pressures are communicated throughout each test specimen so that the pressures are effective. At effective confining pressures (P_e) of 0 and 50 MPa the granodiorite does not water-weaken; at these same effective pressures, the more porous and finer-grained andesite begins to exhibit water-weakening at about 600°C; at $P_e = 0$ and 870° - 900°C, the andesite's strength averages 20 MPa whereas dry specimens at the same P and T exhibit a strength of 100 MPa; at $P_e = 50$ MPa and 920°C, the andesite wet strength is 45 MPa compared to 160 MPa dry. The ductility of the water-saturated specimens deformed at temperatures less than that of melting exhibit ultimate strengths at less than 3% shortening and then work-soften along faults; that is, like their dry counterparts the rocks remain essentially brittle up to the onset of melting. Both dry and wet specimens deform primarily by microscopic fractures that coalesce into one or more macroscopic faults, and the temperature for incipient melting of the andesite

decreases about 150°C in the water-saturated state. Our conclusions about borehole stability did not change.

Measurements of mean coefficient of linear thermal expansion (α) were completed for these rocks (Bauer and Handin, 1983). Differences between measured and calculated values are related to new crack porosity (ϕ) resulting from anisotropic thermal expansions. For the granite a two or three order of magnitude increase in permeability (k) is predicted from the relation $k \sim \phi^3$.

Westerly Granite is weaker and more ductile when deformed wet in creep tests than in our short-term experiments (Bauer, 1982). The weakening increases with increasing pore pressure at a given effective pressure. To determine if the assessment of borehole stability needed revision, we extrapolated the times to failure for the extreme case (800°C). Using the expected borehole stresses at 8 and 10 km and $\sigma_h/\sigma_v = 1$, we concluded that a borehole in granite would remain stable 10^4 and 1 years, respectively. Thus our assessment of stability did not change significantly.

1982

The mechanical problem of deep drilling, as well as others associated with high-temperature crustal environments, cannot be fully resolved without adequate knowledge of both short-term and time-dependent rock strength and ductility. Accordingly, during this period we emphasized the static-fatigue tests on room-dry and water-saturated Westerly granite at 100-MPa effective confining pressure and 300 to 1800°C that were begun during the previous year. The deformation is multimechanistic; microfractures of extensile and shear origin, gliding in quartz and biotite, microfracture healing, dissolution, and mineral alteration are mechanisms

observed and evaluated as functions of temperature and strain. The bulk specimen deformation is partitioned into that contributed by the individual mineral species. The systematic change in micromechanisms with increasing temperature is compatible with the origin of the gradational succession of macroscopic deformation modes from a single fault ($T/T_m \leq 0.9$).

Flow data for 600°C and below (brittle field) are well fit by an equation of the form:

$$t_{\text{failure}} = t_0 p^{-\alpha} \exp (E/RT - K\sigma),$$

where t_0 , α , and K are experimental constants, P is the partial pressure of H_2O , E is the activation energy, σ is the differential creep stress, R is the gas constant, and T is the absolute temperature. This equation had been shown to predict the static fatigue time for many rocks. Using this expression, we calculated apparent activation energies for the total fatigue process in the α -quartz field of about 30 kJ/mole (300-400°C, dry) and 90 kJ/mole (400-600°C, wet).

Quasi-steady state flow occurs in the β -quartz field (700-800°C) and the flow behavior is well described by $\dot{\epsilon} = A \exp (-Q/RT) \sigma^n$, where the activation energy $Q = 290$ kJ/mole, $n = 2.7$, $\dot{\epsilon}$ is the strain rate, σ is the steady-state differential stress, and the constant $A = 25.2$. These results are significant because previously flow had not been so characterized where fracture (cataclasis) is a major, if not the dominant, deformation mechanism. Our results show that fracture is a time-dependent, thermally activated process.

1983

To provide a physical basis for the empirical flow laws, it was deemed necessary to investigate in detail the thermal-activation and stress-corrosion aspects of fracture in wet and dry quartzite, granite, and basalt in extension tests, wherein true macroscopic tensile fractures could be induced at a constant strain rate of 10^{-4}s^{-1} and less over a temperature range of 25° to 800°C, under effective confining pressures of 20 to 200 MPa and absolute pore-water pressures of 10 to 180 MPa.

Furthermore, borehole-stability problems in fractured rock require study of the high-temperature (300-900°C), low pressure (10-100 MPa) frictional behavior of the pre-existing fractures (joints). We began with smooth, bare-rock surfaces in granite, granodiorite, andesite, and basalt to determine under what conditions of P and T frictional strengths might be equal to the strengths of the unfractured, intact material. That is, under what conditions can the rock mass be modelled as a continuum?

With regard to our previous conclusions about borehole stability we noted that (1) creep-rupture strength of granite is only about 20 percent below the short-time strength so that for geotechnical applications the effect of time may be ignored; (2) results from twenty-four frictional-sliding tests indicate that the frictional strength of smooth surfaces of granite is only about 50 percent of the intact strength at all temperatures between 400° and 800°C; (3) twenty-two triaxial-extension tests show that the tensile strength of granite drops from about 10 MPa at 25°C to sensibly zero at 800°C. (The data have not been corrected for the considerable stress concentrations in the "dogbone-shaped" extension specimens. However, the relative values are regarded as significant. Later extension

tests have been done on botched cylindrical specimens in which the axial stress concentration is only about 20 percent.)

1984

In continuation of our studies of borehole stability and the fracture of rock in environments of hot fluids, we investigated the (1) tensile strengths of Sioux quartzite in extension tests to follow up our similar work on Westerly granite, (2) fracture surfaces and microfracture abundances in the extension experiments in an effort to understand why strength decreases with increasing temperature and how fractures may be thermally activated, and (3) frictional strength variation with temperature of mineralized fractures, as opposed to that of smooth rock surfaces previously studied.

We tentatively concluded that:

(1) Tensile strengths for both the quartzite and granite appear to be essentially independent of confining pressure, as expected from earlier work by Handin et al. (1967).

(2) The strength of Westerly granite decreases with increasing temperature especially in the presence of water.

(3) Ductility increases significantly with increasing temperature only in wet tests (at constant effective pressure of 100 MPa).

(4) Macroscopic shear fracture tends to be favored over tensile fracture by increasing temperature or confining pressure.

(5) Distinct patterns of grain-boundary cracks and intragranular microfractures are developed and associated with both macroscopic tensile and shear failure. The main controlling parameter seems to be confining

pressure which suppresses the grain-boundary cracking and promotes ordered and abundant intragranular microfractures.

(6) In dry specimens at 150 MPa confining pressure and in wet ones at 100 MPa effective pressure, decreases in strength are accompanied by decreases in the abundance of microfractures, mainly the intragranular ones. Apparently fracture initiation, propagation and coalescence occur at lower inputs of distortional strain energy with increasing temperature and when wet. That is, it takes less mechanical energy for the fracture process to occur as temperature increases, and supports the view that fractures are thermally activated.

(7) When due consideration is given to the moisture content and pore pressures with gouges of different composition it is clear that there is a drop in frictional strength with increasing temperature. This is best observed in wet gouges under constant effective confining pressure.

(8) If the country rock is impermeable enough or if dehydration generates pore pressure quickly enough, the effective normal stress in the gouge can be reduced, thus lowering frictional sliding strength.

(9) Comparisons of the frictional-sliding strengths of bare-rock surfaces of Tennessee sandstone with the dry gouge-filled ones at 150°C and 50 MPa confining pressure, show that gypsum and anhydrite gouges are about 25% and 15% weaker, respectively, and that dry serpentine gouge (at both 150°C and 400°C) exhibits a strength equal to or slightly stronger than that of the bare-rock surfaces.

1985

Studies continued on the strengths of Sioux quartzite and Westerly granite including fracture surfaces and microfracture types, abundances and

coalescing tendencies in extension experiments in an effort to understand why strength decreases with increasing temperature, particular when the specimen is water-saturated, and if fracture is a thermally activated process. We began study of the role of biotite crystals as the stress risers and facilitators of slip leading to the semi-brittle failure of granite.

Five extension tests of notched Westerly granite specimens were performed at room temperature and confining pressures between 30 and 145 MPa. Tentative corrections to the measured differential forces at catastrophic failure give an average tensile strength of about 20 MPa and confirm the very important finding that the tensile strength of a rock is independent of the confining pressure (P_c) up to the point where $P_c = C_0$, the unconfined compressive strength.

Experimental data-reduction procedures to accurately calculate the state of stress in the extension experiments with notched specimens were re-examined and an analysis of the change in elastic strain energy in these specimens was initiated.

Our microscopic observations of the extended granite specimens reveals that (Hopkins, 1986):

(1) In dry tests, intragranular cracks (IGC) and grain-boundary cracks (GBC) increase in linear and areal density (abundance) with increasing T at constant confining pressure (P_c); IGC increase in abundance with increasing P_c at constant T ; and IGC tend to be shorter with increasing T .

(2) In wet tests, IGC decrease with increasing T ; GBC tend to be better developed with increasing T (as perhaps expected in that water

occurs primarily at the grain boundaries); and the IGC tend to be shorter with increasing T.

(3) The common thread is the enhancement of grain-boundary cracks with increasing T. The GBC play a greater role in the crack coalescence history leading to failure as T increases. For wet conditions, the GBC are best developed and consequently fewer and shorter IGC are required to achieve coalescence and failure. For dry conditions, the GBC are only slightly better developed with increasing T so that more IGC, even though shorter, are needed to affect macroscopic failure.

(4) Perhaps the most surprising result of this study is the indication that the length of IGC decreases with increasing T, wet or dry.

1986-Present

Conclusions warranted from the work completed during this period are listed by section, this report, as follows.

Section 2 - Strength and Microfracturing in Extended Westerly Granite.

In addition to the information on microfractures in the extended Westerly granite specimens, mentioned immediately above, conclusions from further evaluation of the data are as follows:

1. The strength of Westerly granite specimens deformed in triaxial extension decreases significantly with increasing temperature with a disproportionately greater decrease occurring at the higher temperatures. The effect of temperature is equally important at tensile and compressive stress states such that the ratio of the biaxial compressive strength to the tensile strength remains constant at 16 at each temperature (25°, 600°, and 800°C).

2. Water greatly enhances the temperature effect as it enhances the development of grain-boundary cracks.
3. Microstructural analyses indicate microfracturing is the dominant deformation mechanism at all temperatures. Two distinctive microcrack patterns occur. Pattern A is characterized by mechanically induced intragranular cracks (IGC) strongly oriented normal to σ_3 and grain boundary cracks (GBC) that in the main are not completely developed about each grain. Pattern B is characterized by more randomly oriented IGC and wider and more completely developed GBC. These patterns exhibit systematic relationships with temperature and stress state at failure, with pattern B being favored at higher temperatures and particularly by the presence of water.
4. Observational data and comparisons with room-temperature strength data of thermally-cycled specimens indicate that the observed decrease in tensile and unconfined compressive strength at temperature primarily reflects the effects of thermal cracks, i.e., both IGC and GBC produced by differential thermal expansion of neighboring grains rather than by the applied mechanical loads.
5. Presumably, the coalescence of microcracks into through-going macrofractures is enhanced by the development of thermal cracks, particularly along grain boundaries. Easy coalescence leads to failure at lower differential stress levels, i.e., lower strengths as temperature increases particularly in the presence of water.

Section 3 - Extension Experiments on Sioux Quartzite and Westerly Granite

1. Maximum differential loads supported by samples of Sioux quartzite extended at $\dot{\epsilon}$ rates of 5.2×10^{-6} to $6.7 \times 10^{-4} \text{ s}^{-1}$, a confining pressure P_c of 100 MPa, and temperatures T of 25° to 500°C are insensitive to temperature and strain rate.
2. Tensile fracture strengths of Sioux quartzite at these conditions (~ 11 MPa at $T = 25^\circ\text{C}$ and ~ 4.5 MPa at $T = 500^\circ\text{C}$) are comparable to unconfined tensile strengths (of 10 - 11 MPa at $T = 25^\circ\text{C}$) reported by Krech et al. (1974).
3. Prior to tensile failure, samples of Sioux quartzite extended at temperatures of 25° to 412°C exhibit nearly perfect linear elastic behavior.
4. In contrast, Sioux quartzite samples extended at $T = 500^\circ\text{C}$ exhibit significant deviations from linear elasticity prior to failure.
5. Corresponding to the change in mechanical response prior to failure, calculated values of inelastic work done on Sioux quartzite samples prior to failure show a sharp increase from $\sim 0.22\text{J}$ at $T = 25^\circ - 412^\circ\text{C}$ to $\sim 1.15\text{J}$ at $T = 500^\circ\text{C}$.
6. Tensile fracture strengths measured for Westerly granite, using methods identical to those employed for the quartzite experiments, are, within experimental uncertainty, comparable to those reported by Bauer (1983) and Krech et al. (1974).
7. As contrasted with Sioux quartzite, specimens of Westerly granite extended at room temperature exhibit significant nonlinearity prior to failure.

Section 4- Fractography and Microcrack Distribution in Extended Sioux Quartzite

1. Optical and SEM examination of fracture surfaces of experimentally extended samples of Sioux quartzite has revealed two scales of fracture propagation:

- A) Macroscopic zones comprised of smooth (mirror) and rough (hackle) fracture surface indicate a consistent direction of overall fracture propagation across the specimen notch.
- B) On a finer scale, however, fracture propagation inferred from surface features on individual quartz grains show a nearly random distribution of propagation directions.

2. The two scales of fracture propagation are associated with the development of the macroscopic fracture surface and with the process zone of microcracking. Even within the zone of stress concentration associated with the induced fracture origin, the local direction of crack propagation appears to be controlled by local critical flaws, such as grain boundaries and microcracks initially present.

3. Interpretation of these fracture surfaces suggest a mechanism of catastrophic fracture propagation in quartzite similar to that for stable fracturing of various quartz-bearing crystalline rocks (ie. propagation effected by localized growth and coalescence of parallel microcracks within a narrow zone about the primary fracture front.

4. An estimate of the fracture energy for Sioux quartzite ($G = 30 \text{ J/m}^2$) has been made, based upon the inelastic work done on the sample and total fracture surface area created (including microcracks of the process zone), which is smaller than values previously reported for quartzite, and closer to G typical for quartz single crystals (1 to 10 J/m^2).

Section 5 - Rock Friction and Analysis of Borehole Stability

1. Although the effects of chemically reactive fluids are not yet well understood, it appears that Byerlee's empirical relation ($\tau = 50 \text{ MPa} + 0.6\sigma_n$) between frictional strength (shearing resistance, τ) and normal stress (σ_n) is a good enough approximation for igneous rocks under conditions of concern to the borehole-stability problem, regardless of rock composition, temperature and loading rate.
2. The only certain exception to (1) is that the phyllosilicates are much weaker with μ as low as 0.15 in the swelling clays. Joints in typical igneous rocks may be filled with alteration products like chlorite and clays. Accordingly, for natural fault gouges, μ may be less than 0.5.
3. Borehole stability in fractured rock has been analyzed from a general point of view and with the following simplifying assumptions:
 - a. the state of stress around the borehole is not altered by the presence of the fracture (a common assumption for preliminary analysis in engineering rock mechanics),
 - b. the far-field horizontal principal stresses are equal,
 - c. the Kirsch solution for the stress field around a circular hole in a linear elastic, isotropic, homogeneous material is applicable,
 - d. the influence of possible thermal gradients around the borehole is negligible,
 - e. both vertical and horizontal of stresses are principal stress components with magnitudes proportional to depth, pore pressures may exist in the rock mass and the pressure in the borehole may exceed the pore pressure,
 - f. the frictional strength of the fracture plane is proportional to the normal stress acting on the fracture plane (Byerlee's universal law of friction), is independent of temperature, and the coefficient of sliding friction, μ , ranges between 0.5 and 0.85,

- g. slip is imminent when shear stress on the fracture exceeds shear strength of the fracture at some point on the borehole periphery, and
- h. the borehole is vertical and the fracture is dipping with an arbitrary dip angle, $\beta \leq 90^\circ$.

4. Results of the analysis indicate the following:

- a. for a fracture dip angle $\beta < 90^\circ$, stability depends on two parameters, the coefficient of sliding friction, μ , and the ratio, R , of mean normal stress, $(\sigma_1 + \sigma_3)/2$, to maximum shear stress, $(\sigma_1 - \sigma_3)/2$,
- b. as expected, for a constant value of μ , the unstable range of dip angle β decreases and ultimately disappears as R increases,
- c. for a constant value of R , the unstable range of dip angles decreases and finally disappears for larger values of μ ,
- d. the stress ratio, R , decreases as pore pressure increases, which leads to an increase in the size of the unstable range of the dip angle, β ; conversely, decreases in pore pressure tend to stabilize the fracture and lead to a more stable borehole,
- e. increasing borehole pressure above pore pressure by using denser drilling muds, decreases the stress ratio, R , which leads to an increase in the size of the unstable range of the dip angle, β , and thus has a destabilizing effect on borehole stability, which is in contrast to the case of a borehole in intact rock where increasing the mud density leads to a more stable wellbore as long as the excess borehole pressure does not cause hydraulic fracturing, and
- f. for a vertical fracture, $\beta = 90^\circ$, an unstable region exists for all stress states as long as the fracture does not pass through a region near the center of the borehole which is defined by the angle, α_{cr} . The angle α_{cr} is defined as the angle between a line which is parallel to the vertical fracture plane and passing through the center of the borehole and the intersection between the borehole periphery and the fracture plane. The angle, α_{cr} , depends only on the coefficient of sliding friction, μ , according to the relationship

$$\tan(\alpha_{cr}) = \mu.$$

Again, increasing the density of the mud does not stabilize the fracture.

Section 6- Deformation of Biotite Single Crystals

1. Biotite single crystals, oriented for basal slip, deform by dislocation glide at confining pressures as low as 100 MPa, with transitional behavior between frictional gliding and slip at ~50 MPa.
2. The mechanical response of biotite single crystals deformed at high confining pressures is characterized by a steep elastic slope, a rather sharply-defined yield point, and continued deformation at nearly constant stress at strains greater than 1 to 4%.
3. Within the high pressure, ductile field, biotite crystals oriented for shear in the crystallographically nearly identical directions [010] and [310], while significantly stronger, exhibit the same temperature dependence ($Q \approx 33\text{kJ/mol}$) as those oriented for shear in the directions [100] and [110] ($Q \approx 30\text{kJ/mol}$).
4. Application of micromechanical dislocation theory to the results for biotite single crystals suggests that slip occurs on the systems $\{001\}\langle 100 \rangle$ and $\{001\}\langle 110 \rangle$ but that these dislocations may be split into paired partial dislocations.

CODA

Thus, we have investigated the strengths and deformation mechanisms in a variety of typical igneous rocks under conditions likely to be encountered in geothermal wells, particularly those drilling into buried magma bodies. Fundamental insights were gained into why rock strengths decrease with increasing temperature, in the presence of water and as a function of time even in the brittle regime. Based on the experimental data we predicted borehole stability as a function of depth, borehole pressure and the ratio of the far-field vertical and horizontal stresses, initially for homogeneous intact rocks and including the influence of time.

As a result of our review of rock friction and the analysis of borehole stability in fractured rock (Section 5) we are now in a position to work on the integration of these parameters into an overall more comprehensive prediction of borehole stability. While, in principle, such an integration is possible, the time and effort required for the effort is beyond the scope of this expired grant. It is work that needs to be done and supported.

8. ACKNOWLEDGEMENTS

The essential element in any experimental investigation is the quality of the apparatus, the devotion toward its calibration and maintenance, the care with which students are trained to use the apparatus and prepare samples, and the dedication to accuracy during the experiments. For their efforts in these areas we wish to thank Jack N. Maquirk and Buckey L. Turk, research specialists in the Center for Tectonophysics, Texas A&M University. Their help has been all important in maintaining continuity of the experimental program over the years.

We also wish to thank Dr. John M. Logan for discussions concerning the design of the apparatus used in this work and for overseeing fabrication of modifications and replacement parts. Conversations with Dr. Neville L. Carter on deformation mechanisms also were helpful. In addition, members of the Magma Energy Research Advisory Committee, Sandia National Laboratories, made valuable comments and suggestions during the early phases of the work.

Most importantly, however, we are very grateful to the Office of Basic Energy Sciences, U.S. Department of Energy, particularly Dr. George A. Kolstad, for support and encouragement.

9. REFERENCES

- Amadei, B. and M. J. Robison, Strength of rock in multiaxial loading conditions, Proc. 27th U.S. Symp. Rock Mech., 155-164, 1986.
- Anderson, J. L., R. H. Osborne, and D. F. Palmer, Cataclastic rocks of the San Gabriel fault--An expression of deformation at deeper levels in the San Andreas fault zone, Tectonophysics, 98, 209-251, 1983.
- Anderson, O. L. and P. C. Grew, Stress corrosion theory of crack propagation with application to geophysics, Rev. Geophys., 15, 7/-104, 1977.
- Atkinson, B. K., Subcritical crack growth in geological materials, J. Geophys. Res., 89, 4077-4114, 1984.
- Atkinson, B. K. and S. M. Dennis, Experimental constraints on the mechanisms of water-induced weakening of fault zones in crustal rocks, Earthquake Prediction Res., 1, 349-376, 1982.
- Bahat, D. and T. Engelder, Surface morphology on cross-fold joints of the Appalachian Plateau, New York and Pennsylvania, Tectonophysics, 104, 299-313, 1984.
- Ball, A. and B. W. Payne, The tensile fracture of quartz crystals, J. Mat. Sci., 11, 731-740, 1976.
- Barenblatt, G. I., The mathematical theory of equilibrium cracks in brittle fracture, Adv. Appl. Mech., 7, 55, 1982.
- Barton, N., The shear strength of rock and rock joints, Int. J. Rock Mech. Min. Sci., 13, 255-279, 1976.
- Barton, N., Review of new shear criterion for rock joints, Eng. Geol., 7, 287-332, 1973.
- Barton, N. and S. Bandis, Effects of block size on the shear behavior of jointed rock, Proc. 19th U.S. Symp. Rock Mech., Univ. Nevada, Reno, 739-760, 1978.
- Barton, N. and V. Choubey, The shear strength of rock joints in theory and practice, Rock Mech., 10, 1-54, 1977.
- Bauer, S. J., Brittle-ductile transition in dry and water-saturated Westerly Granite (Abstract), Am. Geophys. Union Trans., 61, 437, 1982.
- Bauer, S. J., Semibrittle Deformation of Granite at Upper Crustal Conditions, Ph.D. Dissertation, Texas A&M University, 1983.
- Bauer, S. J., B. Johnson, Effects of slow uniform heating on the physical properties of the Westerly and Charcoal granites, Proc. 20th U.S. Symp. Rock Mech., Univ. Texas, Austin, 7-18, 1979.

- Bauer, S. J., M. Friedman, and J. Handin, Effects of water saturation on strength and ductility of three igneous rocks at effective pressures to 50 MPa and temperatures to partial melting, Proc. 22nd U.S. Symp. Rock Mech., MIT, 73-78, 1981.
- Bauer, S. J. and J. Handin, Thermal expansion of three water-saturated igneous rocks at 800°C at effective confining pressures of 5 and 50 MPa (Abstract), Am. Geophys. Union Trans., 62, 393, 1981.
- Bauer, S. J. and J. Handin, Thermal expansion and cracking of three, confined, water-saturated rocks to 800°C, Rock Mechanics and Rock Engineering, 16, 181-198, 1983.
- Bauer, S. J. and J. V. Ross, Textures and rheologies of synthetic anhydrite/halite mylonites, Trans. Am. Geophys. Union, 64, 839-840, 1973.
- Bieniawski, Z. T., Determining rock mass deformability: experience from case histories, Int. J. Rock Mech. Min. Sci., 15, 237-247, 1978.
- Bieniawski, Z. T., Propagation of brittle fracture in rock, Proc. 10th U.S. Symp. Rock Mech., Univ. Texas, Austin, 409-430, 1972.
- Blackwell, M. L., The Influence of Pore Fluids on the frictional properties of quartzose sandstone, M.S. Thesis, Texas A&M University, 1973.
- Bowden, F. P. and D. Tabor, Friction and Lubrication of Solids, 1, Clarendon Press, Oxford, 1964.
- Bowden, F. P. and D. Tabor, Friction and Lubrication of Solids, 2, Clarendon Press, Oxford, 544 pp., 1964.
- Brace, W. F., Brittle fracture of rocks, Proc. Intern. Conf. on State of Stress in Earth's Crust, 111-174, Elsevier, New York, 1964.
- Brace, W. F., Micromechanics in rock systems, in Structure, Solid Mechanics, and Engr. Design, Edited by M. Te'eni, Wiley and Sons, London, 1971.
- Brace, W. F., Laboratory studies of stick-slip and their application to earthquakes, Tectonophysics, 14, 189-200, 1972.
- Brace, W. F., Recent laboratory studies of stick-slip and their application to earthquakes, Phys. Earth., 25, 185-202, 1971.
- Brace, W. F. and D. L. Kohlstedt, Limits on lithospheric stress imposed by laboratory experiments, J. Geophys. Res., 85, 6248-6252, 1981.
- Bridgman, P. W., Studies in Large Plastic Flow and Fracture, 362 pp., McGraw-Hill, New York, 1952.
- Broek, D., Elementary Engineering Fracture Mechanics, Noordhoff, Leyden, The Netherlands, 1974.

- Brown, E. T. and E. Hock, Trends in the relationships between measured in-situ stresses and depth, Int. J. Rock Mech. Min. Sci., 15, 211-215, 1978.
- Byerlee, J. D., Frictional characteristics of granite under high confining pressure, J. Geophys. Res., 72, 3639-3648, 1967.
- Byerlee, J. D., Brittle ductile transition in rock, J. Geophys. Res., 73, 4741-4650, 1968.
- Byerlee, J. D., Friction in rocks, Pure and Appl. Geophys., 116, 615-626, 1978a.
- Byerlee, J. D., A review of rock mechanics studies in the United States pertinent to earthquake prediction, Pure Appl. Geophys., 116, 586-602, 1978b.
- Byerlee, J. D. and P. Vaughn, Dependence of friction on slip velocity in water saturated granite with added gouge, Trans. Am. Geophys. Union, 65, 1078, 1984.
- Carter, N. L., Steady-state flow of rock, Rev. Geophys., 14, 301-360, 1976.
- Carter, N. L., and H. C. Heard, Temperature and interdependent deformation of halite, Am. J. Sci., 269, 193-249, 1970.
- Carter, N. L. and S. H. Kirby, Transient creep and semibrittle behavior of crystalline rocks, Pageopa., 116, 807-839, 1978.
- Carter, N. L., D. A. Anderson, F. D. Hansen, and R. L. Kranz, Creep and creep-rupture of granitic rocks, Mechanical Behavior of Crustal Rocks, Am. Geophys. Union, Monograph. 24, 61-82, 1981.
- Chang, H. C., Effect of displacement rate and gouge composition on the sliding behavior of simulated gouge, M.S. Thesis, Texas A&M University, 69 pp., 1982.
- Chester, F. M., Correlation of halite gouge texture with sliding mode and velocity dependence in experimental faults (Abstract), Trans. Am. Geophys. Union, 66, 1100-1101, 1985.
- Chester, F. M., Relation of mechanical behavior and mode of failure in semibrittle shear zones, Geol. Soc. Am. Abstracts with ranes Prog., 18, 564, 1986.
- Chester, F. M. and J. M. Logan, Implications for mechanical properties of brittle faults from observations of the Punchbowl fault zone, California, in: Internal Fault Zone Structure edited by C-Y, Wang. Pure Appl. Geophys., 124, 79-106, 1986.
- Chu, C. L., C-Y. Wang, and W. Lin, Permeability and frictional properties of San Andreas fault gouges, Geophys. Res. Ltrs, 8, 565-568, 1981.

- Dengo, C. A., and J. M. Logan, Implications of mechanical and frictional behavior of serpentinite to seismogenic faulting, J. Geophys. Res., 86, 10771-10782, 1981.
- Dennis, P. F. and B. K. Atkinson, The influence of water on the stress supported by experimentally faulted Westerly Granite, J. Geophys. Roy Astron. Soc., 71, 285-294, 1982.
- Dieterich, J. H., Time-dependent friction and the mechanics of stick-slip, Pure Appl. Geophys., 116, 790-806, 1978.
- Dieterich, J. H., Modeling of rock friction. 1. Experimental results and constitutive equations, J. Geophys. Res., 84, 2161-2168, 1979.
- Dieterich, J. H., Constitutive properties of faults with simulated gouge, in The Mechanical Behavior of Crustal Rocks: The Handin Volume, Am. Geophys. Union Monogr. 24, 103-120, 1981.
- Dieterich, J. H., A model for the nucleation of earthquake slip, in Earthquake Source Mechanics, Am. Geophys. Union Monogr. 37, 37-47, 1986.
- Dieterich, J. H. and G. Conrad, Effect of humidity on time- and velocity-dependent friction in rocks, in Chemical Effects of Water on the Strength and Deformation of Crustal Rocks, J. Geophys. Res., 89, 4196-4202, 1984.
- Dula, W. F., Jr., High Temperature Deformation of wet and dry artificial quartz gouge, Ph.D. Dissertation, Texas A&M University, 382 pp., 1985.
- Dunning, J. D. and M. E. Miller, Effects of pore fluid chemistry on stable sliding of Berea sandstone, Unpublished manuscript, 1986.
- Dunning, J. D., D. Petrovski, J. Schuyler, and A. Owens, The effects of aqueous chemical environments on crack propagation in quartz, J. Geophys. Res., 89, 4115-4123, 1984.
- Dunning, J. D., W. C. Lewis, and D. E. Dunn, Chemomechanical weakening in the presence of surfactants, J. Geophys. Res., 85, 5344-5354, 1980.
- Engelder, J. T., Cataclasis and the generation of fault gouge, Geol. Soc. Am. Bull., 85, 1515-1522, 1974.
- Engelder, J. T., J. M. Logan, and J. W. Handin, The sliding characteristics of sandstone on quartz fault-gouge, Pure Appl. Geophys., 113, 60-86, 1975.
- Evans, B., The effect of temperature and impurity content on indentation hardness of quartz, J. Geophys. Res., 89, 4213-4222, 1984.
- Evans, B. and C. Goetze, The temperature variation of hardness of olivine and its implication for polycrystalline yield stress, J. Geophys. Res., 84, 5505-5524, 1979.

- Feucht, L. J., The Effects of Chemical environment on the frictional properties of a quartzose sandstone, M.S. Thesis, Texas A&M University, 52 pp., 1985.
- Field, J. E., Brittle fracture: its study and application, Comtemp. Phys., 12, 1-31, 1971.
- Flinn, D., Transcurrent faults and associated cataclasis in Shetland, J. Geol. Soc. London, 133, 231-248, 1977.
- Friedel, J., Dislocations, Pergamon Press, Reading, Mass., 1964.
- Friedman, M., Fracture in rock, Rev. Geophys., 13, 352-358, 1975.
- Friedman, M., J. Handin and G. Alani, Fracture surface energy of rocks, Int. J. Rock Mech. Min. Sci., 9, 757-766, 1972.
- Friedman, M., J. M. Logan, and J. A. Rigert, Glass-indurated quartz gouge in sliding-friction experiments on sandstone, Geol. Soc. Am. Bull., 85, 937-942, 1974.
- Friedman, M. and B. Johnson, Thermal cracks in unconfined Sioux quartzite, Proc. 19th U.S. Symp. Rock Mech, Univ. Nevada, Reno, 423-430, 1978.
- Friedman, M., J. Handin, N. G. Higgs, and J. R. Lantz, Strength and ductility of four dry igneous rocks at low pressures and temperatures to partial melting, Proc. 20th U.S. Symp. Rock Mech., Univ. Texas, Austin, 35-49, 1979.
- Friedman, M., J. Handin, N. G. Higgs, J. R. Lantz, and S. J. Bauer, Strength and ductility of room-dry and water-saturated igneous rocks at low pressures and temperatures to partial melting, Sandia National Laboratories, Report SAND80-7159, 87 pp., 1980.
- Friedman, M. and N. G. Higgs, Calcite fabrics in experimental shear zones, in The Mechanical Behavior of Crustal Rocks: The Handin Volume, Am. Geophys. Union Monogr. 24, 11-28, 1981.
- Friedman, M., J. Handin, and S. J. Bauer, Deformation mechanisms in granodiorite at effective pressures to 100 MPa and temperatures to partial melting, Proc. 23rd U.S. Symp. Rock Mech., Univ. California, Berkeley, 279-289, 1982.
- Friedman, M., J. Handin, J. E. Russell, A. K. Kronenberg, T. W. Hopkins and F. M. Chester, Mechanical properties of rocks at high temperatures and pressures, Summary Report 1, Dept. of Energy, Contract DE-A505-79ER10361, 1986.
- Goodman, R. E. and P. N. Sundaram, Fault and system stiffness and stick-slip phenomena, Pure Appl. Geophys., 116, 873-887, 1978.
- Griggs, D., F. J. Turner and H. C. Heard, Deformation of rocks at 500°-800°C, in Rock Deformation, Geol. Soc. Am. Mem., 79, 39-104, 1960.

- Gu, J.-C., J. R. Rice, A. L. Ruina, and S. T. Tse, Slip motion and stability of a single degree of freedom elastic system with rate and state dependent friction, J. Mech. Phys. Solids, 32, 167-196, 1984.
- Handin, J., On the Coulomb-Mohr failure criterion, J. Geophys. Res., 74, 5343-5348, 1969.
- Handin, J., H. C. Heard, and J. N. Magouirk, The effect of the intermediate principal stress on the failure of limestone, dolomite, and glass at different temperatures and strain rates, J. Geophys. Res., 72, 611-640, 1967.
- Handin, J., M. Friedman, and S. J. Bauer, Borehole stability and rock-drillability studies (Abstract), Am. Geophys. Union Trans., 62, 105/, 1981.
- Handin, J. and J. M. Logan, Experimental tectonophysics, Geophys. Res. Ltrs., 8, 647-650, 1981.
- Hansen, F. D., Semibrittle creep of selected crustal rocks at 1000 MPa, Ph.D. Dissertation, 224 pp., Texas A&M University, 1982.
- Hansen, F. D. and N. L. Carter, Semibrittle creep of dry and wet Westerly Granite at 1000 MPa. Proc. 24th U.S. Symp. Rock Mechanics, Texas A&M U., 429-447, 1983.
- Hansen, K. S. and W. R. Purcell, Earth stress measurements in the South Belridge Oil Field, Soc. Pet. Eng., SPE15641, (61st Annual Technical Conference, New Orleans), 1-10, 1986.
- Hendersen, R., (project engineer), SEM/TEM Fractography Handbook, Metals and Ceramics Information Center, Battelle Mem. Inst., Columbus, Ohio, 1975.
- Herget, G., Stress assumptions for underground excavations in the Canadian Shield, Int. J. Rock Mech, Min. Sci., 24, 95-97, 1987.
- Heuze, F. E., Scale effects in the determination of rock mass strength, Rock Mech., 12, 167-192, 1980.
- Higgs, N. G., Mechanical Properties of Ultrafine Quartz, Chlorite, and Bentonite in Environments Appropriate to Upper-Crustal Earthquakes, Ph.D. Dissertation, Texas A&M University, 267 pp., 1981.
- Hoagland, R. G., G. T. Hahn and A. R. Rosenfield, Influence of microstructure on fracture propagation in rock, Rock Mech, 5, 77-106, 1973.
- Hock, E. and E. T. Brown, Empirical strength criterion for rock masses. J. Geotech. Eng. Div., Am. Soc. Civil Engineers, 106, GT9, 1013-1035, 1980.
- Hockey, B. J. and R. R. Lawn, Electron microscopy of microcracking about indentations in aluminum oxide and silicon carbide, J. Mater. Sci., 10, 1275-1284, 1975.

- Hopkins, T. W., Microfracturing in Westerly Granite Experimentally Extended Wet and Dry at Temperatures to 800°C and Pressures to 200 MPa, M.S. Thesis, Texas A&M University, 71 pp., 1986.
- Horn, H. M. and D. U. Deere, Frictional characteristics of minerals, Geotechnique, 12, 319-335, 1962.
- Hungr, O. and N. R. Morgenstern, High velocity ring shear tests in sand, Geotechnique, 34, 415-421, 1984.
- Ishido, T. and O. Nishizawa, Effects of zeta potential on microcrack growth in rock under relatively low uniaxial compression, J. Geophys. Res., 89, 4153-4159, 1984.
- Jaeger, J. C. and N. G. W. Cook, Fundamentals of Rock Mechanics, Chapman and Hall, London, 3rd Edition, 593 pp., 1979.
- Jamison, D. B. and N. G. W. Cook, Note on measured values for the state of stress in the earth's crust, J. Geophys. Res., 85, 1833-1838, 1980.
- Johnson, B., A. F. Gangi and J. Handin, Thermal cracking of rocks subjected to slow, uniform temperature changes, Proc. 29th U.S. Symp. Rock Mech., Univ. Nevada, Reno, 259-267, 1978.
- Johnson, B., M. Friedman, T. W. Hopkins, and S. J. Bauer, Strength and microfracturing of Westerly granite extended wet and dry at temperatures to 800°C and pressures to 200 MPa, Proc. 28th U.S. Symp. Rock Mech., Univ. Arizona, Tucson, 1987.
- Johnson, T., Time dependent friction of granite: implications for precursory slip on faults, J. Geophys. Res., 86, 6017-6028, 1981.
- Kern, H., The effect of high temperature and confining pressure on compressional wave velocities in quartz-bearing and quartz free igneous and metamorphic rocks, Tectonophysics, 44, 185-203, 1978.
- Kerr, R. A., Is the San Andreas weak at heart?, Science, 236, 388-389, 1987.
- Kragelskii, I. V., Friction and Wear, Butterworths, Washington, D.C., 346 pp., 1965.
- Kranz, R. L., Microcracks in rocks--A review, Tectonophysics, 100, 449-480, 1983.
- Kranz, R. L., W. J. Harris, and N. L. Carter, Static fatigue of granite at 200°C, Geophys. Res. Ltrs., 9, 1-4, 1982.
- Krech, W. W., F. A. Henderson and K. E. Hjelmstad, A standard rock suite for rapid excavation research, U.S. Bureau of Mines, R.I., 7865, 29, pp. 1974.
- Kronenberg, A. K., J. C. Pinkston, and S. H. Kirby, Basal slip of biotite single crystals, Trans. Am. Geophys. Union, 66, 366, 1985a.

- Kronenberg, A. K., S. H. Kirby, and J. C. Pinkston, Compression of biotite single crystals perpendicular to cleavage, Trans. Am. Geophys. Union, 66, 1085, 1985b.
- Lawn, B. R. and T. R. Wilshaw, Fracture of Brittle Solids, Cambridge Univ. Press, Cambridge, UK, 1975.
- Lockner, D. A. and J. D. Byerlee, Laboratory measurements of velocity dependent frictional strength, U.S. Geological Survey, Open File Rept., 86-417, 1986.
- Lockner, D. A., R. Summers, and J. D. Byerlee, Effects of temperature and sliding rate on frictional strength of granite, Pure Appl. Geophys., 124, 1986.
- Logan, J. M., Friction in rocks, Rev. Geophys., 13, 358-361, 1975.
- Logan, J. M., Creep, Stable sliding and premonitory slip, Pure Appl. Geophys., 116, 773-789, 1978.
- Logan, J. M., Brittle phenomenon, Rev. Geophys., 17, 1121-1132, 1979.
- Logan, J. M. and M. L. Blackwell, The influence of chemically-active fluids on the frictional behavior of sandstone (abstract), Trans. Am. Geophys. Union, 64, 835, 1983.
- Logan, J. M. and K. A. Rauenzahn, Velocity-dependent behavior of mixed quartz-montmorillonite gouge, unpublished manuscript, 1986.
- Logan, J. M. and L. W. Teufel, Comparison of contact areas and temperatures measured during frictional sliding of a sandstone and limestone; Part 2: Effect of normal stress on the real area of contact (Abstract), Geol. Soc. Am. Abstracts with Programs, 8, 982, 1976.
- Logan, J. M. and L. W. Teufel, The effect of normal stress on the real area of contact during frictional sliding in rocks, Pure Appl. Geophys., 124, 4/1-486, 1986.
- Logan, J. M., M. Friedman, N. G. Higgs, C. Dengo, and T. Shimamoto, Experimental studies of simulated gouge and their application to studies of natural fault gouge, in Proc. Conf. VIII, Analysis of Actual Fault Zones in Bedrock, U.S. Geological Survey, Open-file Rept. 79-1239, 305-343, 1979.
- Logan, J. M., N. G. Higgs, and M. Friedman, Laboratory studies on natural fault gouge from the U.S. Geological Survey Dry Lake Valley No. 1 Well, San Andreas fault zone, in Mechanical Behavior of Crustal Rocks: The Handin Volume, Am. Geophys. Monogr. 24, 121-134, 1981.
- Logan, J. M., T. Iwasaki, M. Friedman, and S. A. Kling, Experimental investigation of sliding friction in multilithologic specimens, in Geological Factors in Rapid Excavation, Geol. Soc. Am. Engr. Geol. Case History, 9, 55-67, 1972.

- Martin, R. J., III and W. B. Durham, Mechanisms of crack growth in quartz, J. Geophys. Res., 80, 4837-4844, 1975.
- McColm, I. J., Ceramic Science for Material Technologists, Leonard Hill, Glasgow, UK, 1983.
- McGarr, A., Some constraints on levels of shear stress in the crust from observations and theory, J. Geophys. Res., 85, 6231-6238.
- Mecholsky, J. J., S. W. Frieman and R. W. Rice, Fractographic analysis of ceramics, in Fractography in Failure Analysis, edited by B. M. Straus and W. H. Cullen, ASTM STP 645, 363-379, 1978.
- Mogi, K., Effect of the intermediate principal stress on rock failure, J. Geophys. Res., 72, 5117-5131, 1967.
- Moore, D. E., R. Summers, and J. D. Byerlee, Strengths of clay and non-clay fault gouges at elevated temperatures and pressures, Proc. 24th U.S. Symp. Rock Mech., Texas A&M Univer., College Station, 489-500, 1983.
- Moore, D. E., R. Summers, and J. D. Byerlee, The effects of sliding velocity on the frictional and physical properties of heated fault gouge, Pure Appl. Geophys., 124, 31-52, 1986.
- Morrow, C. A., L. Q. Shi, and J. Byerlee, Permeability and strength of San Andreas fault gouge under high pressure, Geophys. Res. Ltrs, 8, 325-328, 1981.
- Morrow, C. A., L. Q. Shi, and J. D. Byerlee, 1982. Strain hardening and strength of clay-rich fault gouges, J. Geophys. Res., 87, 6771-6780, 1982.
- Murase, T. and A. F. McBirney, Properties of some common igneous rocks and their melts at high temperatures. Geol. Soc. Am. Bull., 84, 3563-3592, 1973.
- Norton, M. G. and B. K. Atkinson, Stress-dependent morphological features on fracture surfaces of quartz and glass, Tectonophysics, 77, 283-295, 1981.
- Okubo, P. G. and J. H. Dieterich, State variable fault constitutive relations for dynamic slip, in Earthquake Source Mechanics, Am. Geophys. Union Monogr., 37 25-35, 1986.
- Paillet, F. L. and K. Kim, Character and distribution of borehole breakouts and their relationship to in-situ stresses in deep Columbia River basalts, J. Geophys. Res., 92, 6223-6234, 1987.
- Paterson, M. S., Experimental Rock Deformation--The Brittle Field, Springer-Verlag, Berlin, 1978,
- Peck, L. and R. B. Gordon, The effect of compressive stress on the fracture energy of Sioux quartzite, J. Geophys. Res., 9, 186-189, 1982.

- Peterson, R. E., Stress Concentration Factors, Wiley, New York, 1974.
- Poirier, J. P., Creep of Crystals. High Temperature deformation processes in metals, ceramics and minerals, Cambridge University Press, Cambridge, UK, 260 pp., 1985.
- Pratt, H. R., A. D. Black, and W. F. Brace, Friction and deformation of jointed quartz diorite, Proc. 3rd Cong. Intern. Soc. Rock Mech., IIA, 306-310, 1974a.
- Pratt, H. R., A. D. Black, W. S. Brown, and W. F. Brace, A new technique for determining the deformation and frictional characteristics of in situ rock, in Field Testing and Instrumentation of Rock, Spec. Tech. Pub. 554, Am. Soc. Testing Materials, 3-19, 1974b.
- Rabinowicz, E., Friction and Wear of Materials, Wiley, New York. 244 pp., 1965.
- Ratigan, J. L., A statistical fracture mechanics approach to the strength of brittle rock, Ph.D. Dissertation, 92 pp., U. California, Berkeley, 1981.
- Rice, J. R., Constitutive relation for fault slip and earthquake instabilities, Pure Appl. Geophys., 121, 443-475, 1983.
- Rice, R. W., Ceramic fracture features, observations, mechanisms and uses, in Fractography of Ceramic and Metal Failures, ASTM STp 827, 5-101, 1984.
- Rice, J. R. and J-C. Gu, Earthquake after effects and triggered seismic phenomena, Pure Appl. Geophys., 121, 187-219, 1983.
- Rice, J. R. and A. L. Ruina, Stability of steady frictional slipping, J. Appl. Mech., 50, 343-349, 1983.
- Rice, R. W., The Compressive Strength of Ceramics, 5, Plenum, New York, 195-229, 1971.
- Rosengren, K. J. and J. C. Jaeger, The mechanical properties of an interlocked low-porosity aggregate, Geotechnique, 18, 317-326, 1968.
- Ross, J. V., S. J. Bauer, and N. L. Carter, Effect of the α β quartz transition on the creep properties of quartzite and granite, Geophys. Res. Ltrs., 10, 1129-1132, 1983.
- Ruina, A. L., Friction Laws and Instabilities: A Quasistatic Analysis of Some Dry Frictional Behavior, Ph.D. Dissertation, Brown Univ., 99 pp., 1980.
- Ruina, A. L., Slip instability and state variable friction laws, J. Geophys. Res., 88, 10359-10370, 1983.
- Rutter, E. H., Pressure solution in nature, theory and experiment, J. Geol. Soc. London, 140, 725-740, 1983.

- Rutter, E. H., R. H. Maddock, S. H. Hall, and S. H. White, Comparative microstructure of natural and experimentally produced clay bearing fault gouge, Pure Appl. Geophys., 124, 3-30, 1986.
- Rutter, E. H. and D. H. Mainprice, The effect of water on stress relaxation of faulted and unfaulted sandstone, Pure Appl. Geophys., 116, 634-654, 1978.
- Rutter, E. H. and D. H. Mainprice, On the possibility of slow fault slip controlled by a diffusive mass transfer process, Gerlands Beitr. Geophysik, 88, 154-162, 1979.
- Rutter, E. H. and S. H. White, The microstructures and rheology of fault gouges produced experimentally under wet and dry conditions at temperatures up to 400°C, Bull. Mineral., 102, 101-109, 1979.
- Ryan, M. P. and C. G. Sammis, Cyclic fracture mechanisms in cooling basalt, Geol. Soc. Am. Bull., 89, 1295-1308, 1978.
- Schneider, H. J., The friction and deformation behavior of rock joints, Rock Mech., 8, 169-184, 1976.
- Schock, R. N. and H. Louis, Strain behavior of a granite and a graywacke sandstone in tension, J. Geophys. Res., 87, 7817-7823, 1982.
- Scholz, C. H. and J. T. Engelder, The role of asperity indentation and ploughing in rock friction, I, asperity creep and stick-slip. Int. J. Rock Mech. Min. Sci., 13, 149-154, 1976.
- Scholz, C. H. and C. A. Aviles, The fractal geometry of faults and faulting, in Earthquake Source Mechanics, Am. Geophys. Union, Monogr. 37, 147-156, 1986.
- Segall, P. and D. D. Pollard, Joint formation in granitic rock of the Sierra Nevada, Geol. Soc. Am. Bull., 94, 563-575, 1983.
- Shand, E. B., Experimental study of fracture of glass: II. Experimental data, Am. Ceram. Soc. J., 37, 559-573, 1954.
- Shimamoto, T., Confining pressure reduction experiments: a new method for measuring frictional strength over a wide range of normal stress. Int. J. Rock Mech. Min. Sci., 22, 227-236, 1985a.
- Shimamoto, T., The origin of large great thrust-type earthquakes along subducting plate boundaries, Tectonophysics, 119, 37-65, 1985b.
- Shimamoto, T., Transition between frictional slip and ductile flow for halite shear zones at room temperature, Science, 231, 711-714, 1986.
- Shimamoto, T. and J. M. Logan, Effects of simulated fault gouge on the sliding behavior of Tennessee Sandstone, J. Geophys. Res., 86, 2902-2914, 1981a.

- Shimamoto, T. and J. M. Logan, Effects of simulated clay gouges on the sliding behavior of Tennessee sandstone, Tectonophysics, 75, 243-255, 1981b.
- Shimamoto, T. and J. M. Logan, Laboratory friction experiments and natural earthquakes: an argument for long-term tests, Tectonophysics, 109, 165-175, 1984.
- Shimamoto, T. and J. M. Logan, Velocity-dependent behavior in simulated halite shear zones: an analog for silicates, in Earthquake Source Mechanics, Am. Geophys. Union Monogr. 37, 49-63, 1986.
- Simmons, G. and D. Richter, Microcracks in rocks, in The Physics and Chemistry of Minerals and Rocks, edited by R. G. J. Strens, New York, Wiley, 1976.
- Solberg, P. and J. D. Byerlee, A note on the rate sensitivity of frictional sliding of Westerly Granite, J. Geophys. Res., 89, 4203-4206, 1984.
- Sosman, R. B., The Phases of Silica, 388 pp., Rutgers University Press, New Brunswick, N.Jersey, 1965.
- Straus, B. M. and W. H. Cullen, Fractography in Failure Analysis, ASTM STP 645, 1978.
- Stesky, R. M., Mechanisms of high temperature frictional sliding in Westerly Granite, Can. J. Earth Sci., 15, 361-375, 1978a.
- Stesky, R. M., Rock friction--effect of confining pressure, temperature and pore pressure, Pure Appl. Geophys., 116, 690-704, 1978b.
- Stesky, R. M., W. F. Brace, D. K. Riley, and P. Y. F. Robin, Friction in faulted rock at high temperature and pressure, Tectonophysics, 23, 177-203, 1974.
- Stierman, D. J., Geophysical and geological evidence for fracturing, water circulation and chemical alteration in granitic rocks adjacent to major strike-slip faults, J. Geophys. Res., 89, 5849-5857, 1984.
- Summers, R. and J. Byerlee, A note on the effect of fault gouge composition on the stability of frictional sliding, Int. J. Rock Mech. Min. Sci., 14, 155-160, 1977.
- Tabor, D., The hardness of solids, Rev. Phys. Technol., 1, 145-179, 1970.
- Tabor, D., The Hardness of Metals, Clarendon Press, Oxford, U.K., 1951.
- Tapponnier, P. and W. F. Brace, Development of stress-induced microcracks in Westerly granite, Int. J. Rock Mech. Min. Sci., 13, 103-112, 1976.
- Teufel, L. W. and J. M. Logan, Effect of displacement rate on the real area of contact and temperature generated during frictional sliding of Tennessee Sandstone, Pure Appl. Geophys., 116, 840-865, 1978.

- Tse, S. T. and J. R. Rice, Crustal earthquake instability in relation to the depth variation of frictional slip properties, J. Geophys. Res., 91, 9452-9472, 1986.
- Tullis, T. E. and J. D. Weeks, Constitutive behavior and stability of frictional sliding of granite, Pure Appl. Geophys., 124, 1986.
- Vesic, A. S. and G. W. Clough, Behavior of granular materials under high stresses, J. Soil Mech., 94, 661-688, 1968.
- Wallace, R. E. and H. I. Morris, Characteristics of faults and shear zones as seen in mines at depths as much as 2.5 km below the surface, in Proc. Conf. VIII, Analysis of actual fault zones in bedrock, U.S. Geological Survey Open File Rep. 79-1239, 79-100, 1979.
- Wang, C-Y. and N. Mao, Shearing of saturated clays in rock joints at high confining pressures, Geophys. Res. Ltrs, 6, 825-828, 1979.
- Wawersik, W. R. and W. F. Brace, Post-failure behavior of a granite and diabase, Rock Mech., 3, 61-85, 1971.
- Weeks, J. D. and T. E. Tullis, Frictional sliding of dolomite: a variation in constitutive behavior, J. Geophys. Res., 90, 7821-7826, 1985.
- Westwood, A., Control and application of environment sensitive fracture processes, J. Mater. Sci., 9, 1871-1895, 1974.
- Wu, F. T., L. Blatter, and H. Roberson, Clay gouges in the San Andreas fault system and their possible implications, Pure Appl. Geophys., 113, 87-95, 1975.
- Wu, F. T., Mineralogy and physical nature of clay gouge, Pure Appl. Geophys., 116, 655-689, 1978.
- Zoback, M. D., D. Moos, L. Mastin, and R. N. Anderson, Well Bore breakouts and in-situ stress, J. Geophys. Res., 90, 5523-5530, 1985.