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Strength and Ductility of Room-Dry and Water-Saturated Igneous Rocks at Low Pressures and Temperatures to Partial Melting

Final Report

MASTER

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AT LOW PRESSURES AND TEMPERATURES TO PARTIAL MELTING

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ABSTRACT

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 are determined at temperatures to partial melting ($> 1050^{\circ}\text{C}$) 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 of 50 MPa), and (c) room-dry Newberry Rhyolite Obsidian at 0 and 50-MPa confining pressure. The strengths of the dry crystalline rocks are a nearly linear function of mean stress, and they decrease nearly linearly with increasing temperatures from 25°C to $800\text{--}900^{\circ}\text{C}$. At higher temperatures the strengths decrease rapidly and vanish upon partial melting when the deformation becomes quasiviscous. The obsidian is stronger than the crystalline rocks at 600°C where glass softening begins and its strength goes to zero at about 800°C . Water-weakening is evident only in the andesite. Its strength degrades dramatically between 850° and 880°C , at which temperature its ultimate strength is between 5 and 25 MPa compared to an average dry strength of 75 MPa at 1000°C . All rocks are brittle throughout the entire temperature range until melting or softening occurs. Shortenings at failure (ultimate strengths) are 3 percent or less in dry tests and ≤ 1 percent in most wet ones.

Time-dependent flow characteristics are 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 3×10^{10} N·s/m² 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 s⁻¹), $Q = 380$ 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 crystalline rocks 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 specimens, however, probably is due to an inherent temperature effect on the load-induced fracturing process. Each wet-specimen also fails along multiple macroscopic 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 solid surfaces.

In that the crystalline rocks remain brittle up to the temperatures

of partial melting, they are very likely to remain drillable by conventional means at the highest temperatures to be encountered in the roof-regions of buried magma chambers. Moreover, the penetration rate ought to increase as the magma body is approached because rock strength decreases with increasing temperature.

Instantaneous rock strength governs the stability of the borehole walls upon or shortly after drilling. Failure in compression would occur if stress differences between borehole pressure and the stresses in the walls of the borehole exceed the failure strength of the rock at a given mean pressure. Potential borehole failure at this stage could lead to stuck drill bits and possible loss of the hole. In general the maximum depth for stable boreholes increases with decreasing temperature, decreasing σ_h/σ_v ratio, and for dry rocks with filled versus open boreholes. Furthermore, other factors being equal, boreholes in the granodiorite should remain stable to greater depths than those in basalt or andesite, with the andesite providing the least stable ones. In the most optimistic case, smooth, even unfilled boreholes in dry, impervious, fresh crystalline rock are apt to be stable over short time periods to depths > 10 km provided temperatures do not exceed 700°C and the σ_h/σ_v ratio is 0.5. Even the water-saturated andesite, which is water-weakened, would provide stable boreholes to > 10 km and temperatures to 880°C if the effective pressure was zero and σ_h/σ_v is 1.0. On the other hand, unfilled boreholes in the andesite are apt to fail at only 1 to 2 km depth even at 25°, if σ_h/σ_v is 2.0.

INTRODUCTION

Borehole stability is one of many severe engineering problems to be anticipated should attempts ever be made to recover directly the geothermal energy stored in magma chambers, as temperatures of the order of 1000°C are expectable. How strong are the rocks to be drilled in this hostile environment? Previous laboratory work is not very helpful because virtually all testing above 500°C has been done at either atmospheric pressure (Murrell and Chakravarty, 1973) or very high confining pressures of 400 MPa or more (Handin, 1966; Carter, 1976; Carter and Kirby, 1978). What is needed is testing at high temperature but low confining pressure. Thus, under laboratory conditions that simulate the expectable natural environment of engineering concern, we are testing rock types that are likely candidates for drilling. Reported herein are the short-time ultimate strengths and ductilities determined at temperatures of 25° to 1050°C and a strain rate of 10^{-4}s^{-1} of (a) room-dry Mt. Hood Andesite, Cuerbio Basalt, and Charcoal (St. Cloud Gray) Granodiorite at confining pressures of 0, 50, and 100 MPa, (b) water-saturated specimens of the same three rocks at zero effective pressure (both pore and confining pressures of 50 MPa), and (c) room-dry Newberry Rhyolite Obsidian at 0 and 50 MPa. These strengths are then compared with the stresses developed at the wall of a borehole in an elastic medium at the appropriate temperatures and mean pressures to assess the problem of borehole stability.

APPARATUS AND PROCEDURES

Some of the triaxial compression tests have been done in a screw-driven machine, similar to that designed by Heard (1963), at strain rates of 10^{-4} and 10^{-7} /s on 0.9 by 2.0-cm specimens. Others, including all at temperatures above 400°C, have been carried out in the new apparatus (Figure 1) briefly described below.

Cylindrical specimens, 2 cm in diameter and 4 cm in length, are axially loaded in a simple four-pole hydraulic press, driven by a variable-speed, constant-displacement, duplex oil pump so as to shorten specimens at constant axial strain rates of 10^{-4} or 10^{-6} /s. Force (up to 1 MN) is measured by a Baldwin CXX load-cell hooked to an X-Y recorder. Displacement is measured by a DCDT hooked to the same recorder.

The confining pressure of argon up to 300 MPa is generated by an intensifier driven by an air pump and charged by an oil-gas separator driven by a second air pump. It is measured by a precision Heise gauge and regulated by a Brown strip-chart recorder.

The triaxial cell, similar in design to that of Goetze (1971), with an inside diameter of 3.8 cm and an outside diameter of 16.5 cm, is made of 4340 steel, drawn to a hardness of Rockwell 34C for optimum toughness. Inside the cell the specimen is jacketed in a thin-walled copper sleeve and surrounded by the internal furnace. This furnace consists of two independently powered coils of Kanthal wire separated by a distance equal to the length of the specimen, and embedded in Norton EA 162 alumina and Sauereisen No. 78 electric-resistor cement prior to wrapping with Fiberfrax, refractory, insulating paper to the maximum diameter. The absence of furnace windings in the central region serves to flatten previously

Figure 1. High-temperature triaxial cell.

convex temperature profiles, related to the low conductivity of the sample. By adjusting the power distribution between the upper and lower windings, severe thermal gradients caused by convection in the gap between the furnace and the specimen assembly are subdued. For unconfined tests, nearly flat temperature profiles are achieved by dissipating only 40% of the total power to the upper furnace coil. For confined tests, convection in the dense argon confining-medium has been a bigger problem, but this has been overcome by infilling the annulus between the furnace wall and specimen with boron-nitride powder, an inert, non-abrasive material, combining excellent dielectric characteristics with good thermal conductivity, and effectively eliminating convection. Temperature is measured on the specimen by a thermocouple inserted through a hole in the lower plug and regulated by a West stepless controller. Furnace power is fed through Bridgman-type, insulated cones. Temperature differences along the length of the specimen are held to as small as 30°C at 1000°C, and temperature measurements are accurate to $\pm 5^\circ\text{C}$ (Figure 2).

The cell is cooled externally by a water jacket, and loss of heat through the upper loading piston and lower plug is reduced by lucalox spacers, which are excellent thermal insulators, but which also have adequate compressive strengths at the temperatures of interest. Seals on the moving piston and lower closure can then be made by ordinary rubber O-rings.

Each dry test is done at a constant confining pressure of 0, 50, or 100 MPa, a temperature between 25° and 1050°C, and a strain rate of 10^{-4} s^{-1} . True stress-strain curves are derived from the force-displacement record. Axial differential stress at each increment of axial strain is calculated by dividing the recorded differential force by the area of the specimen, corrected by assuming Poisson's ratio of 0.25 in the elastic range and

Figure 2. Schematic diagram (a) shows modified furnace design with accompanying axial temperature profiles for an unconfined test with an air gap between furnace and specimen assembly. Boron nitride powder fills the gap when test is to be made at confining pressure. In (b) a circuit diagram illustrates how total power is distributed between upper and lower furnace coils.

volume conservation and homogeneous deformation in the inelastic range. Conventional strain is calculated by dividing the change in length by the original length, the displacement record having been corrected for elastic distortion of the apparatus, and expressed as percent shortening. The accuracies of stress, strain, and confining pressure are 2 to 3 percent.

In the triaxial compression test, the principal stresses are $\sigma_1 > \sigma_2 = \sigma_3 = p_c$, where p_c is confining pressure and compressive stresses are counted positive. Later on, we shall use the laboratory data to predict failure of the borehole where $\sigma_1 > \sigma_2 > \sigma_3$, and we shall assume that ultimate compressive strength ($\sigma_1 - \sigma_3$), at failure, is a function only of mean stress, $\sigma_m = (\sigma_1 + \sigma_2 + \sigma_3)/3$, and temperature. Thus, for each test we compute ($\sigma_1 - \sigma_3$) and σ_m and list them together with the total shortening at macroscopic failure that serves as a measure of ductility (Tables 1-5). Also listed is the angle θ between the load axis (σ_1) and the macroscopic fault, as measured with a contact goniometer outside the copper jacket. "Shattered" specimens have failed by multiple faulting, and θ cannot be determined.

For the water-saturated tests, each specimen is jacketed with excess water in thin-walled copper tubing, and is placed within the internal furnace in the pressure cell. The confining pressure and pore pressure are elevated, e.g., to 50 and 45 MPa, respectively. The 5 MPa difference helps to drive water into pores and cracks and helps to keep it out of the interface between rock and jacket. Temperature is then elevated to test conditions at about $10^\circ\text{C min}^{-1}$ and the pore pressure is adjusted to 50 MPa. Use of spacers and pistons with a central borehole make these

drained tests. If during an experiment the pore space is fully saturated and the pore pressure is equilibrated throughout the system, then the fluid pressure in the specimen is truly registered by the external gauge; and when the internal pore pressure, p_p , equals the external confining pressure, p_c , the effective confining pressure, p_e , is zero. If the effect of the fluid pressure is purely mechanical, the mechanical properties should then be the same as those measured for the dry rock at $p_c = 0$. If the rock is not fully saturated and/or the permeability is very low, the true value of the pore pressure may be unknown, and the effective confining pressure may not be zero. Pore water is pumped into a specimen through a hole in the top piston; thus visual evidence of water at the lower end, at least confirms pervasion.

STARTING MATERIALS

Mount Hood Andesite

Our sample was collected from a Recent flow, located just below the parking lot of Timberline Lodge on the south slope of Mount Hood. The block is pink and porphyritic with an aphanitic groundmass. Porosity varies from 10 to 12%. Phenocrysts comprise about 40% of the rock, 30% being plagioclase, 5% pyroxene, and about 1% each of hornblende, olivine, and magnetite. The feldspars are zoned, twinned labradorite, and along with the other phenocrysts, are conspicuously fractured. The long axes of the phenocrysts range up to 5 mm; overall they are randomly oriented, although in small domains they often show parallel alignment. Microfractures are confined to the phenocrysts, and none was observed to transect both phenocrysts and groundmass. The microfractures are randomly oriented, and the average number of microfractures per mm is less than 1 in the groundmass and about 7 in the phenocrysts.

The groundmass consists of small crystals (<1 mm) of lath-shaped feldspar, dispersed magnetite, and a small amount of interstitial glass. It completely envelopes the phenocrysts. This nearly holocrystalline groundmass should serve as the load-bearing framework that controls the gross mechanical behavior of the rock. A chemical analysis is given by Murase and McBirney (1973) who also list liquidus temperatures of 1100°C for pyroxene, 1240°C for plagioclase, and 1210°C for the opaque oxides.

Charcoal Granodiorite

This rock is properly known as the St. Cloud Gray Granodiorite of

Precambrian age from St. Cloud, Minnesota; "Charcoal" is its trade name. Our sample was purchased from the Cold Springs Granite Co. in Minnesota. It is one of the rocks adopted as standard by the U.S. Bureau of Mines (Krech et al., 1974). Its modal composition is: 38% labradorite, 26% orthoclase, 21% quartz, 12% hornblende and 5% biotite. Grain size ranges from < 0.1 to 10.0 mm and averages about 2 mm. Quartz crystals are free of deformation lamellae; but 49% exhibit undulose extinction, 43% are polygonized, and 12% are undeformed. Some 67% contain healed fractures, and 21% contain fresh fractures. The feldspars are all twinned and nearly all contain some fractures. Overall, however, the average number of microfractures per mm is less than 1.

There is a fracture anisotropy in this rock. Experimentally induced tensile fractures are controlled by pre-existing arrays of microfractures; ultrasonic velocity is minimum, and attenuation is maximum perpendicular to the plane of fracture anisotropy (Friedman and Bur, 1974). All our test specimens are loaded perpendicular to this plane. Porosity is less than 0.1%.

Cuerbio Basalt

A large block of the Cuerbio Basalt of Early Pleistocene age was collected from the lower of two flows exposed in a roadcut along U.S. Highway 85, near the top of La Bajada Hill, south of La Cienega, New Mexico. The specimen contains some macroscopic, elongated vesicles up to 2 cm long. Sawed slabs from the block were X-rayed at Sandia Laboratories by Dr. E. J. Graeber and his associate, so that we could core our specimens from domains free of the larger vesicles. This approach worked successfully, and the few small vesicles (< 1.0 mm) exposed at the surface

of some cylinders were filled with a high-temperature cement to prevent jacket failure. The reproducibility and consistent trends of the experimental data attest to the success of this approach. Porosity of test specimens ranges between 5 and 8%.

The basalt is dark gray and it contains an aphanitic groundmass, phenocrysts, and vesicles, both with maximum lengths of about 1 mm. Thin-section analyses reveal that the larger phenocrysts are composed of 80% euhedral, fractured olivine, 17% cleaved pyroxene, and 4% twinned and fractured feldspar; in aggregate these phenocrysts compose 14% of the rock. The groundmass (86%) is holocrystalline and consists of microphenocrysts of olivine and pyroxene (< 0.1 mm), magnetite (< 0.06 mm), and lath-shaped feldspars (up to 0.35 mm). These comprise 10%, 10%, and 80% of the groundmass, respectively. Microscopically, the rock shows a flow layering manifest by the dimensional alignment of feldspar laths in the groundmass and the larger olivine phenocrysts. All our specimens are loaded parallel to this planar anisotropy.

Newberry Rhyolite Obsidian

Our sample was collected from the north end of the Big Obsidian Flow along the trail southeast of Paulina Lake, Newberry Caldera, Oregon. The block contains flow banding which locally produces a planar anisotropy and complicates the coring of perfect, right-circular cylinders for experiments. Microscopically, this rock consists entirely of a clear, black glass which completely envelope minute crystals of feldspar. The crystallites are strongly oriented dimensionally and reflect local, primary flow of lava prior to freezing. This fabric cannot be related to the experimental deformation; accordingly, petrofabric studies are macroscopic.

A specimen deforms by either brittle fracture or by uniform flow with no macroscopic manifestation except barrelling. Data on chemical composition, viscosity, and other physical properties are given by Murase and McBirney (1973).

EXPERIMENTAL RESULTS

Crystalline Rocks

Dry Experiments. Data sets for the instantaneous failure strength and strain of room-dry Mt. Hood Andesite, Charcoal Granodiorite, and Cuerbio Basalt, shortened at confining pressures of 0, 50, and 100 MPa, temperatures to 1000°C, and a strain rate of 10^{-4} /s, are given in Tables 1, 2, and 3, and corresponding stress-strain curves appear in Figures 3, 4, and 5. These data reveal the following significant facts.

(1) The reproducibility of data on strength and shortening at failure is reasonably good for specimens deformed under similar conditions and for specimens of different size deformed under the same conditions, but in a different apparatus (e.g., Table 2, specimens T1 and M3).

(2) The failure strengths of all three rocks decrease with increasing temperature at constant confining pressure and increase with increasing confining pressure at constant temperature (Figures 6, 7, and 8). The average strength-reduction of the andesite is essentially linear from 125 MPa at 25°C to 78 MPa at 1000° unconfined, and from 290 MPa at 25° to 200 MPa at 900°C, at 50-MPa confining pressure, or 1.2 and 3.3 MPa/25°C. This strength reduction is more dramatic at 100 MPa-confining pressure, i.e., 14 MPa/25°C between 700 and 1000°C. Strength reductions for the granodiorite are larger than those for the andesite at each confining pressure, i.e., 11 MPa/25°C at 50 MPa and 27 MPa/25°C at 100 MPa. Unconfined, the reduction is non-linear from 350 MPa at 25°C to 35 MPa at 1000°C. The strength reduction for the basalt is also nearly linear to about 900°C; gradients are 30 MPa/25°C at 100, 8 MPa/25°C at 50-MPa confining pressure

TABLE 1. EXPERIMENTAL DATA FOR ROOM-DRY MOUNT HOOD ANDESITE

SPECIMEN NO.	CONFINING PRESSURE (MPa)	TEMP. (°C)	MEAN STRESS (MPa)	ULTIMATE STRENGTH (MPa)	SHORTENING AT FAILURE (%)	COMMENTS ^a & FAULT ANGLE (°)
20	0	25	50	150	1.1	S
21	0	25	30	100	1.1	S
481 ^b	0	25	35	110	0.8	S
482 ^b	0	25	30	90	0.8	S
483 ^b	0	25	40	115	1.3	23
493	0	400	50	145	1.3	S
496	0	400	35	110	1.2	S
499	0	400	35	105	1.2	23
503 ^b	0	400	35	110	1.0	S
507 ^b	0	400	45	135	1.1	30
36	0	700	40	115	1.0	25, S
37	0	700	35	100	0.9	S
8	50	25	155	320	2.3	27
225	50	25	135	250	3.2	S
27 ^b	50	25	125	220	6.8	34
28 ^b	50	25	125	220	6.9	32
9	50	400	125	230	2.7	32
16	50	400	150	305	1.8	32
31	50	700	130	235	0.9	I
34	50	700	130	235	1.6	I
84	0	1000	30	85	1.5	S
85	0	1000	25	75	1.5	S
97	50	1000	85	110	1.4	S
123 ^c	50	930	105	165	1.6	
123 ^c	50	900	115	195	3.5	
123 ^c	50	880	115	195	5.4	
125 ^d	50	25	180	390	2.1	32
128	50	25	140	275	2.0	32
221 ^e	0	25	30	90	1.7	S
223 ^f	0	25	40	120	1.5	S
226 ^g	50	25	155	315	2.5	29
258	100	720	230	395	2.2	30
271	100	700	255	460	3.0	S
256	100	960	210	330	3.6	S
269	100	1000	190	275	3.2	32, S

a. S = shattered (2 or more fractures or faults), I = incipient fault.

b. Strain rate is 10^{-7} s^{-1} .

c. Specimen work-hardened during temperature decrease.

d. Specimen 125 was heated to 930°C at 10°C/min, cooled, and then deformed at 25°C and 50 MPa.

e. Unconfined specimen heated to 905°C at 1.5°C/min, cooled, and then deformed at 25°C.

f. Unconfined specimen heated to 1000°C at 1.5°C/min, cooled, and then deformed at 25°C.

g. Unconfined specimen heated to 1000°C at 1.5°C/min, cooled and then deformed at 25°C and 50 MPa.

TABLE 2. EXPERIMENTAL DATA FOR ROOM-DRY CHARCOAL GRANODIORITE

SPECIMEN NO.	CONFINING PRESSURE (MPa)	TEMP (°C)	MEAN STRESS (MPa)	ULTIMATE STRENGTH (MPa)	SHORTENING AT FAILURE (%)	COMMENTS, ^a FAULT ANGLE (°)
22	0	25	120	360	0.8	S
24	0	25	115	340	0.6	S
38	0	400	40	115	1.0	S
44	0	400	40	115	0.8	S
511	0	400	60	175	1.3	S
512	0	400	60	175	1.1	18, S
513 ^b	0	400	55	165	1.1	S
514 ^b	0	400	55	160	1.1	S
564	50	25	270	665	2.1	29
T1	50	25	255	620	1.7	S
M3	50	25	260	630	1.8	35
T2	50	400	220	510	1.4	27
18	50	400	220	520	1.5	S
233	50	400	245	580	1.4	S
236	50	400	235	560	1.4	27
45	50	400	230	535	1.5	S
46	0	700	15	45	1.2	S
47	0	700	25	70	1.2	S
231	50	700	205	465	2.0	27, S
232	50	700	180	390	1.8	29, S
51	0	>1050	0	0	27.0	PM
86	0	1000	15	40	1.7	30, S
87	0	1000	10	30	1.9	30, S
118	50	1000	85	100	1.7	S
119	50	930	135	255	1.9	S
132	50	1000	140	275	1.8	36
133 ^c	50	25	260	635	1.5	30
216 ^d	0	25	20	55	1.8	S
217 ^e	0	25	20	60	2.2	S
218 ^e	0	25	20	60	1.8	S
227 ^g	50	25	235	560	2.5	27
238	100	700	305	615	1.9	26
239	100	700	320	655	1.9	27, S
247	100	885	250	445	2.2	31
257	100	965	205	315	2.0	S

- a. S = shattered (2 or more fractures [or faults]). PM = partial melting.
b. Strain rate is 10^{-7} sec⁻¹.
c. Confined specimen heated at 10°C/min to 1000°C and then deformed at 25°C and 50 MPa confining pressure.
d. Unconfined specimen heated to 1000°C at 10°C/min and then deformed at 25°C, unconfined.
e. Unconfined specimens heated to 900°C at 1.5°C/min and then deformed at 25°C, unconfined.
g. Unconfined specimen heated to 1000°C at 1.5°C/min, cooled and then deformed at 25°C and 50 MPa.

TABLE 3. EXPERIMENTAL DATA FOR ROOM-DRY CUERBIO BASALT

SPECIMEN NO.	CONFINING PRESSURE (MPa)	TEMP. (°C)	MEAN STRESS (MPa)	ULTIMATE STRENGTH (MPa)	SHORTENING AT FAILURE (%)	COMMENTS ^a & FAULT ANGLE (°)
201	0	25	100	300	0.8	S
204	0	25	100	305	0.8	S
196	0	575	70	210	0.6	S
208	0	900	70	205	1.2	S
212	0	1030	0	5	0.2	PM
202	50	25	185	400	1.6	S
203	50	25	230	540	1.2	28
209 ^b	50	25	215	500	1.2	28
193	50	600	165	345	0.8	S
194	50	600	150	300	0.8	S
197	50	700	150	300	0.9	28
199	50	925	120	205	0.3	40, S
210	50	940	100	145	1.2	28
198	50	965	70	65	1.3	42
205	50	1000	80	95	1.4	UF
220 ^c	0	25	70	210	1.4	S
222 ^d	0	25	75	220	1.8	S
224	50	25	235	550	1.3	32, S
246	100	700	255	465	1.3	30
277	100	720	240	425	1.2	28
253	100	910	175	230	1.2	30
279	100	950	140	115	1.5	S
248	100	1000	160	180	2.0	S

a. S = shattered (2 or more fractures or faults), PM = partial melting; UF = uniform flow and multiple fractures.

b. Specimen heated to 750°C at 10°C/min, cooled, and deformed at 25°C and 50 MPa.

c. Unconfined specimen heated to 905°C at 1.5°C/min, cooled, and deformed at 25°C.

d. Unconfined specimen heated to 1000°C at 1.5°C/min, cooled, and deformed at 25°C.

Figure 3. Stress-shortening curves for room-dry Mt. Hood Andesite deformed at the confining pressures (P_c) and temperatures shown.

Figure 4. Stress-shortening curves for room-dry Charcoal Granodiorite deformed at the confining pressures (P_c) and temperatures shown.

Figure 5. Stress-shortening curves for room-dry Cuerbio Basalt deformed at the confining pressures (P_c) and temperatures shown.

MT. HOOD ANDESITE

Figure 6. Ultimate strength versus temperature plot for Mt. Hood Andesite. Solid and open circles are data from room-dry specimens at confining pressures of 0, 50, and 100 MPa. Solid stars are data for water-saturated specimens with zero effective pressure. Triangles represent dry specimens heated to 930°-1000°C, cooled and deformed at 25°C, 50 MPa. The unconfined, dry strength at 25°C is virtually the same whether or not a specimen has been preheated and cooled.

CHARCOAL GRANODIORITE

Figure 7. Ultimate strength versus temperature plot for Charcoal Granodiorite. Solid circles are data points at confining pressures of 0, 50, and 100 MPa, room-dry specimens. Solid stars are data for water-saturated rocks with zero effective pressure. Triangles represent specimens heated dry to 1000°C, then cooled and deformed at 25°C, 50 MPa. Open square represents three dry, unconfined specimens heated to 900°-1000°C, then cooled and deformed at 25°C.

Figure 8. Ultimate strength versus temperature plot for Cuerbio Basalt. Solid and open circles are data from room-dry specimens at confining pressures of 0, 50, and 100 MPa. Solid stars are data for water-saturated specimens at zero effective pressure. Triangle represents a dry specimen heated to 750°C, then cooled and deformed at 25°C, 50 MPa. Open squares represent unconfined, dry specimens heated to 905°- 1000°C, then cooled and deformed at 25°C.

and 3 MPa/25°C unconfined. Our data are bracketed by those of Lindholm et al. (1974) on Dresser Basalt (a little stronger) and Murrell and Chakravarty (1973) on dolertie (a little weaker).

(3) The effect of confining pressure is to strengthen the rocks at each temperature (Figures 6, 7, and 8). However, this effect diminishes as T_m is approached, i.e., the lines connecting the 0, 50, and 100-MPa confining pressure data tend to converge at temperatures of 1000°C and higher. Plots of ultimate strength versus mean stress are reasonably linear at all temperatures as discussed below.

(4) The strengths of the rocks vanish when even partial melting occurs. This takes place at a nominal temperature of about 1030°C at the hotter end of a basalt specimen. A few unconfined tests were done on andesite and granodiorite at about 1250°C (not listed). Nearly complete melting locally was achieved as would be expected as the liquidus temperatures of all phases range from 1100° to 1240°C.

We have determined the equivalent viscosity, $\eta = (\sigma_1 - \sigma_3)/3\dot{\epsilon}$, of the unconfined, partially melted rocks under a constant axial load due merely to the weight of the loading piston, about 0.1 MPa on the 2-cm specimens, by measuring the ensuing axial strain as a function of time (creep test). For andesite and granodiorite viscosities are of the order of 10^7 N · s/m², and for the lesser melted basalt about 10^9 . These values agree well with those of Murase and McBirney (1973), determined by a quite different technique, beam-bending.

(5) All three rocks are brittle throughout the temperature range explored until melting occurs (Tables 1-3). Shortenings at failure are less than 3 percent. Between 25° and 700°C the shortening at failure

tends to be either constant or even to decrease.

(6) Most of the stress-shortening curves exhibit pronounced work-softening after the onset of macroscopic failure. This reflects displacement along faults that have not lost total resistance to differential stress.

(7) A few tests on the andesite and granodiorite were made at 10^{-7} /s, 25° and 400°C, unconfined and 50 MPa (Tables 1 and 2). Except for a 30 percent weakening of the andesite at room temperature and 50 MPa, the 1000-fold decrease in strain rate from 10^{-4} /s has negligible effects on strength or ductility.

(8) Two types of tests were made to investigate the cause(s) of the temperature-weakening. All three rocks were first heated at 50 MPa-confining pressure to temperatures between 750° and 1000°C, cooled to 25°C, and then deformed at confining pressure. The rocks are as strong after thermal cycling as before (Figures 6, 7, and 8), so that the thermal cracking caused by the relatively rapid heating and cooling (1.5 °C/min or 10°C/min) is not the cause of strength reductions. Indeed, specimens of the andesite are an average of 25 percent stronger after thermal treatment (Figure 6), for reasons not yet understood. On the other hand, similar thermal treatment greatly reduces the strength of the unconfined rocks presumably because the effect of thermal cracking has not been suppressed by confining pressure.

In one inadvertant experiment, a specimen of andesite was heated to 980°C under 50 MPa-confining pressure and then deformed (Figure 3). During the shortening the internal furnace failed and the temperature fell to 880°C when the test was terminated at 6 percent strain, short of macroscopic

failure. Instead of failing at 1 or 2 percent strain, as expected from previous experiments at constant temperature, the specimen work-hardened from 165 MPa at 930°C to 190 MPa at 900°C to 195 MPa at 880°C. The strengths at these temperatures fall on the curve established by constant-temperature experiments. These data need to be reproduced, but these preliminary results certainly suggest that temperature affects fracture strength, and does not merely degrade the strength of the rock through thermal cracking, at least under confining pressure.

Water-Saturated Experiments. To date 18 successful experiments have been conducted on water-saturated specimens of the three rocks at zero effective confining pressure, temperatures from 400° to 1015°C, and at a constant strain rate of 10^{-4} s^{-1} (Table 4). The stress-strain curves from these tests all show nearly linear rise to a peak differential stress (ultimate strength) at strains of ≤ 1.0 percent and then a work-softening phase to axial strains up to 4.5 percent (Figure 9). Displacements during work softening probably are along multiple macroscopic fractures (shattered specimens) or along discrete faults (Table 4). All tests were stopped prior to complete loss of cohesion.

Ultimate (fracture) strengths of the wet granodiorite show no water-weakening (Figure 7). In fact, between 400° and 1015°C the specimens are somewhat stronger than their dry counterparts. This suggests that the pore fluid probably is not in pressure equilibrium throughout the specimen so the effective pressure ($P_c - P_p$) is > 0 (but much less than 50 MPa). In contrast, the Mt. Hood Andesite shows pronounced weakening between 850 and 880°C, at which point its fracture strength is between 5 and 25 MPa (Figure 6). In comparison its dry strength is 75 MPa at

Table 4 Experimental Data for Water-Saturated Charcoal Granodiorite, Cuerbio Basalt and Mt. Hood Andesite Shortened at 10^{-4} s^{-1} .

Rock Type and Specimen No.	Confining Pressure (MPa)	Pore Pressure (MPa)	Effective Pressure (MPa)	Temp. (°C)	Mean Stress (MPa)	Ultimate Strength (MPa)	Shortening at Failure (%)	
CHARCOAL GRANODIORITE								
280	100	100	0	400	50	150	0.7	Shattered, work-softened to 50 MPa at 1.9% shortening.
286	50	45	5	460	50	150	-	θ = 23°, } Heated at differential stress of 150 MPa until specimen failed. Shattered)
287	50	45	5	610	50	150	-	
289	50	50	0	740	35	100	1.0	Shattered, work-softened along multiple fractures to 45 MPa at 2.7% shortening.
291	50	50	0	835	25	75	1.0	θ = 34°, work-softened along fault to 15 MPa at 2.5% shortening.
294	50	50	0	945	20	55	0.9	θ = 34°, work-softened along fault to 15 MPa at 2.1% shortening.
308	50	50	0	1015	15	40	0.6	Shattered, work-softened along fault to 25 MPa at 3.2% shortening.
CUERBIO BASALT								
296	50	50	0	415	85	255	0.6	Shattered, work-softened along multiple fractures to 115 MPa at 1.7% shortening.
304	50	50	0	765	80	235	0.8	Shattered, work-softened with stick-slip to 135 MPa at 2.5% shortening.
307	50	50	0	916	30	90	0.9	Shattered, work-softened along multiple fractures to 60 MPa at 3.4% shortening.
309	50	50	0	850	45	135	0.3	Shattered.
310	50	50	0	945	20	60	0.1	Shattered, work softened along multiple fractures to 45 MPa at 2.7% shortening.
MT. HOOD ANDESITE								
306	50	50	0	615	35	105	0.6	Shattered, work-softened along multiple fractures to 25 MPa at 1.8% shortening.
290	50	50	0	720	25	70	0.6	Shattered, work-softened along multiple fractures to 25 MPa at 2.0% shortening.
300	50	50	0	800	25	70	0.6	Shattered, work-softened along multiple fractures to 20 MPa at 3.5% shortening.
301	50	50	0	845	20	65	0.8	Shattered, work-softened along multiple fractures to 15 MPa at 4.0% shortening.
302	50	50	0	855	15	40	0.8	Shattered, work-softened along multiple fractures to 15 MPa at 4.3% shortening.
303	50	50	0	870	10	35	0.8	Shattered, work-softened along multiple fractures to 20 MPa at 3.1% shortening.
297	50	50	0	870	0	5	1.0	Ductile, no fracturing or faulting, total shortening 2.8% before test stopped.
305	50	50	0	880	10	25	0.7	Shattered, work-softened along multiple fractures to 10 MPa at 2.3% shortening.

H_2O -SATURATED TESTS, $P_e=0$

Figure 9. Stress-shortening curves for water-saturated specimens of the three crystalline rocks deformed at zero effective confining pressure (confining and pore pressures equal to 50 MPa) and at temperatures shown.

1000°C. The andesite has a much higher porosity than the granodiorite so that pervasive water-pressure equilibration is more likely. Certainly the lower strength of wet andesite suggests P_e is small if not zero, and there does appear to be a water-weakening effect. Ultimate strengths are available for only five wet specimens of the basalt (Figure 8). Although sparse, the data suggest no weakening or apparent strengthening at temperatures to 763°C. There is, perhaps some weakening at >850°C. Clearly more data are needed for this rock, and conclusions made below are based primarily on the andesite and granodiorite.

To date, only systematic macroscopic observations of the deformed specimens have been made. Most of the specimens exhibit two or more shear fractures (noted as "shattered," Table 1) with their traces clearly imprinted on the copper jackets. Only three of the granodiorite specimens deformed along single discrete shear fractures. Clearly, the rocks are remaining brittle in their mechanical response to the test conditions. Granodiorite specimen 308 (1015°C) and andesite specimen 297 (870°C) were welded to their end pieces so that fragments of the rock adhered to the end piece upon disassembly. Some local melting was suspected but none was observed with a binocular microscope or in thin sections cut from these specimens.

Obsidian

Short-Term Behavior. The strength and ductility of dry Newberry Rhyolite Obsidian has been determined at 0 and 50 MPa pressure, at temperatures to 800°C, and at a strain rate of 2×10^{-4} /s (Table 5, Figures 10 and 11). Ultimate strength decreases dramatically from 665 MPa at 600°C to 15 MPa at 800°C (Figure 12); the brittle-ductile transition

Table 5. Experimental Data for Dry, Newberry Rhyolite Deformed at 10^{-4} Sec $^{-1}$ *.

Specimen No.	Confining Pressure (MPa)	Temp. (°C)	Mean Stress (MPa)	Ultimate Strength (MPa)	Shortening at Failure (F) at Yielding (Y) (%)	Comments
144	0	600	222	666	3.5 (F)	Shattered
145	0	700	47	141	1.0-2.0 (Y)	Shortened to 6.2% then allowed to relax
135	0	700	76	228	2.0-4.0 (Y)	Shortened to 7.2% then allowed to relax
149	0	750	9	27	0.7 (Y)	Shortened to 3.8% then allowed to relax
151	0	750	10	30	0.7 (Y)	Shortened to 4.6% then allowed to relax
153**c	0	770	17	50	0.5-1.0 (Y)	Shortened to 4.2% then allowed to relax
153a	0	800	5	14	0.6 (Y)	Shortened to 3.1% then allowed to relax
153b	0	800	5	15	0.4 (Y)	Shortened to 2.9% then allowed to relax
134	50	400	323	>820	4.1 (F)	Shattered, possible fracture in lucalox spacer at $\Delta\sigma$ of 670 MPa; specimen failure related to failure of lucalox
168	50	700	155	315	0.8 (Y)	Shortened to 9.7% then allowed to relax
170	50	700	165	344	0.8 (Y)	Shortened to 9.0% then allowed to relax
172c**	50	735	100	148	0.5 (Y)	Shortened to 5.3% then allowed to relax
169	50	750	80	91	0.5 (Y)	Shortened to 8.3% then allowed to relax
172b	50	765	61	32	0.1 (Y)	Shortened to 2.6% then allowed to relax
172a	50	800	52	5	0.1 (Y)	Shortened to 2.0% These specimens are too
171	50	800	52	7	0.1 (Y)	Shortened to 6.8% } weak for relaxation tests

* Strain rates actually vary from 2.2×10^{-4} sec $^{-1}$ to 4.4×10^{-4} sec $^{-1}$, however effects of these small differences on strength and ductility are negligible.

** Specimens 153 and 172 were deformed three times in the order a, b, and then c. For each cycle of loading the same specimen was shortened, relaxed, and then reloaded.

Figure 10. Stress-shortening curves for room-dry Newberry Rhyolite Obsidian at temperatures to 800°C . Experimental conditions and specimen numbers (in parentheses) are listed after each curve. Note scale change for insert plot.

Figure 11. Stress-shortening curves for Newberry Rhyolite Obsidian at 50 MPa-confining pressure, temperatures to 800°C, and strain rate of about $3 \times 10^{-4} \text{ sec}^{-1}$. Temperature and specimen numbers are listed after each curve.

Figure 12. Ultimate strength versus temperature plot for room dry Newberry Rhyolite Obsidian.

occurs between 600° and 700°C; and work-hardening ceases and steady-state flow obtains between 700° and 750°C.

Data at 50 MPa are not appreciably different (Table 5). Between 700° and 750°C the obsidian is somewhat stronger at 50 MPa than unconfined, but the opposite is true between 750° and 800°C. In short, the increment of 50 MPa in confining pressure has insignificant effects on short-time strength and ductility.

Time-Dependent Flow. Stress-relaxation tests (Lee and Hart, 1971; Rutter et al., 1978) were made on 7 unconfined and 5 confined specimens (50 MPa) at temperatures between 700° and 800°C (Table 5). In these experiments, the loading pump is switched off at the end of the usual constant strain-rate path. The load supported by the specimen then decays with time as the stored elastic strains in the machine and specimen are converted into permanent strain within the specimens.

Assuming that the volume of hydraulic fluid in the loading ram remains constant during the relaxation, and that any anelastic relaxations of the machine are negligible, one can relate the permanent displacement, Δd_p , induced in the specimen to the force drop, ΔF , through the machine stiffness, k_m , and the specimen stiffness, k_s , as

$$\Delta d_p = -\left(\frac{1}{k_m} + \frac{1}{k_s}\right)\Delta F. \quad (1)$$

For specimens that are essentially inelastic (viscous) in behavior, the equation reduces to:

$$\Delta d_p = -\frac{1}{k_m} \Delta F. \quad (2)$$

Taking time derivatives, one can write

$$\dot{d}_p = -\frac{1}{k_m} \dot{F}. \quad (3)$$

Thus, by plotting curves of F against t and taking the gradient at specific points on the ensuing curve, one finds a data set relating F and $\dot{d}_p$, provided that the stiffnesses of the relevant parts of the machine are known. For these particular tests, the total machine stiffness, $1.23 \times 10^8 \text{ N/m}$ is relevant. Strain rate, $\dot{\epsilon}$, and differential stress ($\sigma_1 - \sigma_3$), are then calculated from the known length and cross-sectional area of the specimen at the start of the relaxation. Since the total induced strain during relaxation is very small ($< 0.5\%$), it is not necessary to correct for changes in length or area. By this method we have obtained steady-state strength data at strain rates from 10^{-4} to $10^{-6}/\text{s}$.

Equivalent Viscosities. From data on $\dot{\epsilon}$ and ($\sigma_1 - \sigma_3$) during relaxation, equivalent viscosities are calculated for a given temperature. Unconfined viscosities at 10^{-4} s^{-1} decrease linearly from an average of 3.4×10^{11} at 700°C to about $3 \times 10^{10} \text{ N} \cdot \text{s/m}^2$ at 800°C . These values agree closely with those determined for this same obsidian at the same temperatures, but with the beam-bending method by Murase and McBirney (1973, Figure 5). At 50 MPa-confining pressure, the equivalent viscosities decrease with increasing temperature along a slightly steeper slope, but the effect of the pressure increment is insignificant. The specimens deformed at 50 MPa and 800°C were too weak for meaningful stress-relaxation testing (Table 5).

Plot of $\log \dot{\epsilon}$ against ($\sigma_1 - \sigma_3$) show a systematic variation with temperature at both confining pressures (Figure 13). The data fit a power law, $\dot{\epsilon} = (\sigma_1 - \sigma_3)^n$, with linear least-squares values for n ranging

Figure 13. Log strain rate versus log steady-state flow stress plot for dry Newberry Rhyolite Obsidian at different temperatures and confining pressures.

from about 2.0 at 700°C to about 1.6 at 800°C unconfined, and 1.3 to 1.5, confined.

Finally, a flow law of the type

$$\dot{\epsilon} = A \exp(-Q/RT) (\sigma_1 - \sigma_3)^n \quad (4)$$

is fit to the data between 700° and 800°C. The constant A is 2×10^{13} when σ is in MPa and $\dot{\epsilon}$ is s^{-1} ; the activation energy for viscous flow is $Q = 380$ kJ/mol; R is the gas constant; and $n \approx 1.5$. (By using $n = 1.5$, we ignore the very small effect of confining pressure.) Our value of Q agrees remarkably well with the 400 kJ/mol of Murase and McBirney (1973, Figure 6). We emphasize that this flow law has been determined only for temperatures between 700 and 800°C and $\dot{\epsilon}$ from 10^{-4} to $10^{-6} s^{-1}$, and that the parameter, n , varies with temperature and strain rate.

PETROFABRIC OBSERVATIONS

All specimens have been examined macroscopically, and thin- and polished-sections from over 50 dry specimens have been studied in detail to characterize the operative deformation mechanisms as functions of temperature and pressure.* Quantitative data consist of the angle (θ) between the load axis (σ_1) and the macroscopic fault zone (Tables 1-4), microfracture abundance, and chemical composition of glasses developed upon partial melting. These are accompanied by qualitative assessments of microfracture orientation and distribution, thermal alteration of minerals, gliding flow in mafic minerals, and development of glass in partially melted rocks.

Macrofractures

Studies of deformation mechanisms are made to gain insight into temperature effects on strength and ductility. All the rocks deform primarily by brittle fracture until partial melting (crystalline rocks) or glass softening (obsidian) occurs. Most specimens are shattered, i.e., they contain multiple faults (Tables 1-5). About 45 percent of the andesite, granodiorite and basalt specimens contain a discrete fault from which a reliable measure of θ can be obtained. This angle ranges between 23° and 34° for the andesite, 18° and 36° for the granodiorite, and 28° and 42° for the basalt. In general, θ tends to increase with increasing temperature. There is also a tendency for the fault zones to widen with increasing temperature and pressure.

*Thin section analyses of the H_2O -saturated specimens will not be made in time for inclusion in this report. They will be reported as part of the continuation of this effort funded by DOE.

By comparing the data from the few tests on the andesite at 25° and 400°C and 0 and 50 MPa done at strain rates of both 10^{-4} and 10^{-7} /s (Table 1), we note an unexpectedly large effect of time on the fracture angle, nearly 8°, not previously reported in the literature as far as we know (Figure 14).

Microfractures

Microscopically each fault zone (Figures 15 and 16) consists of pervasive, precursive, extension microfractures (oriented within 20° of σ_1) and adjacent microscopic feather fractures (Friedman, 1975; Conrad and Friedman, 1976). Excellent examples are found of incipient faults along which en-echelon arrays of the precursive axial microfractures define the path of the eventual fault (Figures 15 and 16). Thermal fractures also occur; they are distinguishable by being relatively short, randomly oriented with respect to external coordinates, and located close to grain boundaries and/or larger fractures.

Detailed studies of microfracture abundances in the granodiorite and andesite have been made, and abundance is quantified as the average number of microfractures per mm intersected along linear traverses oriented perpendicular to the cylinder (load) axis. This abundance factor is obtained from 5 to 9 traverses with cumulative lengths of from 4.4 to 16.1 mm, spaced 2 to 5 mm apart for each specimen. Traverses across gouge either ignore it or are equated to discrete microfractures by arbitrarily assigning 100 microfractures for each millimeter of finely comminuted gouge. Both load-induced and thermoelastic fractures are counted. (A measure of the latter is afforded by study of specimens subjected only to

Figure 14. Plot of θ , the angle between fault surface and σ_1 across boundaries of specimen, against strain rate.

Figure 15. Specimens show that fracture is the major mechanism of deformation in both the granodiorite and andesite to temperatures $\geq 1000^{\circ}\text{C}$.

(a) Photograph of two specimens of granodiorite in their copper jackets. Outlines of macrofractures in copper clearly show the deformation of this rock still is brittle at 920° and 1000°C . Cylinders are 4.0 cm long.

(b) Polished section of specimen No. 87, deformed dry and unconfined at 1000°C , shows a fault zone consisting of coalesced, en echelon extension fractures. Cylinder is 4.0 cm long.

(c) Photomicrograph of a right-lateral fault zone in granodiorite specimen No. 132, deformed dry at 50 MPa and 1000°C . The long axis of the picture corresponds to the orientation of σ_1 across the boundaries of the specimen during deformation. Especially conspicuous are microscopic feather fractures in the host feldspar grains immediately adjacent to the shear zone. Crossed-polarized light. Scale line is 0.05 mm.

(d) Photomicrograph shows microfractures in feldspar phenocryst and finer-grained groundmass of the andesite have the same orientation and configuration in this specimen (No. 85) deformed at 1000°C as they do in specimens deformed at 25° , 400° , or 700°C . Plane polarized light. Scale line is 0.5 mm.

a

b

c

d

Figure 16. Photomicrographs show brittle deformation of the Mount Hood Andesite at 700°C, confining pressures to 50 MPa, and at a strain rate of 10^{-4} sec^{-1} . (a) Specimen 36, deformed unconfined, shows traces of macroscopic extension (vertical) and shear fractures (inclined, see arrow). (b) Specimen 35, deformed at 50 MPa, shows an incipient fault zone (see outlined zone) details of which are shown in (c, d, and e). Microfractures (arrows) in Specimen 35 are subparallel to the greatest principal stress (σ_1) across the boundaries of the specimen and lie en-echelon arrangement along the fault zone (b). They cross phenocrysts and glassy ground mass without major change in direction. Plane polarized light. Scale line for (a and b) is 1.0 cm, that for (c, d, and e) is 0.05 cm.

a

b

c

d

e

thermal treatment.) Data for all three crystalline rocks are similar, and those shown for the granodiorite and andesite are representative (Figures 17 and 18, respectively).

Clearly, the total number of both load-induced and thermal cracks increases with increasing temperature, about equally for all three rocks, and in the andesite about equally in the very-fine grained groundmass and in the phenocrysts. The increase in microfracture density of about 5 per mm from that in the starting material to that in specimens deformed at room temperature is essentially identical to that in the Westerly Granite under similar experimental conditions (Tapponnier and Brace, 1976, Figure 3). Subtracting the amount of thermal cracking from the total abundances of microfractures (Figure 17), reveals that the degree of microfracturing caused by loading alone is about the same at all temperatures up to 1000°C, above which partial melting occurs and thermal cracking increases dramatically (not shown in the figure where up to 1000°C the increase is linear).

The work-softening exhibited by many of the specimens (Figures 3-5, and 9) occurs during displacements along the macroscopic faults. Thin-section observations reveal that this softening is accompanied by local ductile flow in the groundmass of the andesite, the development of microscopic feather fractures and of gouge, and kinking of favorably oriented biotite and other types of gliding flow in mafic minerals immediately adjacent to the faults. Some of these same mechanisms, along with thermal alteration of hornblende, biotite and pyroxene, probably also operate prior to macroscopic failure. However, in light of the strength and ductility trends, these mechanisms, if significantly influential at all, apparently contribute more to the weakening of the rocks than to the enhancement of their ductility, for virtually all of the permanent

Figure 17. Microfracture abundance as functions of pressure and temperature in deformed, room dry Charcoal Granodiorite. Specimen 133 (▲) was heated to 1000°C, then cooled and deformed at 25°C, 50 MPa; the higher result includes gouge, and lower one does not.

MT HOOD ANDESITE

Figure 18. Relations between microfracture abundance and temperature for hypocrystalline groundmass and phenocrysts in Mt. Hood Andesite, deformed dry at 10^{-4} sec $^{-1}$ and at 0 and 50 MPa confining pressures.

shortening is accountable to the macroscopic faulting in the brittle regime, even at 1000°C.

Partial Melting

The partial melting in the andesite and granodiorite rocks (Figures 19-22) produces two types of glass. Under the microscope in transmitted light, one is brownish-amber exhibits flow-banding, contains relatively few bubbles (Figure 20). Crystallites produced upon devitrification are common (Figure 21). The other glass is colorless and highly vesicular, and flow herein produces elongated vesicles (Figure 20). In granodiorite specimens with ancipient partial melting the amber glass, clearly derived from the hornblende occurs first as the temperature rises, i.e., the hornblende melts at a lower temperature than does the quartz, feldspar or other mafics (Figure 21a). Electron-microprobe analyses by Drs. P. F. Hlava and E. J. Graeber of Sandia Laboratories confirm the petrologic evidence that the brownish-amber and clear glasses arise from melting of mafic minerals and feldspar, respectively (Table 6).

Figure 19. Photograph shows partially melted specimen of Mt. Hood Andesite. Item (1) is upper thermocouple wire, (2) glass, (3) furnace windings and ceramic, and (4) melted copper jacket. Specimen is about 4.0 cm long. Microscopic details are shown in Figure 22.

Figure 20. Photomicrograph of flow in two glasses formed during partial melting and deformation of Charcoal Granodiorite. Item (1) is flow banding in amber glass; (2) elongated vesicles in colorless glass. Plane-polarized light; scale line is 0.25 mm.

Figure 21. Photomicrographs of amber glass development in Charcoal Granodiorite. (a) View shows relation between original hornblende grain (H), altered hornblende, opaque oxidized zone, and amber glass with devitrified crystallines (arrow). (b) Example of crystallines formed upon devitrification of glass. Plane-polarized light, scale lines are 0.5 mm for (a) and 0.25 mm for (b).

a

b

c

Figure 22. Photomicrographs of partially melted Mt. Hood Andesite. (a) Reflected light view of polished surface shows vesicular melted zone, ceramic furnace and furnace wires. (b) Thin section view of same areas as (a) shows melted zones surrounding darker unmelted area. Opaque areas are oxidized iron-rich minerals biotite and hornblende. Vesicles and furnace are conspicuous. (c) Detail of lower left corner of (b) shows unmelted rock (short arrow), amber glass (straight arrow), highly vesicular colorless glass (curved arrow), and large vesicles. Scale lines are 7.5 mm in (a and b) and 0.5 mm for (c). Plane-polarized light.

Table 6. Electron Microprobe Analysis¹ of Two Glasses from Partially Melted Mount Hood Andesite and Charcoal Granodiorite.

	Mount Hood Andesite		Charcoal Granodiorite	
	Brown Glass	Colorless Glass	Brown Glass	Colorless Glass
	Wt. %	Wt. %	Wt. %	Wt. %
SiO ₂	55.95	62.78	58.35	62.38
TiO ₂	2.15	1.17	1.37	0.95
Al ₂ O ₃	9.82	13.74	10.28	15.58
FeO	12.46	6.96	9.52	2.88
MnO	0.24	0.09	0.18	0.05
MgO	8.48	3.26	5.93	1.87
CaO	4.57	5.50	6.42	4.42
Na ₂ O	3.13	4.20	3.40	4.79
K ₂ O	1.00	1.46	4.04	7.05
P ₂ O ₅	0.46	0.38	0.32	0.26
Others	1.74	0.46	0.19	0.08

1. Analyses made at the Sandia Laboratories by P. F. Hlava, courtesy of E. J. Graeber.

BOREHOLE STABILITY

Borehole Stresses

Ideally borehole stresses depend on the regional state of stress in the undisturbed rock, and this is not generally known a priori. Few measurements of the in-situ stress have been made in crystalline igneous rocks at depths of more than a few meters (Table 7). Using the hydraulic-fracturing technique, the staff of the Los Alamos Scientific Laboratory has determined the stress at a depth of 2926 m in the Precambrian crystalline basement of the Jemez Mountains in connection with its experiment on geothermal-energy extraction from dry hot rock. Taken as the weight per unit area of the superincumbent rock, the vertical stress, σ_v , is 72 MPa, and the minimum horizontal stress, σ_h , is 34 MPa (Brown, 1976). Hydraulic-fracturing tests by Haimson (1976), and Haimson and Voight (1977) in granitic rocks in the United States and basalt in Iceland, respectively, reveal that the horizontal stresses may be greater or less than the vertical stresses at depths in the range of 100 to 400 m. Strain-relief measurements in a South African dunite disclose an average horizontal stress at 350 m that greatly exceeds the overburden stress (Gay, 1975).

Since available data are too few for meaningful generalizations, we cannot predict the ratios of vertical to horizontal stresses elsewhere. For simplicity let us assume, therefore, that all horizontal stresses are equal and that σ_v/σ_h is 0.5, 1, or 2. Values falling much outside these extremes are unlikely, albeit possible.

TABLE 7. SOME IN-SITU STRESS MEASUREMENTS IN IGNEOUS ROCKS

PLACE	ROCK	DEPTH (M)	VERTICAL (OVERBURDEN)	STRESSES (MPa) MAXIMUM HORIZONTAL	MINIMUM HORIZONTAL	REF.
CALIFORNIA	GRANITE	300	8.2	9.5	5.4	Haimson, 1976
S. CAROLINA	GRANITE	230	6.2	22.8	15.9	Haimson, 1976
WISCONSIN	GRANITE	136	3.5	16.0	7.0	Haimson, 1976
N. MEXICO	GRANITE	2926	71.7	?	34.1	Brown, 1976
ICELAND	BASALT	203	5.4	6.3	4.1	Haimson and Voight, 1977
ICELAND	BASALT	375	10.0	7.0	6.4	Haimson and Voight, 1977
S. AFRICA	DUNITE	350	11.5	15.5 (average)		Gay, 1975

At by far most locations where the complete in-situ stress tensor has been determined, σ_v is indeed equivalent to the weight per unit area of the overburden and it lies in a principal plane or close to it (Haimson, 1976). Let us assume that before the rock is drilled the state of stress is

$$\begin{aligned}\sigma_x &= \sigma_y = \sigma_h \\ \sigma_z &= \rho_r g z = \sigma_v \\ \sigma_h &= b \sigma_v,\end{aligned}\tag{5}$$

where $b = 0.5, 1$, or 2 ; z (m) is depth below the free surface; ρ_r (kg/m^3) is the average bulk density of the overburden; g (m/s^2) is acceleration of gravity; and compressive stresses (MPa) are counted positive.

In cylindrical coordinates (Figure 23) the vertical axial stress, σ'_z , and the horizontal radial and tangential stresses, σ'_r and σ'_θ , in the undisturbed rock are in general

$$\begin{aligned}\sigma'_r &= \frac{\sigma_x + \sigma_y}{2} + \frac{\sigma_x - \sigma_y}{2} \cos 2\theta \\ \sigma'_\theta &= \frac{\sigma_x + \sigma_y}{2} - \frac{\sigma_x - \sigma_y}{2} \cos 2\theta \\ \sigma'_z &= \rho_r g z,\end{aligned}\tag{6}$$

where the angle θ is measured counterclockwise from the x direction, and $\sigma_x > \sigma_y$.

Now suppose that a vertical, perfectly cylindrical hole is drilled into the stressed mass of dry, statistically homogeneous and isotropic, ideally elastic rock. The original far-field stresses are perturbed in

Figure 23. Radial and tangential stresses, σ_r and σ_θ , around a cylindrical hole of radius, a , in an elastic, isotropic plate. Far-field horizontal stresses are $\sigma_x > \sigma_y$; vertical stress is σ_z . Borehole fluid pressure is p_b , and pore pressure in a porous medium is p_p .

the vicinity of the hole where the new stresses can be computed from the Kirsch equations (see, for example, Timoshenko and Goodier, 1970, p. 90-93).

$$\begin{aligned}\sigma_r &= \frac{\sigma_x + \sigma_y}{2} \left(1 - \frac{a^2}{r^2}\right) + \frac{\sigma_x - \sigma_y}{2} \left(1 - \frac{4a^2}{r^2} + \frac{3a^4}{r^4}\right) \cos 2\theta \\ \sigma_\theta &= \frac{\sigma_x + \sigma_y}{2} \left(1 + \frac{a^2}{r^2}\right) - \frac{\sigma_x - \sigma_y}{2} \left(1 + \frac{3a^4}{r^4}\right) \cos 2\theta ,\end{aligned}\tag{7}$$

where a is the radius of the hole and σ_x and σ_y are the original far-field stresses. Since we are concerned with the stresses at the wall, let $r = a$. Then

$$\begin{aligned}\sigma_r &= \tau_{r\theta} = 0 \\ \sigma_\theta &= \sigma_x + \sigma_y - 2(\sigma_x - \sigma_y) \cos 2\theta.\end{aligned}\tag{8}$$

The stress changes in plane strain are

$$\begin{aligned}\Delta\sigma_r &= \sigma_r - \sigma'_r = -\left(\frac{\sigma_x + \sigma_y}{2}\right) + \left(\frac{\sigma_x - \sigma_y}{2}\right) \cos 2\theta \\ \Delta\sigma_\theta &= \sigma_\theta - \sigma'_\theta = \frac{\sigma_x + \sigma_y}{2} - \frac{3(\sigma_x - \sigma_y)}{2} \cos 2\theta \\ \Delta\sigma_z &= \nu(\Delta\sigma_r + \Delta\sigma_\theta) = -2\nu(\sigma_x - \sigma_y) \cos 2\theta ,\end{aligned}\tag{9}$$

where ν = Poisson's ratio, about 0.25 for rock.

Adding these stress changes to the initial stresses (equation 6), assuming that $\sigma_x = \sigma_y = \sigma_h$ (equation 5), and solving for the maximum values at $\cos 2\theta = 1$ gives

$$\begin{aligned}\sigma_r &= 0 \\ \sigma_\theta &= 2\sigma_h \\ \sigma_z &= \rho_r g z.\end{aligned}\tag{10}$$

Since $\tau_{r\theta} = 0$, these are principal stresses.

For an average bulk density, $\rho_r = 2.7 \times 10^3 \text{ kg/m}^3$, the vertical stress, σ_z , increases at the rate of 27 MPa/km.

Although well completions at near-magmatic temperature may require work in an open hole, let us suppose that a hydrostatic borehole pressure, p_b , can be maintained by a drilling fluid which cannot penetrate the surrounding impermeable rock. Then the stresses are

$$\sigma_r = p_b$$

$$\sigma_\theta = 2\sigma_h - p_b \quad (11)$$

$$\sigma_z = \rho_r g z.$$

Conventional, heavy, oil-well drilling muds can be formulated for a density as high as $2 \times 10^3 \text{ kg/m}^3$, but they are not stable at the high temperatures of concern, so we suppose that the hole is filled with water with a density close to $1 \times 10^3 \text{ kg/m}^3$ at 25°C. (The density is really less at high temperatures.) The borehole pressure, p_b , then increases at a rate of 10 MPa/km.

All crustal rocks have finite porosity, owing to intergranular voids and intragranular microcracks, and commonly to unsealed macrofractures as well. Let us suppose that these spaces are saturated with ground water under a hydrostatic pore-pressure, p_p , which increases at the rate of 10 MPa/km. To find the effective stresses, $\bar{\sigma}$, in the solid rock around the open borehole, we subtract p_p from all the normal stresses in equations 10 and 11. The vertical stress, $\bar{\sigma}_z = (\rho_r - \rho_w) g z$, increases at a rate of about 17 MPa/km.

Should the wallrock be permeable, the pore pressure in the vicinity of an open hole would bleed off, and the stresses would approach those given by equation 10. Around the water-filled well the pore pressure and borehole pressures would equalize, and the stresses would approach those given by equation 11.

The principal stresses ($\sigma_1 > \sigma_2 > \sigma_3$) given by equations 10 and 11 together with differential stresses, $\sigma_1 - \sigma_3$, and mean stresses, $\sigma_m = (\sigma_1 + \sigma_2 + \sigma_3)/3$, are given in Table 8 for the different ratios, σ_h/σ_v . Note that for a water-filled hole in water-saturated rock, the tangential stress can become tensile when the far-field horizontal stress is low relative to the overburden stress. This state of stress would favor unintentional hydraulic fracturing and lost circulation of the drilling fluid.

Our estimates of the borehole stresses should be regarded as minimal because the stresses are higher where $\sigma_x \neq \sigma_y$, as is commonly observed in nature, and because the walls of a real drillhole are not smooth, so that stress concentrations may be much higher than we have assumed ideally. On the other hand, thermoelastic stresses, introduced by temperature difference between borehole and country rock, have not been considered. If the borehole were cooler than the rock mass, the thermoelastic stresses would subtract from the differential stresses at the borehole wall and thus tend to stabilize the borehole.

Estimation of Borehole Stability

Dry Rocks. Recall that in the triaxial compression test the principal stresses are $\sigma_1 > \sigma_2 = \sigma_3 = P_c$, whereas in the wall of the borehole, $\sigma_1 > \sigma_2 > \sigma_3$. In order to use the laboratory data to predict failure of

TABLE 8. BOREHOLE STRESSES (MPa/km)

CONDITION		$\frac{\sigma_h}{\sigma_v}$	σ_1	σ_2	σ_3	$\sigma_1 - \sigma_3$	σ_m
DRY ROCK	OPEN HOLE	2	$\sigma_\theta = 108$	$\sigma_z = 27$	$\sigma_r = 0$	108	45
		1	$\sigma_\theta = 54$	$\sigma_z = 27$	$\sigma_r = 0$	54	27
		0.5	$\sigma_\theta = 27$	$\sigma_z = 27$	$\sigma_r = 0$	27	18
	FILLED HOLE	2	$\sigma_\theta = 98$	$\sigma_z = 27$	$\sigma_r = 10$	88	45
		1	$\sigma_\theta = 44$	$\sigma_z = 27$	$\sigma_r = 10$	34	27
		0.5	$\sigma_z = 27$	$\sigma_\theta = 17$	$\sigma_r = 10$	17	18
SATURATED ROCK	OPEN HOLE	2	$\bar{\sigma}_\theta = 88$	$\bar{\sigma}_z = 17$	$\bar{\sigma}_r = 0$	88	35
		1	$\bar{\sigma}_\theta = 34$	$\bar{\sigma}_z = 17$	$\bar{\sigma}_r = 0$	34	17
		0.5	$\bar{\sigma}_z = 17$	$\bar{\sigma}_\theta = 7$	$\bar{\sigma}_r = 0$	17	8
	FILLED HOLE	2	$\bar{\sigma}_\theta = 78$	$\bar{\sigma}_z = 17$	$\bar{\sigma}_r = 10$	68	35
		1	$\bar{\sigma}_\theta = 24$	$\bar{\sigma}_z = 17$	$\bar{\sigma}_r = 10$	14	17
		0.5	$\bar{\sigma}_z = 17$	$\bar{\sigma}_r = 10$	$\bar{\sigma}_\theta = -3$	20	8

$\sigma_z = \sigma_v = \rho_r gz = 27 \text{ MPa/km}$ for $\rho_r = 2.7 \times 10^3 \text{ kg/m}^3$.

$p_b = p_p = \rho_w gz = 10 \text{ MPa/km}$ for $\rho_w = 10^3 \text{ kg/m}^3$.

the borehole, we assume that ultimate compressive strength, $\sigma_1 - \sigma_3$, at failure, is a function only of mean stress, $\sigma_m = (\sigma_1 + \sigma_2 + \sigma_3)/3$, and temperature. We realize that this strength is not independent of the relative magnitude of the intermediate principal stress, σ_2 , but we have measured it at three different confining pressures ($P_c = 0, 50$, and 100 MPa) and find it is linearly proportional to mean pressure. However, the effect of σ_2 is second-order (Handin et al., 1967), and the relation between strength and mean pressure is very nearly linear for many rocks in the brittle state (Handin and Hager, 1957; Borg and Handin, 1966). Given all the other uncertainties of the predictions, we regard this failure criterion as good an approximation as any proposed (Handin et al., 1967).

Figure 24 shows an example of one way to associate strength data with borehole failure. Using the quantities listed in Table 8 for dry rock ($p_p = 0$) and the ratio, $\sigma_h/\sigma_v = 1$, we plot differential stress, $\sigma_1 - \sigma_3$, as a function of mean pressure, σ_m , at the borehole wall for an open hole ($p_b = 0$) and a water-filled hole ($p_b = 10$ MPa/km) in impermeable rock. Choosing the proper stress gradients, $\sigma_1 - \sigma_3 = 54$ and 34 MPa/km, respectively, we relate the stresses to depth below the surface. At any given depth, the borehole pressure reduces the differential stress and so tends to stabilize the hole.

On this diagram of the borehole stresses as functions of depth, we superimpose the plots of ultimate compressive strength versus mean pressure, using the data of Table 1 and Figure 6 for dry andesite at temperatures of 25° , 400° , 700° , and 900°C and confining pressures of 0 , 50 , and 100 MPa. At a given mean pressure, say 50 MPa, the strengths exceed the borehole

Figure 24. Assessment of borehole stability for dry Mt. Hood Andesite. Differential stress versus mean pressure at wall of an open hole (solid line and depth scale) and a "water"-filled hole (dashed line and depth scale) in impermeable rock. Superposed are ultimate compressive strength versus mean stress data. Each data point is the average of two or more tests at a given confining pressure and temperature.

differential stress, so that even the open hole is stable at all temperatures. The intersections of the two sets of curves mark the boundaries between the fields of stability, containing all points to the left of the loci of borehole stresses, and of instability, containing all points on the right-hand side. Since the strength curves are subparallel to each other and to the borehole-stress curves, these boundaries are not sharp. Nevertheless, they can be related approximately to depth, about 5, 4, 3, and 2 km at 25, 400, 700, and 900°C, respectively, for the open hole. The water-filled hole is stable at all temperatures to a depth of at least 10 km. Similar data for the dry basalt and granodiorite are plotted in Figures 25 and 26, respectively, and the maximum depth for stable boreholes in all three rocks are listed in Table 9 for a variety of conditions.

Examination of Figures 24-26 and Table 9 leads to certain generalizations concerning borehole stability based on the compressive strength data for the three dry crystalline rocks, namely:

1. The maximum depth for stable boreholes increases with decreasing temperature, decreasing σ_h/σ_v ratio, and with filled versus open hole.
2. For $\sigma_h/\sigma_v = 1.0$, boreholes filled with a medium whose average specific gravity is 1.0 will be stable in all three rocks to depths ≥ 10 km at temperatures $> 900^\circ\text{C}$. The single exception is dry basalt that would fail at 5 km depth at temperatures $> 900^\circ\text{C}$ (Figure 25).
3. Open holes are much less stable. With $\sigma_h/\sigma_v = 1.0$ borehole failures are likely at only 2 to 3 km depth if temperatures are $> 900^\circ\text{C}$. At 400°C the weakest rock (andesite) would begin to fail at 4 km, and most would be stable to > 10 km.
4. For dry rocks the σ_h/σ_v ratio of 0.5 affords the greatest

Figure 25. Assessment of borehole stability for dry Charcoal Granodiorite.

Figure 26. Assessment of borehole stability for dry Cuerbio Basalt.

TABLE 9. MAXIMUM DEPTH (KM) OF A STABLE BOREHOLE BASED ON STRENGTH AND MEAN STRESS DATA FOR DRY ROCKS

CONDITIONS		σ_h/σ_v	25°C	400°C	700°C	>900°C
ANDESITE	DRY	OPEN HOLE	2.0	2*	1	1
			1.0	4	3	2
			0.5	>10	>10	4
		FILLED HOLE	2.0	3	2	1
			1.0	>10	>10	>10
			0.5	>10	>10	>10
	SATURATED	OPEN HOLE	2.0	2	1	1
			1.0	7	5	3
			0.5	>10	>10	>10
		FILLED HOLE	2.0	4	3	2
			1.0	>10	>10	>10
			0.5	**	**	**
GRANODIORITE	DRY	OPEN HOLE	2.0	7	4	1
			1.0	>10	>10	2
			0.5	>10	>10	>10
		FILLED HOLE	2.0	>10	6	1
			1.0	>10	>10	>10
			0.5	>10	>10	>10
	SATURATED	OPEN HOLE	2.0	6	1	1
			1.0	>10	>10	2
			0.5	>10	>10	4
		FILLED HOLE	2.0	>10	>10	2
			1.0	>10	>10	>10
			0.5	**	**	**
BASALT	DRY	OPEN HOLE	2.0	4	--	1
			1.0	>10	--	2
			0.5	>10	>10	4
		FILLED HOLE	2.0	>10	4	1
			1.0	>10	>10	5
			0.5	>10	>10	>10
	SATURATED	OPEN HOLE	2.0	4	3	1
			1.0	>10	9	3
			0.5	>10	>10	7
		FILLED HOLE	2.0	>10	5	2
			1.0	>10	>10	>10
			0.5	**	**	**

* Rounded to nearest kilometer.

** Tangential stress is tensile.

stability. This far-field stress condition is most likely to occur in regions affected by horizontal extension of the crust.

5. For impervious, saturated rock with filled boreholes where σ_h/σ_v is 0.5 the tangential borehole stress is actually tensile increasing at -3 MPa/km. This would tend to favor tensile failure at a few kilometers depth; accordingly, no corresponding depths are indicated in Table 9. However, the gradient of borehole stresses for this condition (Table 8, $(\sigma_1 - \sigma_3) = 20$ MPa/km and $\sigma_m = 8$ MPa/km) constitute a "worst case" that is utilized below for evaluation of the water-saturated results. Thus, a σ_h/σ_v ratio of 0.5 is the most favorable case for stability in dry rocks and the least favorable one for wet ones.

These predictions are likely to be optimistic. Recall that the borehole stresses would be higher than supposed if the far-field horizontal stresses were not equal and if the stress concentrations in a real rough-surfaced hole were greater than those idealized for the calculations. Furthermore, we have not yet fully accounted for the effects of pore water (see below) and time. On the other hand superposed thermoelastic stresses from cooled boreholes in hot rock would diminish borehole stress differences and enhance stability.

We have emphasized the testing of room-dry rocks at a strain rate of 10^{-4} /s because data can be collected quickly and easily. Moreover the dry tests may be valid simulations of in situ conditions if "dry-out" zones exist immediately above buried magma chambers and if drilling is accomplished without water. The results are conservative in the sense that the measured strength is the highest and ductility is the lowest that can be expected. In other words, if a borehole in dry rock is potentially

unstable under short-time loading, then it will surely fail under long-term loading or where the rock has been water-weakened.

Water-Saturated Rocks. If dry-out zones are not present in the roof-rocks of buried magma chambers, the pores, grain boundaries, cracks, and macrofractures will be water-saturated. Moreover, water may be used in the drilling process and to cool the borehole walls. The influence of water on rock strength is two-fold. The purely mechanical effect is to reduce the effective borehole stresses (Table 6), according to Terzaghi's principle (Handin et al., 1963; Brace and Martin, 1968), and to extend somewhat the maximum depth of the stable field (Table 9). On the other hand, chemical effects involve hydration of the silanol bonds in silicates. The higher is the temperature, the more significant is this phenomenon likely to be (Griggs, 1967). Water-weakening can occur due to (a) chemisorption or stress corrosion effects at crack tips that reduce fracture strengths (e.g., see review by Anderson and Grew, 1977), (b) lowered melting points, and (c) lowering of the critical resolved shear stresses for dislocation gliding, i.e., yield strengths for crystal plastic mechanisms (e.g., Griggs and Blacic, 1965; Griggs, 1974; Balderman, 1974; Tullis, 1979). Without detailed thin section and perhaps TEM studies (scheduled for future work), it is not possible to say which of these mechanisms has produced the reduced strength of the andesite at elevated temperatures demonstrated above. But, it is generally recognized that H_2O must gain access to the crystal surfaces (item a, above) and then to their substructures (items b and c, above) to effect the hydrolytic weakening. It is not surprising therefore, that of the three rocks tested, the andesite shows the greatest weakening (Figures 6, 7, and 8) as its relatively

large porosity (10-12 percent) permits greater access of the pore water to the solid interfaces.

In the ductile regime it has been recognized that there is a temperature effect on the water-weakening of quartz (Griggs, 1967; Balderman, 1974). Our temperatures are much in excess of these temperatures (300-500°C). More recently however, it has been shown that there also is a pressure effect, as water is driven from the surface of pores and cracks into the adjacent crystal structures with increases pressure (Tullis, 1979). Our pore pressures 50 and 100 MPa are small compared to those (1000 and 1500 MPa) needed in the laboratory to recognize this effect (Tullis and Yund, 1978). Thus, although our temperatures are more than sufficient, our pressures may be too low to foster the ductile aspect of water weakening (item c, above). Coupling this view with the brittle response and macroscopic appearance of the specimens, we tentatively conclude that the water weakening of the andesite is a stress corrosion and/or reduced melting-temperature effect.

Borehole stability for wet andesite and granodiorite is estimated using borehole stress loci for impervious, saturated rocks and the assumptions that σ_h/σ_v is 0.5 and 1.0 and that the effective pressure in the rock is zero (Figures 27 and 28). As noted above, the ratio of 0.5 presents a "worst case" condition and involves the development of tangential tensile stress. Failure in tension with the development of vertical fractures oriented radially to the borehole is a distinct possibility under these conditions at a few kilometers depth. Such failure would not cause borehole collapse, but it would lead to the loss of drilling fluids.

Figure 27. Assessment of borehole stability for water-wet Mt. Hood Andesite at zero effective pressure. Temperatures correspond to data points located to the left. Deviation from a linear 3:1 trend is due solely to rounding strength data to the nearest 5 MPa. Note three borehole stress conditions and corresponding depth scales are given.

Figure 28. Assessment of borehole stability in water-wet Charcoal Granodiorite at zero effective pressure. Details are the same as for Figure 27.

Below we ignore this type of failure, and assume that the rocks will fail in compressive shear fractures according to the differential stress-mean stress relationship. The σ_h/σ_v ratio of 1.0, and a corresponding open hole condition, also is shown in Figures 27 and 28 to permit comparison with data for the dry rock in that we find water-weakening of the andesite is appreciable at temperatures above 800°C (Figure 6).

Data points for both rocks lie very close to, but still to the left of the worst-case borehole-stress loci (Figures 27 and 28). These results suggest, conservatively, that borehole stability would require smooth borehole walls no hotter than 600°C. At higher temperatures the data are too close to the borehole-stress loci to predict stability. Indeed at 850-900°C failure is likely at depths ≤ 1 km for the andesite, and at ≤ 3 km for the granodiorite if at 900 to 1000°C. On the other hand, even though the andesite is weakened by the water, it would appear that it would sustain stable boreholes in open holes to depths ≥ 10 km and temperatures to 880°C if the effective pressure were zero and $\sigma_h = \sigma_v$.

Time Effects. Thus far we have dealt solely with instantaneous compressive strengths of the crystalline rocks. Corresponding predictions of borehole stability involve possible shear failure upon or shortly after penetration by the drill. Such failures could lead to stuck drills and subsequent loss of the hole, and thus are significant to the borehole stability problem. This does not speak, however, to the additional problem of borehole maintenance over the multiyear life of a successful magma-energy-extraction well. It is well known that lowering the strain rate or increasing the duration of loading in a creep test decreases strength (e.g., Carter and Kirby, 1978). The higher is the temperature, the stronger is the time dependence, especially with approach to the melting point.

Only a few studies of time-dependent deformation of crystalline rocks have been made at low pressures and elevated temperatures (see recent review by Carter et al., 1981). There is the unconfined uniaxial-compression creep testing of Murrell and Chakravarty (1973) on crystalline rocks to partial melting, that of Rummel (1969) on a granite at temperatures of 100 to 400°C, and that of Goetze (1971) on Westerly Granite to temperatures of melting at 500 MPa confining pressure and 100 MPa pore-fluid pressure. A compilation of their data and that of Carter et al. (1981) at higher confining pressure is given in Table 10 (Carter et al., 1981, Table 2). In each study the data are empirically well fit by the equation shown in Table 10. Corresponding plots of log creep strain (ϵ_t) versus time (log t, s) for temperatures to 600°C and differential stresses of 10 and 100 MPa (Figure 29, Carter et al., 1981, Figure 13) show that for durations up to say 10 years (life expectancy of a geothermal well) the creep strains are small (10^{-2} to 10^{-4}). While this is an encouraging result relative to the borehole stability problem much work remains systematically to evaluate time-dependent flow at realistic conditions and with other rock types and to evaluate the validity of corresponding constitutive equations.

Table 10. TRANSIENT CREEP PARAMETERS FOR GRANITIC ROCKS [§]

Material	Temperature Range (°C)	Effective Pressure (MPa)	Stress (MPa)	$\epsilon_t = \beta_o \sigma^n t^m \exp(-E/RT)$				Source
				$\beta_o^\dagger$	n	m	E(kJ/mole)	
Granite	20-400	0	87	2×10^{-6}	1.35*	0.25*	6	Rümmel [1969]
Granite	711	400	64	3.9	1.7	0.49	159	Goetze [1971]
Granite	600-800	1000	600-1200	7×10^{-5}	2.2	0.51	128	This study
Granodiorite	830-1045	0	100	3×10^2	1	0.37	176	Murrell and Chakravarty, 1973; Murrell, 1976

§ Copy of Table 2 in Carter et al. (1981)

† β_o is calculated for longitudinal differential stress expressed in MPa and time in seconds.

* These values are of doubtful significance as the material deformed dominantly by logarithmic creep.

Figure 29. Extrapolation in log strain - log time space of power-law high-temperature transient creep equation and parameters given in Table 10. (R) Rummel (1969), (G) Goetze, 1971), (M+C) Murrell and Chakravarty (1973), and (C et al.) Carter et al. (1981). Figure is courtesy of Carter and co-workers.

CONCLUSIONS

The major conclusions to be drawn from the experimental and petro-fabric results now available 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 are 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 deform primarily by fracture, with some ductile flow in the groundmass where it is present, particularly adjacent to faults, 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.

4) 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.

5) 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. (1) Brittle fracture accounts for virtually all the permanent strain. (2) The degree of microfracturing associated with the mechanical loading, as distinguished from the thermoelastic stressing, is about the same at all temperatures from 25° to 1000°C. (3) 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 by fitting their data to an equation of the form of equation 4, 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 claim that a lot more research on fracture must be done before a meaningful constitutive "law" can be written for inelastic deformations in the brittle state.

6) 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.

7) 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 later 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 have 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 have focussed initially

on the instantaneous strengths and ductilities of the rocks with borehole stability predictions pertinent only to relative short periods of time immediately after penetration by the drill. On the other hand, 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 are meager and pertinent testing has just begun. Also we have not considered the influence of thermoelastic stresses that would develop around cooled boreholes and that upon superposition would tend to stabilize the borehole. Predictions must remain qualitative until sophisticated numerical methods can be applied to the problem.

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