THEME 2

1ST INTERNATIONAL CONGRESS ON ROCK MECHANICS

DESCRIPTION OF ROCKS AND ROCK MASSES WITH A VIEW TOWARD THEIR PHYSICAL AND MECHANICAL BEHAVIOR - A GENERAL REPORT

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ABSTRACT

A review of the intrinsic characteristics of crystals, rocks, and rock masses and their physical and mechanical effects leads to an awareness of the parameters that should be considered in any given application of rock mechanics. Evidence is presented to illustrate that composition, structure, and orientation prescribe the behavior of the single crystals. For rocks, it is shown that lithology, statistical crystallographic orientation, grain size, mineralogical alteration, porosity, water saturation, microfracture development, and primary and secondary anisotropies are important. Within the rock mass the surfaces of mechanical discontinuity are decisive, provided that their shear and tensile strengths are less than those of the adjacent rock. Accordingly, the orientation, size, and frequency, degree of mineral filling, shear and tensile strength, water saturation, and contribution to rock mass permeability of the discontinuities, particularly the macrofractures (joints, fissures, cracks), need to be recognized and quantified. Illustrative examples for each point are taken largely from the contributions to Theme 2, First International Congress on Rock Mechanics.

The techniques for measurement and description utilized by the contributors to this Theme are reviewed. They include use of X-ray diffraction, petrographic modal analysis and quality indices, acoustic velocity measurements, a variety of experimental deformation tests, field investigations, drilling tests, and laser-optical processing of photographs.

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INTRODUCTION

Our theme is, in fact, the recognition that rocks and rock masses are not continuous homogeneous, isotropic bodies. Their physical and mechanical** behavior is influenced by interactions between (a) the intrinsic characteristics of the material (Table 1) and (b) the physical and chemical environmental conditions of the deformation (Table 2).*** The intrinsic aspects (crystal structure, composition, and fabric) are manifest on all scales from that of the constituent crystals or grains, through that of the intact of coherent rock sample, to that of the rock mass and terrain. A review of these characteristics and their mechanical effects is attempted here in order to delineate the nature and scope of the parameters that need to be described in applied rock mechanics.

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**All mechanical properties are of course physical, but they constitute a special category relating to the motions of material particles. Some mechanical properties important in rock mechanics are Young's modulus, Poisson's ratio, linear and bulk compressibility, viscosity, and ultimate, yield, shear, tensile, and unconfining compressive strength. There are other physical properties of nonmechanical nature that are of sufficient interest to warrant special attention. These include thermal conductivity and expansion, density, porosity, permeability, electrical conductivity, and acoustic velocity and attenuation.

***The chemical environment doubtless affects the mechanical behavior of rocks, but very little is known about this subject, and it will not be discussed further here.

Table 1

LIST OF DESCRIPTIVE CHARACTERISTICS OF ROCKS AND ROCK MASSES KNOWN TO INFLUENCE DEFORMATIONAL BEHAVIOR AND THE PERTINENT CONTRIBUTIONS TO THEME 2, 1ST. INTERNATIONAL CONGRESS ON ROCK MECHANICS

Scale	Characteristic	Pertinent Contribution This Congress (by Author - alphabetically)
Single Crystal	composition structure - intracrystalline gliding systems orientation	Albissin Habib and Bernaix Pincus
Coherent Rock	lithology	Dreyer, Mendes et al., Paulmann, Ruiz
	preferred crystal orientation	Albissin, Mauriño and Limousin, Pincus, Siemes
	grain size	Boretti-Onyszkiewicz, Dreyer
	mineral alteration	Denissov et al., Hagerman, Hansági, Iliev, Jumikis, Mendes et al., Ruiz
	porosity	Iliev, Kowalski, Ruiz
	degree of saturation	Hansági, Jumikis, Kowalski, Ruiz
	microfractures	Habib and Bernaix, Hagerman, Hansági, Pincus
	primary anisotropy	Jumikis, Kowalski, Pincus
	secondary anisotropy	Boretti-Onyszkiewicz, Mauriño and Limousin, Paulmann
Rock Mass	mechanical discontinuities - (faults, macrofractures [joints], bedding, foliation and schistosity) their orientation and development	Berger, Boretti-Onyszkiewicz, Denissov et al., Habib and Bernaix, Hagerman, Hansági, Jumikis, Mauriño and Limousin, Norris, Paulmann, Pincus, Silveira et al., Sutic and Bozinović
	permeability and water saturation	Jumikis
	shear and tensile strength	Boretti-Onyszkiewicz, Hagerman, Silveira et al.
	size, topography, tectonics	Boretti-Onyszkiewicz, Hansági, Norris, Sutic and Bozinović

Wöhlbier et al. - a statement calling attention to the multidisciplinary approach incorporated in rock mechanics.

Table 2

PHYSICAL ENVIRONMENTAL CONDITIONS

Conditions	Engineering Applications (to depths of 9km)		Geologic-Geophysic Applications (to center of Earth)
Effective pressure	Static:	$\leq$ 2 kb(1)	$\leq$ 3600 kb(2)
	Dynamic (shock):	$\leq$ 10,000 kb(3)	$\leq$ 10,000 kb
Temperature	Static:	$\leq$ 300°C	$\leq$ 3000°C(2)
	Dynamic:	$\leq$ 1500°C(4)	$\leq$ 1500°C(4)
Strain rate	Static:	10 to 10^{-10}/sec	10 to 10^{-16}/sec
	Dynamic:	10^{10} to 10/sec	10^{10} to 10/sec
Differential stress	Static:	$\leq$ 2 kb(?)	$\leq$ 2 kb(?)
	Dynamic:	?	?
Boundary conditions		Rigid to non-rigid to free air	

(1) 1 kilobar (kb) = 10^9 dyne/cm^2 = 987 atm = 14,500 lb/in.^{2} = 1020 kg/cm^2

(2) Katz (1966)

(3) Brode (1964)

(4) Ahrens and Gregson (1964)

This is accompanied by a review of the techniques for measurement and description. The contributions to this Congress are emphasized in the text whenever possible.

ON THE SCALE OF THE CRYSTAL

Composition and Structure

An understanding of why a rock exhibits a given mechanical behavior must begin with consideration of the constituent crystals (Albissin, 1966). Of primary concern are composition and structure and the resulting crystallographic control of certain physical and mechanical properties. Dislocation densities, impurities, and other imperfections are ubiquitous in real crystals and are not treated separately here. As an obvious example, one would expect crystals of halite (NaCl) and quartz (SiO_2) to exhibit radically different mechanical behavior under specified conditions. The abundance of likely slip systems in halite coupled with its high symmetry and low critical resolved shear stress for slip, τ_c, compared to similar considerations for quartz would suggest that halite should be weaker and more ductile than quartz. Indeed, this is the case. It is especially illustrative, however, to examine the differences in behavior that exist for calcite and dolomite because their structures are _not_ radically different and they differ chemically only in that about half of the calcium ions in the calcite structure are replaced by magnesium ions in dolomite (Figure 1). The mechanical differences between the two minerals are large and are manifest primarily in the different operative slip systems (Figure 1) and in the magnitude of τ_c to initiate the respective slip systems (Figures 2 and 3). As will be discussed later, these differences in the behavior of calcite and dolomite crystals are related to equally different ultimate strengths and ductilities between the corresponding rocks.

Orientation

As the physical and mechanical properties of crystals are structurally controlled, the orientation of a crystal with respect to its boundary conditions is critical. This is illustrated by examination of a list of strength data for different directions of loading of a number of different types of single crystals (Handin, 1966). In calcite, for example, the yield strength for crystals compressed parallel to the c axis (r $\{10\bar{1}1\}$ translation gliding favored) is much larger than that for compression normal to the c axis (e $\{01\bar{1}2\}$ twin gliding favored, Figures 4, 5, and 6). Moreover, under low confining pressure this contrast leads to deformation by fracture when the crystal is loaded parallel to the c axis whereas when loaded normal to the c axis the deformation is still by intragranular flow (Friedman, 1963). In addition, it is well known that the values of the elastic moduli, thermal conductivities and expansions, electrical properties, etc., vary with crystallographic direction (Clark, 1966). It follows, therefore, that the physical and mechanical behavior of a rock is strongly dependent upon the nature and degree of any preferred crystallographic orientation of the constituent crystals (e.g., Albissin, 1966; Brace, 1965; among others).

ON THE SCALE OF THE ROCK

Our knowledge on the deformational behavior of rocks is extensive because most experiments are made with the intact or coherent rock specimen. One hesitates to assign dimensions to the size range under consideration here other than to assume that involved are aggregates of crystals or grains which form a continuous body bounded by nonpenetrative planes of mechanical discontinuity. In nature, then "rock" refers to the continuous body found between macrofractures or between bedding and foliation planes which separate materials of different lithology.

Lithology

The term lithology connotes not only gross rock type (e.g., sandstone, limestone, granite, basalt, etc.) but also the minor variations in mineralogy and texture that exist within rock types (e.g., calcite-cemented sandstone, siliceous limestone, anhydrite-rich rock salt, etc.). Handin (1966, p. 235) has compiled a comprehensive list of experimental data on the short-time mechanical properties of rocks. He finds upon study of 174 references that the measurements of different investigators are consistent, and that most specimens within a given lithologic category are found to behave similarly. These categories are listed below in order of decreasing ultimate strength and increasing ductility (Figure 7) as follows:

1) unfoliated igneous and metamorphic rocks, quartzite, highly silica-cemented sandstone,
2) slate and highly indurated siliceous shale,
3) dolomite,
4) moderately well cemented sandstone,
5) limestone,
6) schist, shale, mudstone, and poorly indurated siltstone,
7) salt and gypsum.

Thus it appears possible to discuss the gross influence of lithology without much regard for consideration of grain size, porosity, or texture, particularly for deformations at confining pressures in excess of 1000 bars.

This uniformity of behavior makes possible the prediction of the gross, relative, mechanical properties of any rock whether tested in the laboratory or not. This is particularly true when the relative ductilities or strengths are explicable on the basis of similar differences in mechanical behavior of the corresponding single crystals. For example, the differences in ultimate strength and ductility between limestone and dolomite (Figure 7) may be illustrated by a test specimen composed of a dolomite core surrounded by calcite marble (Figure 8). The more ductile marble as deformed mainly by intragranular flow whereas the dolomite core has remained brittle and has elongated by fracture. Examples of contrasts in ductility are common in the field (e.g., see Kazanskjy and Yakshis, 1964).

Some specific examples of the role of lithology were submitted to this Congress. Dreyer (1966) correlated the volumetric content (modal analysis) of kieserite ($MgSO_4 \cdot H_2O$), anhydrite ($CaSO_4$), and polyhalite [$K_2Ca_2Mg(SO_4)_4 \cdot H_2O$] with the crushing strength for a suite of rock salt samples. He found that crushing strengths increase from 343 to 366 bars for increasing kieserite and anhydrite content (from 0 to 6 percent, each) and that the crushing strengths are independent of polyhalite content in the range 0 to 4 percent. These small differences could only have been recognized by detailed analyses with care taken to keep other variables constant. Mendes, Barros, and Rodrigues (1966) define a quality index K which for granite is directly proportional to the sum of the percentages of quartz, feldspar and micas in the rock and inversely proportional to the sum of the percentages of altered minerals, microfractures, and voids. They show that the index is related statistically to the modulus of elasticity for corresponding samples (Figure 9). Paulmann (1966) in his study of tectonically induced anisotropic behavior found that the experimental results also depended upon the type of rock as shown by his comparisons between sandstone and shale. Ruiz (1966) presents in tabular form the results of seventeen different tests of physical and mechanical properties for some twenty-six different rock types occurring in the state of São Paulo, Brazil. Examination of Ruiz's data for compressive strength versus lithology illustrates a good agreement with Handin's grouping, cited above. These detailed studies clearly illustrate the effects of lithology on the deformational behavior of rocks.

Preferred Orientation of Constituent Crystals or Grains

In the section on crystal orientation it was noted that the physical and mechanical properties for the individual crystal varied with crystallographic direction. It follows that the behavior of a rock is strongly dependent upon any preferred (nonrandom) crystallographic orientation of the constituent crystals or grains. The orientational anisotropy in Yule marble is an excellent example. This marble is characterized by a strong concentration of calcite $\underline{c}$ axes oriented at high angles to a weakly developed macroscopic foliation. Accordingly, compression parallel to or extension normal to the foliation tends to promote $\underline{e}$ twin gliding (the "easy" gliding system) in the individual crystals. For opposite loading orientations $\underline{r}$ translation is favored. As the critical resolved shear stress τ_c for $\underline{e}$ twin gliding is much less than that for $\underline{r}$ translation at low and moderate temperatures, the yield stresses for the crystals are correspondingly affected, and the Yule marble shows the qualitatively predictable strength anisotropy (Figure 4). This anisotropy tends to diminish with increased temperature as τ_c for $\underline{r}$ translation decreases toward that for $\underline{e}$ twin gliding (Figure 2).

Brace (1965) has shown quantitatively that the variation in measured values of linear compressibility in ten rocks are in close agreement with the variation calculated from preferred orientation, modal analyses, and crystal properties. Moreover, measured values of thermal expansion and conductivity for Yule marble are in close agreement with those predicted from the fabric. In these studies Brace applied a weighting factor to calculate the contribution of each crystal to the bulk anisotropy of the rock. Further, Christensen (1965) demonstrated that velocity anisotropy in metamorphic rocks is related to mineral orientation.

Of course, the preferred crystallographic orientation of the constituent grains does not always dominate the physical and mechanical behavior of the rock. Mauriño and Limousin (1966), for example, report that a compressive strength anisotropy related to macrofracture trends and bedding in the rock mass is greater than that related to a preferred orientation of quartz c axes in an orthoquartzite from Argentina.

Grain Size

It is possible to discuss the influence of grain size on mechanical behavior for two broad types of texture. The first is characterized by interlocking crystals and results from initial growth or recrystallization. The texture of metals, igneous and most metamorphic rocks, and some limestones and dolomites is of this type. The second, typical of clastic sedimentary rocks, is characterized by certain packing arrangements and grain-contact configurations.

In the former resistance to deformation tends to increase with decrease in grain size at constant composition. This effect is well known from study of polycrystalline metal (e.g., see Barrett, 1952, pp. 354-356). The crystal boundaries are viewed as an extremely thin transition region where the atomic positions represent compromises between the structural arrangements in the adjoining crystals. Thus the crystal boundaries can be viewed as a wall of dislocations where pinning is readily developed and where impurities tend to segregate. The net effect is to provide a stronger region than within the adjoining crystals. Since the total crystal surface area within an aggregate of given volume increases with decreasing grain size it follows that finer-grained aggregates will exhibit a greater resistance to deformation than the coarser-grained ones.

The contrasts between coarse-grained Yule marble and exceedingly fine-grained Solenhofen limestone illustrate this point nicely. The latter is a gray, massive, pure calcite, lithographic limestone. In some blocks it is essentially isotropic in mechanical behavior and in others it exhibits a low degree of strength anisotropy and preferred crystallographic orientation of constituent grains (Siemes, 1966). The Yule marble is a pure calcite marble with a median grain size between 0.5 and 1.0 mm. It is remarkably homogeneous in the nature and degree of its anisotropy; and is characterized by a marked strength anisotropy which, as mentioned earlier, is related to a strong preferred orientation of the constituent crystals. Further to the similarities between these rocks (other than for grain size), Siemes has shown that changes in crystallographic orientation of the constituent grains in Solenhofen specimens brought about by experimental deformation are essentially identical to those produced in experimentally deformed Yule marble. Some representative strength data for Solenhofen limestone and Yule marble under given sets of experimental conditions obtained from Griggs et al. (1953), Turner et al. (1956), Heard (1960), and Handin (1966) are listed in Table 3. Only data from the <u>strongest orientation</u> for the Yule marble are listed. It is clear

Table 3

SOME REPRESENTATIVE STRESS-STRAIN DATA
FOR SOLENHOFEN LIMESTONE AND YULE MARBLE

Experimental Conditions	Solenhofen Limestone	Yule Marble
All deformed dry, at strain rate of 2×10^{-4}/sec, at 5070 bars confining pressure, and at temperatures as follows:	Differential stress at 10 percent strain (bars)	Differential stress at 10 percent strain (bars)
24°C	7180	4280 (Av. of 4 tests)
150°C	6320	3730
300°C	5150	2500 (Av. of 5 tests)
400-500°C	4300 (400°C)	1980 (Av. of 4, 500°C)

that the finer-grained Solenhofen limestone is much stronger for a given set of physical conditions than the coarse-grained Yule marble. Similarly, rhyolite has a higher ultimate strength than granite, particularly at 1 kilobar confining pressure and 150°C as compared to the smaller difference at 5 kilobars and 500°C (Borg and Handin, 1966).

This point is further illustrated by Dreyer's (1966) detailed analysis of mineralogy, grain size, sorting, and nature of grain boundaries in rock salt specimens. For constant mineralogy, Dreyer found that the crushing strength increased with decreasing grain size (Figure 10), and to a much less pronounced extent with improved sorting. The influence of grain interlocking was negligible. Thus in Dreyer's study there is a continuous increase in strength with decreasing grain size.

In sandstones the role of grain size is less well documented. Boretti-Onyszkiewicz (1966), for example, in study of the compressive and bending strengths of flysch sandstones from the Podhale region of Poland found that the finer-grained sandstones were stronger than the coarser-grained ones. Borg et al. (1960) in study of experimentally deformed St. Peter sand aggregates of specific size fractions (105-125, 180-210, and 250-300 microns, respectively) found that the compressibility of the aggregates tends to increase with increasing grain size. This correlates with the fact that the coarser sand also exhibits the greatest reduction in porosity, median pore size, and median grain size, and the least percentage of unbroken grains. The coarsest sand, however, is the strongest in triaxial tests. In interpreting this result one must remember that the grain size configuration is materially altered by the confining pressure which is applied before the material is axially loaded. Relating the results to initial grain size, therefore, is at best tenuous. More work needs to be done to clarify the role of grain size in sandstones. That the strength should increase with decreasing grain size in sands (as it appears to do in aggregates with interlocking texture) follows from this view expressed by Borg et al. (Ibid, p. 185) namely: "If it is supposed that all size fractions are packed in about the same manner and that there are point contacts among the grains...the pressure applied to the specimens is borne by fewer grains [in the coarser sands] so that the stress concentrations are greater, and rupture is easier."

Mineral Alteration

This topic embraces the effects on mechanical behavior produced by subaerial weathering and the circulation of ground waters. The mineralogical and textural modifications that result usually degrade the properties of the unaltered rock. For example, Iliev (1966) studied changes in certain physico-mechanical properties as a function of the degree of weathering in a suite of monzonitic rocks from the Vitosha pluton, Bulgaria. He found that the modulus of elasticity, the acoustic velocity, and the ultimate strengths were greatly reduced by weathering while porosity was increased (from 2.3 to 12.0 percent). Similar changes in the modulus of elasticity were recorded by Serafim (1964) and by Mendes, Barros, and Rodrigues (1966) whose quality index is inversely proportional to the percentage of altered minerals in the rock (Figure 9).

As for characterizing the degree of weathering, Iliev (1966) proposed a coefficient K based on the acoustic velocity in the unweathered rock (V_o) as compared to that in the weathered rock (V_w), where $K = (V_o - V_w)/V_o$. This same point is expressed in somewhat different terms by Kitsunezaki (1965). Iliev goes on to show that the acoustic velocity decreases linearly with increasing porosity. This would agree with Hamrol (1961), Serafim (1964), and Ruiz (1966) among others who regard porosity as a suitable basis for an alteration index especially when the rock in the unweathered state exhibits low porosity. As another approach, Denissov et al. (1966) suggest that drilling rates are related to the degree of weathering.

A specific example of mineralogical alteration influencing rock behavior is provided by study of the secondary phyllosilicate content (mica, chlorite, pyrophyllite, and chloritoid minerals) in Witwatersrand quartzites (Black, 1962). As the phyllosilicate content increases in the rocks the violence of fracture, linear compressive strength and Young's modulus decrease.

The engineering aspects of recognizing the changes in rock properties brought about by weathering and alteration cannot be minimized. In this Congress, for example, Jumikis (1966) points out how engineers must plan the methods of excavation and the location and design of highways and foundations with regard to the degree of weathering in the Brunswick shale (Triassic) of New Jersey. Indeed, Hagerman (1966) and Hansági (1966) as well as others attempting to classify rocks for mechanical purposes find it necessary to consider the degree of weathering and alteration and the resulting firmness of the rock.

Porosity

In the previous section on alteration porosity was considered as a suitable base for an alteration index for rocks which in the unweathered state exhibit low porosity. In this section porosity is discussed as it influences the mechanical behavior of initially porous materials, primarily the sedimentary rocks. One of the best illustrations known to this reporter is the contribution to this Congress by W. C. Kowalski (1966). He correlated the compressive strengths of air-dried limestones and marls from Poland with their void ratios (volume of pores/volume of solids). These ratios equally could have been expressed as porosity [volume of pores/(volume of solids + volume of voids) × 100]. The rock samples tested were free of fractures and symptoms of weathering changes. The data (reproduced in Figure 11) conclusively show that the compressive strength decreases with increasing void ratio or porosity. Kowalski states the general relationship between strength R and void ratio ξ as $R = d\xi^{-c}$, where the value of the constants d and c depend on the kind of rock, its water saturation, kind of strength (compressive, shear, tensile), and the direction of loading with respect to bedding or other possible anisotropies.

Mogi (1965) has published the results of triaxial compression tests on ten dry rocks including porous andesites, tuffs, and trachytes. He concludes that the mechanical properties of the rocks under confining pressure depend mainly on mineral composition and porosity. In general the yield strengths, ultimate strengths, and Young's modulus decrease and the macroscopic ductility increases with increasing porosity. Further, in the porous rocks Young's modulus increases dramatically with increasing confining pressure, presumably because the porosity is reduced by compaction. For general relationships the reader is referred to Brown et al. (1964) who have established a strength-porosity relation involving different pore geometry and orientation.

Water Saturation

Closely allied to porosity is the role of fluid saturation. Two aspects of the subject need to be separated (a) the effects of the pore fluid pressure and (b) the physicochemical effects involved in altering the minerals of the rock so as to change the gross mechanical behavior of the aggregate. The role of pore fluid pressure and the concept of effective pressure have been described for rocks by Handin et al. (1963) and Robinson (1959) among others. The ultimate strength and ductility of porous rocks depend on the effective confining pressure (the difference between the external confining pressure and the pore pressure) when the pore fluid is chemically inert and the permeability and pore space configuration permit pervasive uniform pore pressure throughout the solid framework. This is illustrated for Berea sandstone, Mississippian, Ohio, in Figure 12. The fluid pressure in macrofractures (joints) also influences the behavior of the rock mass as will be discussed in a later section.

The second aspect of fluid saturation is discussed by Hansági (1966), Jumikis (1966), Kowalski (1966), and Ruiz (1966) in their contributions to this Congress. These authors generally agree that rock strengths decrease with increasing degree of fluid saturation. Jumikis (1966), for example, found the ultimate unconfined compressive strength of dry, weathered Brunswick shale (Triassic, New Jersey) averaged 451 bars as compared to 16-64 bars for saturated specimens. As these were unconfined tests the reduction in strength is probably caused by physicochemical mineral modification and not because of pore fluid pressure effects.* The nature of these changes that result from water saturation are poorly understood, but their influence on mechanical behavior is important and predictably such as to degrade the structural competence of the materials.

Microfractures

Habib and Bernaix (1966) have compiled a comprehensive review of the role of microfractures (fissures, cracks, minute joints) on the deformational behavior of rocks. They point out that fracturing in the hand specimen as well as in the rock mass differs only in scale, and that fractures influence a number of properties on all scales, namely: anisotropy, water saturation, intrinsic strength, mode of rupture, compressibility, Poisson's ratio, acoustic velocity, change in acoustic velocity with pressure, change in permeability with pressure, and variations in the results of rock mechanics tests between

*In saturated rocks of low permeability, however, high pore fluid pressures can prevail for short durations even in unconfined tests because of the relatively slow pressure equalization throughout the fluid phase.

samples and with sample size. Indeed, Habib and Bernaix hold that rocks are different in their behavior from other solids in that they are fractured. In agreement, Mendes et al. (1966) define their rock quality index as inversely proportional to the percent of microfracturing (Figure 7). Hagerman (1966) and Hansági (1966) also subscribe to the view that the mechanical properties of rocks and their state of anisotropy are mainly brought about by the existence of fractures. The relative importance of the microfractures, however, probably decreases with increasing confining pressure.

Primary Anisotropy

Except for consideration of preferred crystallographic orientations the factors enumerated thus far characterize both isotropic and anisotropic rocks. Morphological anisotropy and the differential mechanical behavior that results are usually caused by the nonrandom spatial and distributional character of certain fabric elements. Some of these are primary (related to the formation and diagenetic and metamorphic history of the rock) and some are secondary (related to the tectonic history). The former are discussed in this section and the latter in the following one, all within the framework of the coherent rock rather than the rock mass.

Anisotropy caused by primary elements of the rock fabric are apt to be most important in the sedimentary and in some metamorphic rocks. Aside from preferred crystallographic orientations that can result from growth or recrystallization, sedimentary and some metamorphic processes commonly produce preferred dimensional orientations of crystals or grains and variations in particle size and density which macroscopically are manifest in pervasive bedding, foliation, schistosity, and cleavage. These frequently are planes of low shear and tensile strength. The resulting strength anisotropies are detected by loading specimens parallel, normal, and at various angles to the layering, and are usually most obvious in tests conducted at low confining pressure.

Jumikis (1966) has amply pointed out some of the engineering difficulties that arise from dealing with highly anisotropic rocks. Specifically, he comments on the low shearing strength parallel to bedding and the differential expansion (swelling) observed in water saturated zones of the Triassic red shales of New Jersey.

Secondary Anisotropy

Of concern here are anisotropies introduced by the geologic deformation of the rocks. Habib and Bernaix (1966) call specific attention to fractures and the role they play in creating anisotropic behavior. Borg et al. (1960) and Friedman (1963) in studies of experimentally deformed sands and sandstones recognized that although the stresses in an aggregate are transmitted from one grain to another at points of grain contacts, microfractures, for example, that develop in the grains statistically do not radiate from these contacts but tend to be oriented such as to reflect the principal stress directions across the boundaries of the aggregate as a whole. Thus in nature it seems that deformation can produce or modify microscopic fabric elements in an ordered manner such as to create intrinsic lines or planes of anisotropy that could influence the subsequent mechanical behavior of the material.

Boretti-Onyszkiewicz (1966), Mauriño and Limousin (1966), and Paulmann (1966) in their contributions to this Congress describe three cases of tectonically induced anisotropy. Their reports are similar in scope even though their samples were taken from widely different geologic situations and different testing techniques were employed. Boretti-Onyszkiewicz (1966) describes a situation in the Podhale region of Poland where open macrofractures occur oriented parallel to older calcite filled macrofractures. The fracture sets are oriented at high angles to bedding in sandstones. Sample blocks were taken from unfractured material between the macrofractures, and were experimentally deformed in unconfined compression tests such as to promote fractures along potential planes inclined at high angles to bedding but free to occur at any azimuth with respect to the bedding. The author reports that the induced fractures trend parallel to the main sets of macrofractures developed in the field (Figure 13). Mauriño and Limousin (1966) found from loading orthoquartzite specimens normal to the bedding that the induced fractures occurred parallel to the regional system of macrofractures. Also rupture strengths for these directions are lower than for other directions of test loading. Paulmann (1966) applied unconfined "ball" and "needle" tests to a suite of sandstone specimens and recorded breaking strengths for different directions. He found directions within the rocks for which the breaking strengths were considerably lower than in others. These directions are oriented parallel to the tectonic anisotropy in the rock mass and depend upon the strike and dip of folds, faults, and macrofractures. Highest values of breaking strength were found in specimens taken from zones of anticlines, synclines, and from the vicinity of faults, that is from zones of high strain.

These examples strongly suggest that pervasive geologic deformation can produce or modify microscopic fabric elements that in turn control the observed anisotropic behavior. Unfortunately, the authors did not present microscopic observations so that a specific statement on this point is not possible. On the other hand, one cannot ignore the hypothesis implied by Boretti-Onyszkiewicz that the observed anisotropic behavior is related to the state of stored elastic strain (residual elastic strain) that was "locked" in the rocks during the geologic deformation. Friedman (1966), for example, has shown that such strains can date from at least the Laramide period of deformation (Late Cretaceous). These strains are not necessarily manifest in observable fabric elements, but are a source of internal energy that is capable of influencing mechanical behavior (Emery, 1964).

ON THE SCALE OF THE ROCK MASS

General Statement

More than half of the contributions to this theme are concerned with the descriptive aspects of the rock mass (Table 1). This reflects in part on the success of those who have argued that the major problems in rock mechanics deal with the physical and mechanical behavior of the rock mass rather than that of the intact rock specimen alone (Denkhaus, 1965; Judd, 1964, 1965; Lang, 1964; Müller, 1964; Serafim, 1964; Talobre, 1964; and Zienkiewicz, 1965) to name a few outside the contributions to this theme. This is not to say that one can begin to solve the problems of the rock mass without some insight into the behavior of the coherent material.

What do we mean by the rock mass? In general it consists of one or more lithologic units that form a structure framework. The framework can be undeformed or a highly faulted and folded terrain (Sutić and Božinović, 1966). The detailed configuration of the mass is defined by the size and spatial distribution of the lithologic units and the orientation of the bedding, foliation, and schistosity surfaces. Macrofractures (joints, fissures, cracks, etc.) generally pervade the mass and are commonly developed in one or more sets each composed of a large number of roughly parallel individual fractures. The properties of the fractures, width, spacing, surface areas, roughness, degree of mineral filling, and the nature and magnitude of initial displacements are all highly variable. Thus the typical rock mass is a discontinuous, compositionally, and structurally complex body that is homogeneous only within relatively small domains.

It follows that the deformation of the rock mass will not be uniform and homogeneous. To date an adequate theory to deal with the required complexities does not exist. However, Steketee (1958), Chinnery (1966), and Young

(1966) have shown how dislocation theory may be applied to analytical solutions for the deformation of the rock mass. In this deformation displacements are thought to be restricted to narrow zones (to the surfaces of the mechanical discontinuity). Young states that "The derivative of strain along a shear zone defines a strain gradient which is physically and mathematically equivalent to a continuous distribution of infinitesimal dislocations. This gradient gives rise to a stress field that is identical to that caused by the dislocation distribution and that is readily calculated by applying dislocation theory." This encouraging start should stimulate further treatment along these lines.

Most workers agree that the important factors in the deformation of the rock mass that must be recognized and quantified include the following:

1) the surface discontinuities presented by macrofractures, faults, bedding, foliation, and schistosity--their orientation and development, degree of mineral filling, water saturation, and contribution to rock mass permeability;

2) the contrasts in shear and tensile strength between the representative coherent rock sample and the surfaces of mechanical discontinuity; and

3) the size of the rock mass and its topographic and tectonic setting.

Orientation of the Discontinuities

In the main the rock mass deforms primarily by displacement along the planes or surfaces of mechanical discontinuity, i.e., bedding, foliation, schistosity, and above all macrofractures. As Hagerman (1966) points out this is particularly true in rock masses composed of strong rocks wherein the surface discontinuities are the decisive factors. In weak rocks, on the other hand, the physical properties of the intact specimens are dominant. In the former case, therefore, any prediction of the deformational behavior of the rock mass must be based in part on a description of the orientation of the discontinuities. Bedding, foliation, and schistosity (including slaty cleavage) are reasonably well defined features and on the scale of the rock mass they are pervasive. Accordingly the techniques to map their orientations and describe their thicknesses and surface characteristics are well established. Once the structure is known, and if it is not highly variable, predictions of the attitude and character of these features at any given location within the mass are dependable. On the other hand, the morphology of macrofractures (joints, fissures, cracks, small faults) are highly variable, and the approaches to mapping and describing them also differ. As a result the accuracy of predictions of macrofracture development within a structural framework are either inherently inadequate or are highly subjective. The contributions by Berger (1966) and Silveira et al. (1966) are noteworthy here as they include techniques for the measurement of macrofractures with a view toward obtaining reliable, reproducible, quantitative information with a minimum of personal bias.

Examination of the macrofracture data in the contributions to this theme by Berger, Boretti-Onyszkiewicz, Denissov et al., Mauriño and Limousin, Norris, Paulmann, Pincus, and Silveira et al., shows certain consistencies even though the data are representative of terrains of widely different composition, topography, and structure. Although the macrofractures vary erratically in spacing and number they tend to be developed at any one locality in a reasonably small number of well-defined sets. In the main these sets occur at high angles to the local bedding, foliation, or schistosity and appear to be geometrically and possibly genetically related to the local or regional structure. It follows that if this geometric and genetic relationship can be recognized and tested, one can develop a suitable basis upon which to predict statistically the orientation of the macrofractures in particular domains within the rock mass.

The fact that joints, fissures, cracks, etc., are either shear fractures or extension (tensile) fractures and that they form with a predicable geometric relationship to the principal stresses in the rocks at the time of fracturing (Figure 14) has proved to be a useful basis for relating fractures to structure (e.g., see DeSitter, 1956, p. 132; Stearns, 1964; and Friedman, 1964). Macrofractures sets in the rock mass are recognized as shear or extension features on the basis of this geometry and if possible by offset criteria. They are correlated geometrically with the local structure (fabric axes and planes, fold axes, faults) or genetically through consideration of the corresponding principal stress axes. For example, certain fracture patterns are now recognized to be ubiquitously associated with folds presumably because they are related to the bending stresses in the folded plates (Figure 15). This has been tested by mapping these fracture patterns in genetically different folds. Similarly, in and adjacent to fault zones macrofractures parallel and conjugate to the fault are invariably developed. It follows that macrofracture orientation patterns can be predicted from stress trajectory information obtained from theoretical analyses or from photomechanical model studies of rock mass problems.

The major limitation to this approach is that one cannot predict what combination of shear and extension fractures will form from a given orientation of the principal stress axes. For example, extension fractures alone might form, or one shear and an extension fracture, or both shears, or only one of the shear fractures etc. Accordingly, predictions of fracture orientation from one locality to another can only be stated ideally in terms of the total shear and extension fracture assemblage possible for a given state of stress.

Norris (1966) related the bedding, macrofractures, cleats, normal faults, reverse faults, and wrench faults to the kinematic axes for thrust plates in his fabric study about some Canadian coal mines. He found that the structural fabric does not vary significantly and that the macrofractures and bedding are the most consistent and predictable fabric elements. The macrofractures can be correlated between some localities by their similar orientations with respect to the local fabric axes. In other cases sets recognized at one stratigraphic level cannot be correlated, one for one, with those at another level, even though at both stations the same stress orientations are deduced from the extant fracture geometry and offset criteria. This is a good illustration of the limitation stated above.

Another example of the relationship between macrofracture and structure is afforded by Mauriño and Limousin (1966). They studied macrofractures in Silurian orthoquartzites, Province of Buenos Aires, Argentina, in a region characterized by basement block faulting. A well-defined macrofracture pattern is developed in the region consisting of two perpendicular sets. Locally a third set is developed parallel to the trend of the major block faults. Boretti-Onyszkiewicz (1966) reports a similar situation in sandstones from the Podhale region of Poland, where a regional orthogonal system of macrofracture sets is developed apparently independently of local fold axes. Locally, however, the macrofractures and faults are parallel. In some cases the fault displacement is manifest by movement along one of the sets.

These studies serve to illustrate that macrofractures do not necessarily randomly pervade the rock mass. They tend to be developed in readily mapped sets and are often geometrically if not genetically related to local and regional structure. This suggests that macrofracture orientations can be predicted for a given unexplored locality from projected structure or from stress trajectories obtained through theoretical and model studies.

Development of Macrofractures

The lengths, widths, surface areas, and the abundance of macrofractures are important from a mechanical viewpoint and are amenable to quantitative description (see e.g., Silveira et al., 1966). The genetic factors controlling the first three of these characteristics are too poorly understood to mention them further. Several generalities can be made on abundance, however, as determined from a review of the comments by our theme contributors. The abundance of macrofractures:

a) increases in the vicinity of faults and other areas of high strain,

b) increases with decreasing bedding thickness, other variables being equal,

c) is greater in brittle as compared to ductile rocks, and

d) increases in localities that are near the air-rock interface and particularly near steep topographic slopes.

One measure of abundance is spacing. This applies once the macrofracture array has been characterized into sets. For each set the spacing is the mean distance between individual fractures as measured along a direction perpendicular to the fracture surface. Another measure is the number of fractures per linear distance along this same direction (Stearns, 1964). Still another locally useful measure is the number of macrofractures that intersect a surface of a given orientation and area (Silveira et al., 1966).

Permeability and Water Saturation

The discontinuities in the rock mass, particularly the macrofractures, provide permeability channels for the circulation of ground waters. Chemical alteration of wall rock and fracture filling are enhanced and the strength of the rock mass may be correspondingly decreased to a degree dependent upon the permeability and the time available for alteration. Serafim (1964) has emphasized the importance of studying permeability trends related to the fracture pattern to determine if grouting curtains are needed and what their design might be to most effectively stabilize the rock mass. He also points out, as does Jumikis (1966) and Secor (1965) that the hydraulic pressure within the fractures influences the behavior of the rock mass just as it does in the pores of the coherent rock sample. Jumikis mentions that noncommunicating ground water pressure can be higher than the normal hydrostatic head and cause lifting up and breaking off of shale slabs. The noncommunication is sometimes caused by clay-fillings in the macrofractures and by local faults which act as barriers. The need for proper drainage should be recognized.

An illustration of the importance of water pressure in macrofractures is afforded by accounts of earthquake activity near Denver, Colorado, which was presumably stimulated by pressure injection of waste water in a disposal well (Evans, 1966; and Bardwell, 1966). The well is bottomed in fractured Precambrian gneiss at a total depth of 12,045 feet. Since the fluid injection began, 710 earthquakes have been recorded (magnitude range: 0.7 to 4.3 on the Richter scale), which have epicenters within a 5 mile radius of the well. The statistical correlation between volume and pressure of the fluid injection and the seismic activity is most convincing. Evans suggests that the rock movements are due to the increase of fluid pressure within the macrofractures. Other interesting aspects of this situation are (1) the suggestion that the rock mass appears to be in a rather delicate state of rest such that fluid injected at wellhead pressures up to 1050 psi cause the rock movements; (2) there is evidence that the displacements causing the seismic activity are occurring along a plane (fault) that dips eastward and passes beneath the disposal well at a depth of 6.5 miles (Wang, 1965); and (3) that macrofractures open to fluids can exist to a depth of at least 12,050 feet and, if point (2) is valid, to as much as 6.5 miles!

Shear and Tensile Strengths

If the rock mass prefers to deform by displacements along the discontinuities, the shear and (or) tensile strengths of the discontinuities must be less than those for initiation of new failure surfaces in the coherent rock. This fact led Hagerman (1966), Coates (1964), and others to base their classification in part on the uniaxial compressive strength of the coherent specimen. Hagerman, in view of local conditions, reports that the discontinuities, mainly macrofractures, are decisive in the deformation of the rock mass only for rocks with compressive strengths greater than 980 bars.

Perhaps a more direct approach is illustrated by the work of Lane and Heck (1964) and Byerlee (1966). They studied the frictional characteristics along natural macrofractures in triaxial tests, and compared the results to the mechanical behavior of the corresponding intact samples. For a given angle between the surface discontinuity and the axial load, the shear stresses for sliding along the surface at different normal stresses are recorded. An envelope for sliding along that surface is obtained and compared to the envelope for the intact sample in order to define fields of failure and stability (Figure 16). Through tests such as these useful engineering data for a given situation can be obtained. In addition they contribute along with Donath (1961), Jaeger (1959), Handin and Stearns (1964), Henkel et al. (1964), Lang (1964), Withers (1964), and others to the general understanding of the factors that influence the coefficient of sliding friction.

Granted the successful completion of these studies, the problem would still remain for the geologist or engineer to estimate the mechanical effects of the discontinuities in the field. Description of the character of the surfaces is a start to at least a qualitative estimate of their past or potential mechanical involvement. Probably significant in this regard are surface roughness (degree of mineral orientation, slickenside development, etc.), ease of parting, area over which a single plane extends, and the degree to which macroscopic asperities interlock. Particularly for filled macrofractures, the nature of the secondary filling, its strength compared to that of the wall rock, and the degree to which it bonds the adjacent blocks, are doubtlessly important. For example, Boretti-Onyszkiewicz (1966) describes one region in which fresh, open, tensile macrofractures that developed after older calcite-filled ones. The conclusion is clear that the tensile strength of the calcite-filled fractures was greater than that of the intact rock.

Size, Topography, and Tectonics

It is a matter of experience that the mechanical behavior of the rock mass is in part dependent upon its size, topographic expression, and tectonic history (e.g., see general comments by Sutić and Božinović, 1966). Size is a factor in at least three areas: (1) Generally, the strength of a rock mass decreases with increasing size primarily because the probability of encountering flaws is greater in the larger body. It is common knowledge, for example, that in a given rock the stability of an excavation or opening increases with decreasing size (Lang, 1964). (2) Body forces, of course, become appreciable in large masses and can be the dominant factor in a given situation. (3) The extrapolation of laboratory test data becomes increasing tenuous with increasing size of the rock mass (Hansági, 1966). This is so not only because of the flaws in the rock mass, but because the simplified boundary conditions of the experimental test become decreasingly realistic with increasing size of the body under consideration (Judd, 1965).

The effects of topography need little statement. It is well known that (a) if other factors are equal, slope stability decreases with increasing profile angle, (b) masses near the air-rock interface exhibit increased fracturing and weathering (Boretti-Onyszkiewicz, 1966), (c) orientation of the topographic exposure relative to climatic factors and to internal structure is an important factor, and (d) the magnitude of topographic relief influences the relative importance of body forces.

Tectonic deformation of the rock mass is probably the broadest and most significant of these three topics. The important relationships between tectonics and the development of macrofractures and other surfaces of mechanical discontinuity have been cited earlier. In addition, at least three other factors are noteworthy, namely: (1) The present in-situ state of stress is the result of the superposition of residual stresses (related to some past tectonic event), current tectonic stresses, and the state of stress caused by the weight of the overburden. Judd (1964), Müller (1964), and Talobre (1964) among others have called attention to the importance of recognizing that the in-situ state of stress is apt to differ from the hydrostatic condition proposed by Heim in 1878. (2) Deformation of rocks up to a certain point produces work hardening which can in certain cases enhance the strength of the material (i.e., the Bauschinger effect). On the other hand, excessive deformation causes a decrease in supporting strength by the development of shearing surfaces as described by Norris (1966). (3) Faulting tends to influence local loading conditions. For example, Parker and Scott (1964) found in the White Pine Mine, Michigan, that pillars outside faulted areas carry more load than those in faulted areas. They attributed this to movement on fault planes which tends to relax stresses in nearby pillars and transfer loads to those in adjacent zones.

TECHNIQUES FOR MEASUREMENT AND DESCRIPTION

The number of observational techniques required in rock mechanics investigations is large because the phenomena of interest range in size over at least fourteen orders of magnitude from the crystal structure (10^{-9} m = 10 Å) to the mountain mass (10^5 m). Standard techniques for study of the crystal to the dislocation level involve use of different etching and decorating methods, optical, electron, and X-ray microscopy, X-ray diffraction, and the electron microprobe. Description of the rock specimen usually includes optical microscopy, petrofabric methods, X-ray diffraction, and chemical techniques. Bulk properties are measured by physical and mechanical tests. The rock mass is commonly described by employing standard field mapping techniques, drilling results, in-situ physical and electrical measurements, and study of aerial photographs. The contributors to Theme 2 make interesting use of many of these techniques, some of unusual nature are mentioned below.

The optical processing of two-dimensional, petrofabric data described by Pincus (1966) is a new approach applicable to the study of phenomena on all scales of observation. The method has been applied previously to the analysis of seismic data. According to the author, a photograph of the subject "is reduced to a transparency that acts as a diffraction grating through which laser light is passed. The frequency distributions of direction and spacing can be taken directly from the resulting two-dimensional Fourier transform." Directional and spatial filtering of the original image is an attractive aspect of the technique. The usefulness of the method is illustrated for calcite twin lamellae spacing and orientation, dimensional grain orientation, and lineations on an aerial photograph.

Seimes (1966) uses X-ray diffractometry to determine the preferred crystallographic orientation in ultra fine-grained Solenhofen limestone. This technique is particularly useful in study of rocks too fine-grained for standard optical analysis.

Attempts to describe rocks quantitatively in terms of indices useful in the systematic exploration of the rock mass are made by several workers.

Mendes and associates (1966) define a quality index based upon detailed microscopic analyses of mineral composition, mineral alteration, fracturing, and voids. They find a good correlation between this index and the modulus of elasticity. Iliev (1966) concludes that acoustic velocity varies significantly with the degree of weathering and that velocity differences between weathered and unweathered counterparts can be used to obtain a coefficient of weathering. This coefficient can in turn be useful in estimating other physical and mechanical properties of rocks that are known to be influenced by the degree of weathering.

Boretti-Onyszkiewicz (1966), Mauriño and Limousin (1966), and Paulmann (1966) use experimental deformation tests to detect latent planes of failure related to tectonically induced anisotropy. Utilization of the rock testing apparatus as a structural tool is an approach likely to see rapid growth. For example, Karp and Donath (1966) experimentally deformed prestrained specimens in an attempt to measure the magnitude of the prestrain and the directions and relative magnitudes of the principal stresses responsible for the initial deformation. Their results are encouraging and suggest that the method may be used in study of naturally prestrained specimens.

Several authors utilize the information obtained during drilling to evaluate the mechanical state of the rocks penetrated by the bit (Denissov et al., 1966; and Hansági, 1966). A core-fissuring factor is defined by Hansági and is used to establish a value of rock-firmness and a practical mining classification for the rock. Denissov and his co-workers use core recovery and drilling speed as parameters. They find that the former can give an incorrect idea about the state of the rock, whereas the latter, drilling rate, yields a certain correlation with the intensity of macrofracturing and the depth of weathering.

The important problem of accurately and quantitatively describing macrofractures in the rock mass is dealt with by Berger (1966) and Silveira et al. (1966). Berger advocates use of sampling grids which enable him to obtain a statistically reproducible and realistic picture of the macrofracture development. Silveira and associates quantify the orientation, area, thickness, and spacing of the macrofractures, and utilize equatorial cylindrical projections as well as upper hemisphere, equal-area projections to illustrate and analyze the macrofracture orientation data.

CONCLUSIONS AND TOPICS FOR GENERAL DISCUSSION

It has been demonstrated that the intrinsic characteristics of crystals, intact rocks, and the rock mass can influence the nature of the physical and mechanical behavior of the material. Factors such as crystal structure and composition, lithology, preferred crystallographic orientation, mineral alteration, fluid saturation, grain size, and rock mass size are probably important throughout the ranges of confining pressure, temperature, and strain rate encountered in nature. On the other hand, porosity, microfractures, macrofractures, residual stresses, and most other primary and secondary anisotropies are most influential in the low confining pressure and low temperature regimes prevailing at shallow depths in the crust. It follows that all of these factors are potential variables to be considered in most engineering applications.

The accurate prediction of the deformational behavior of rocks and rock masses is a great challenge. At present we are at a disadvantage because a general theory to deal with the deformation of real materials does not exist. Moreover, it may not be possible to specify all the interrelationships between the physical environmental conditions and the intrinsic characteristics of the material. Nonetheless, it is essential in applied rock mechanics to be aware of the possible variables that enter into the system, to recognize which are the important factors, and then to undertake whatever measurements and calculations seem necessary at least to improve our understanding of the problems.

It is appropriate to conclude the general report with a list of subjects which reflect the current interests of workers in our field. Accordingly, the following subjects are proposed for discussion, namely:

1) The problems involved in recognizing which of the many possible intrinsic characteristics of rocks and rock masses (in conjunction with the physical environmental factors) are the most significant or critical in governing the deformation behavior of the material.

2) The use of detailed petrographic studies (modal analyses, quality indices, petrofabric analyses) and field studies as a means of predicting at least the relative physical and mechanical properties of rocks and rock masses.

3) The descriptive characteristics of mechanical discontinuities [macrofractures (joints), bedding, foliation, etc.] that tend to influence the physical and mechanical behavior of the rock mass.

4) The extrapolation of quantitative laboratory test data for the small intact rock specimen to problems involving the deformation of the rock mass.

5) The use of the experimental deformation test to detect planes or lines of "easy" potential failure, permanent strain, principal stress axes, and other phenomena that relate to the tectonic history of the sample.

6) The descriptive classification of rocks and rock masses for applied rock mechanics. Is a general classification possible at this time or is a classification practical only when it is designed to fit local conditions and a specific application? In either case what factors should be considered in such a classification?

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List of Captions

Figure Number

1 Diagrammatic representation of the calcite (a) and dolomite (b) structures. In (a) section is drawn normal to the a_2 axis. The structure is twinned on $\underline{e}_1$ $\{01\bar{1}2\}$, with the gliding direction and sense of shear indicated. $\underline{r}_1$ $\{10\bar{1}1\}$ and $\underline{f}_1$ $\{02\bar{2}1\}$ translation planes, gliding directions, and senses of shear are also indicated. In (b) the plane of the section is parallel to a_3 $\{11\bar{2}0\}$. Gliding direction for translation on $\{0001\}$ is not in the plane of the section, but is parallel to any of the three $\underline{a}$ axes. The system for twin gliding parallel to $\underline{f}$ is also illustrated.

2 Critical resolved shear stress (τ_c) for $\underline{e}$ twin gliding and $\underline{r}$ and $\underline{f}$ translation in calcite as functions of temperature at different strain rates (unpublished compilation, H. C. Heard, 1966).

3 Critical resolved shear stress as function of temperature (at strain rate of 1.7×10^{-4}/sec) for basal translation and $\underline{f}$ $\{0\bar{2}\bar{2}1\}$ twinning in dolomite (after Higgs and Handin, 1959, Figure 12).

4 Stress-strain curves for dry Yule marble and calcite single crystals extended or compressed at 10 kilobars confining pressure, strain rate of 2.5×10^{-4}/sec, and at different temperatures. Specimen orientation is indicated for each curve. Each curve represents an average of two or more experiments (after Griggs and Miller, 1951, Figures 3 and 4; and Griggs, et al., 1951, Figures 1 and 2).

5 Stress-strain curves for dolomite single crystals loaded parallel to the $\underline{c}$ axis at 5 kilobars confining pressure, strain rate of 1.7×10^{-4}/sec, and temperatures as noted on each curve (after Higgs and Handin, 1959, Figure 6).

6 Schematic illustration of the calcite structure showing loading conditions favorable for $\underline{e}$ twin gliding (on left) and for $\underline{r}$ translation (on right).

7 Ultimate compressive strengths and ductilities of water-saturated rocks as functions of depth. Effects of confining (overburden) pressure, temperature (30°C/km), and "normal" pore pressure included, strain rate of 1.7×10^{-4}/sec (after Handin, et al., 1963, Figure 27).

8 (a) Photograph of a section through a deformed cylinder consisting of a dolomite core surrounded by calcite marble. Contrasting behavior is shown by marked development of shear fractures in the dolomite and their absence in the marble. (b) Geometry of fractures in the dolomite of (a). Extended 31 percent at 5000 bars confining pressure, strain rate of 1.7×10^{-4}/sec, 300°C (after Griggs and Handin, 1960, Pl. 8).

9 Correlation between the modulus of elasticity and a quality index based upon petrographic analyses for a suite of granite samples. Quality index equals % quartz + % feldspar + % micas + others/% altered mineral + % microfractures + % voids (after Mendes, Barros, and Rodrigues, 1966, Figure 5).

10 Relationship between grain size (decreasing to the right) and crushing strength in rock salt specimens (after Dreyer, 1966, Figure 3).

11 The interdependence between the strength (R_{cs} in kg/cm^2) and void ratios of Polish air-dry limestones and marls (after Kowalski, 1966, Figure 1).

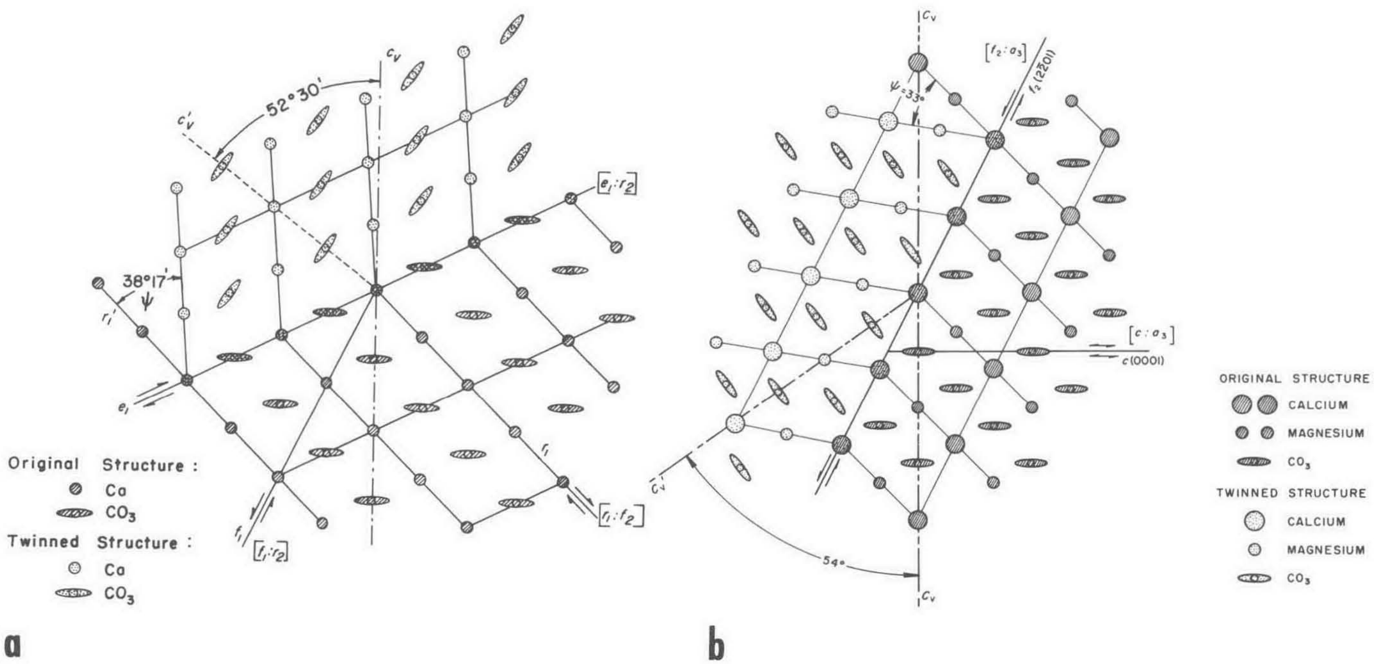
12 Stress-strain curves for Berea sandstone at different pore water pressures (kilobars) and strain rate of 1.7×10^{-4}/sec. Below: all at 2 kilobars confining pressure and 24° or 300°C; all in compression except curve marked Ext (for extension). Above: at confining pressures (p_c) of 0.5, 1.0, and 2.0 kilobars at 24°C; at pore pressures (p_p) of 0, 0.5, and 1.5 kilobars; all at same effective pressure of 0.5 kilobars (after Handin, et al., 1963, Figure 4).

13 Diagram of natural joint directions (solid line) and of fractures induced during laboratory loading (dashed line) of sandstone monoliths from the Podhale region of Poland (after Boretti-Onyszkiewic, 1966, Figure 4).

14 Idealized geometric relationships between the principal stresses ($\sigma_1 > \sigma_2 > \sigma_3$, compressions positive) and the extension (A) and shear fractures (B and C). The senses of displacement are shown. Typical angle between the shears B and C is about 60° in rocks.

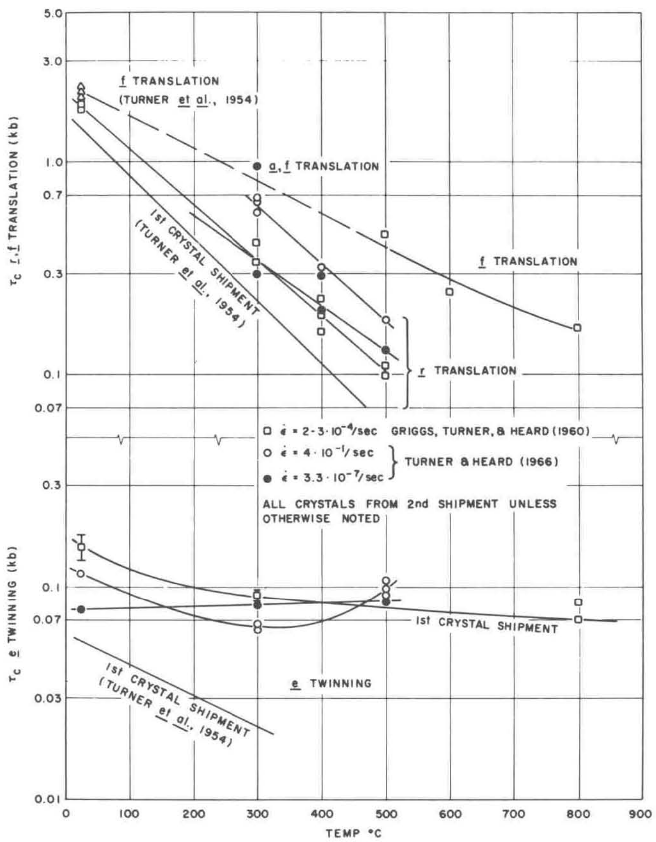
15 Idealized sketch of five fracture patterns found superposed on the Teton anticline, Montana. Pattern 1 is the dominant pattern; sketch grossly distorts prevalence of patterns 2, 3, 4 and 5 (after D. W. Stearns, personal communication, 1964).

16 Mohr envelops for intact cores of medium-grained quartz monzonite and for sliding along natural open fractures inclined at 28° to axial load. Angle ϕ is the friction angle. Specimens with natural fracture are loaded until slip occurs on fracture surface. Each stage in this loading is at a given confining pressure. The second and third stages are at successively higher confining pressures on same specimen. Stage four is obtained by reloading after unloading from stage three (after Lane and Heck, 1964, Figure 8).



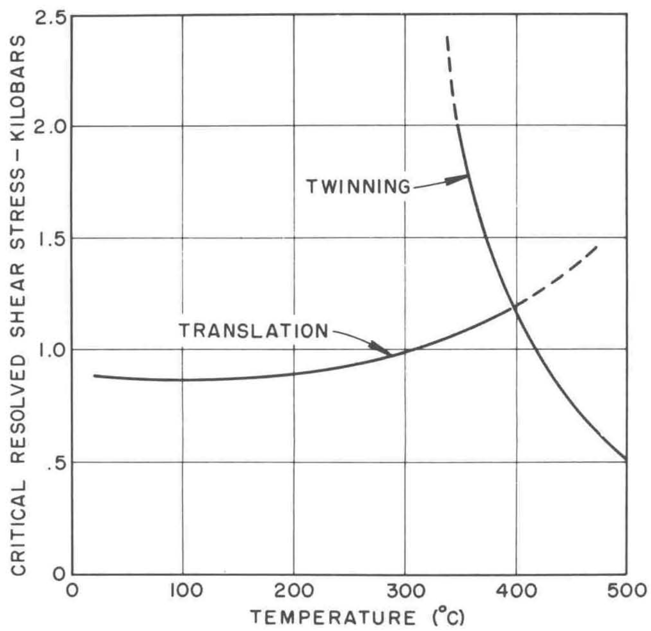
52°30'
c_v
c_v'
$[e_1:r_2]$
38°17'
ψ
r_1'
e_1
f_1
$[f_1:r_2]$
$[r_1:f_2]$
Original Structure :
Ca
CO_3
Twinned Structure :
Ca
CO_3
a
c_v
$[f_2:a_3]$
$f_2(2\bar{2}01)$
$\psi=33°$
$[c:a_3]$
c(0001)
c_v'
54°
ORIGINAL STRUCTURE
CALCIUM
MAGNESIUM
CO_3
TWINNED STRUCTURE
CALCIUM
MAGNESIUM
CO_3
b

Figure 1



f TRANSLATION
(TURNER et al., 1954)
a, f TRANSLATION
1st CRYSTAL SHIPMENT
(TURNER et al., 1954)
f TRANSLATION
r TRANSLATION
□ ε̇ = 2-3·10^-4/sec GRIGGS, TURNER, & HEARD (1960)
○ ε̇ = 4·10^-1/sec
● ε̇ = 3.3·10^-7/sec
TURNER & HEARD (1966)
ALL CRYSTALS FROM 2nd SHIPMENT UNLESS OTHERWISE NOTED
1st CRYSTAL SHIPMENT
e TWINNING
1st CRYSTAL SHIPMENT
(TURNER et al., 1954)
τc r, f TRANSLATION (kb)
τc e TWINNING (kb)
TEMP °C

Figure 2



CRITICAL RESOLVED SHEAR STRESS - KILOBARS
2.5
2.0
1.5
1.0
.5
0
TWINNING
TRANSLATION
0
100
200
300
400
500
TEMPERATURE (°C)

Figure 3

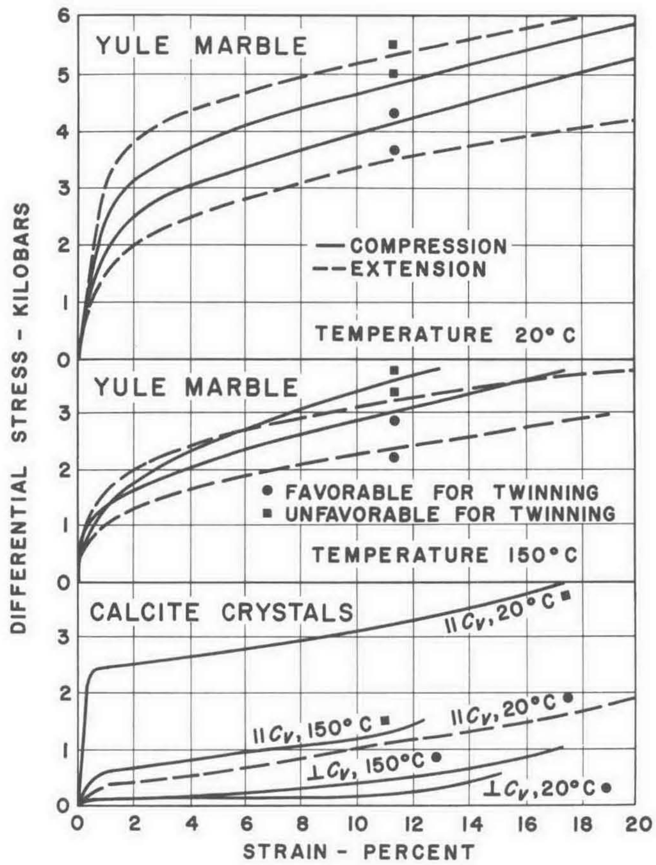

YULE MARBLE
—COMPRESSION
--EXTENSION
TEMPERATURE 20° C
YULE MARBLE
● FAVORABLE FOR TWINNING
■ UNFAVORABLE FOR TWINNING
TEMPERATURE 150° C
CALCITE CRYSTALS
DIFFERENTIAL STRESS - KILOBARS
STRAIN - PERCENT

Figure 4

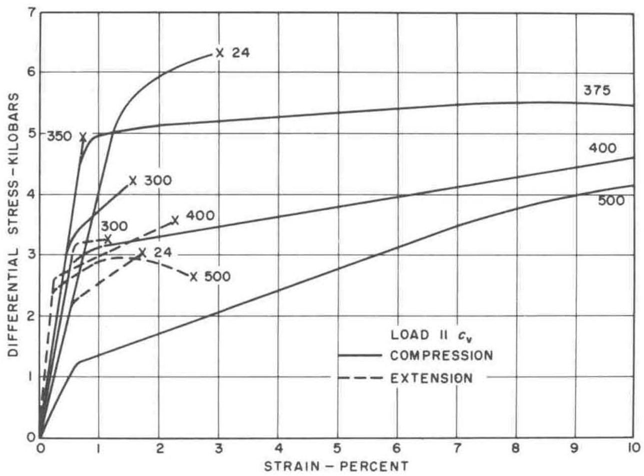

DIFFERENTIAL STRESS - KILOBARS
STRAIN - PERCENT
X 24
375
350 X
X 300
400
500
X 400
300
X 24
X 500
LOAD II c_v
COMPRESSION
EXTENSION

Figure 5

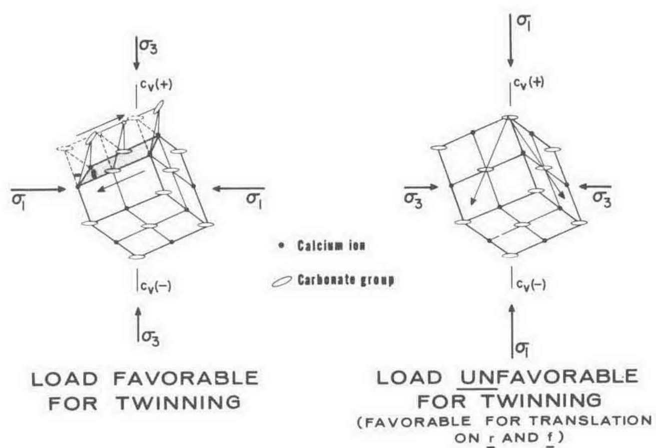

σ3
cv(+)
σ1
σ1
cv(−)
σ3
LOAD FAVORABLE
FOR TWINNING
Calcium ion
Carbonate group
σ1
cv(+)
σ3
σ3
cv(−)
σ1
LOAD UNFAVORABLE
FOR TWINNING
(FAVORABLE FOR TRANSLATION
ON r AND f)

Figure 6

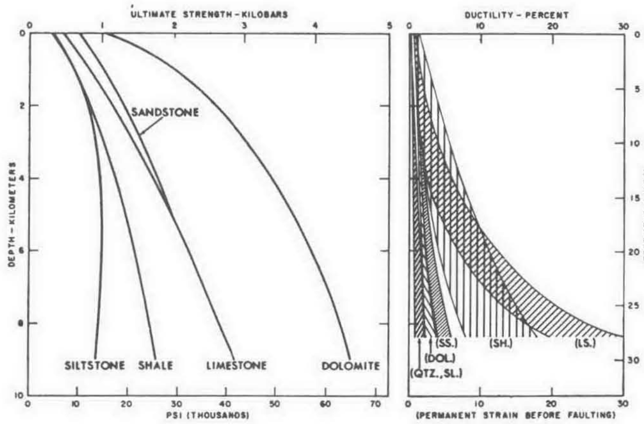

ULTIMATE STRENGTH - KILOBARS
DEPTH - KILOMETERS
SANDSTONE
SILTSTONE
SHALE
LIMESTONE
DOLOMITE
PSI (THOUSANDS)
DUCTILITY - PERCENT
FEET (THOUSANDS)
(SS.)
(SH.)
(LS.)
(DOL.)
(QTZ., SL.)
(PERMANENT STRAIN BEFORE FAULTING)

Figure 7

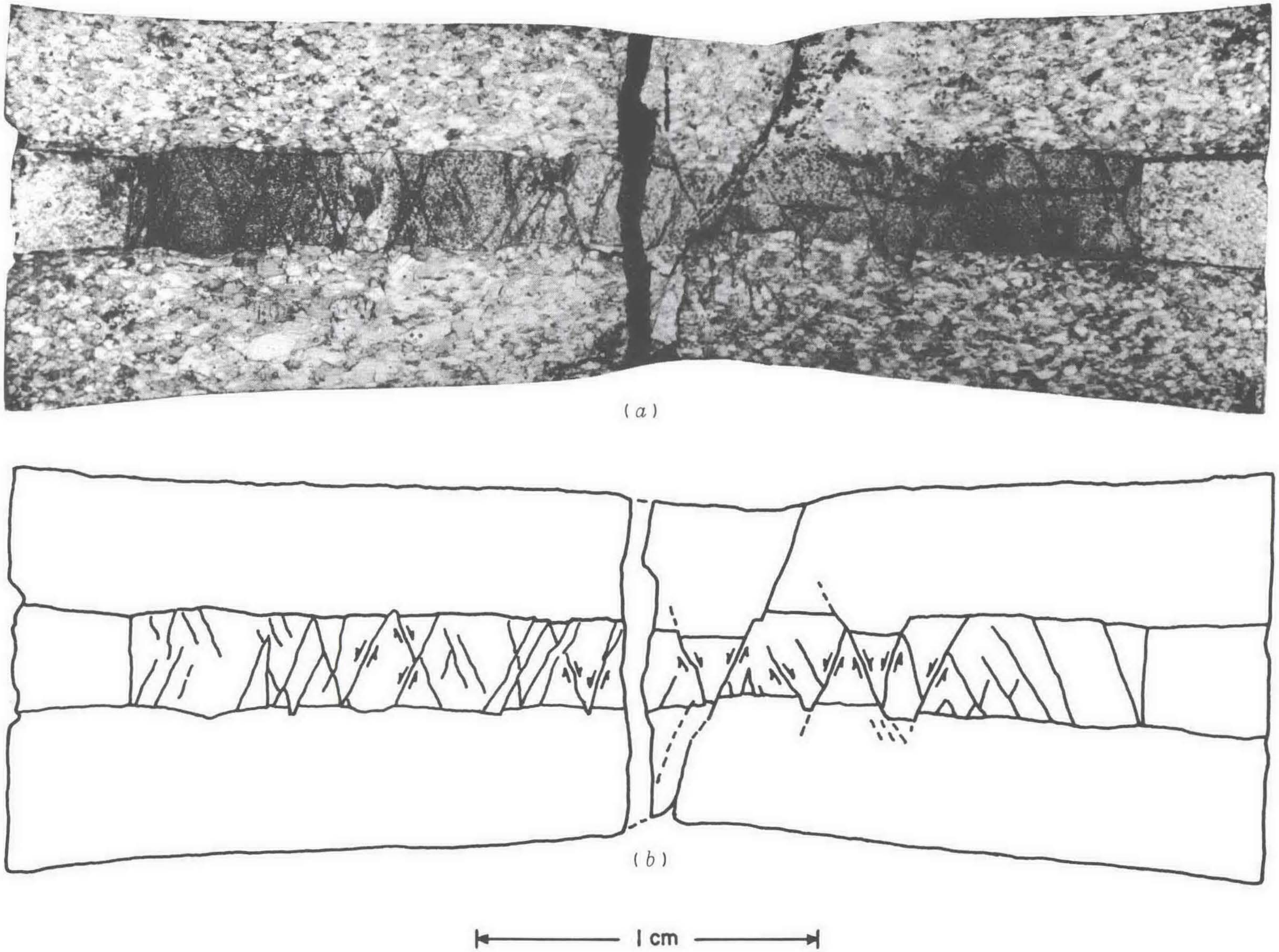


Figure 8

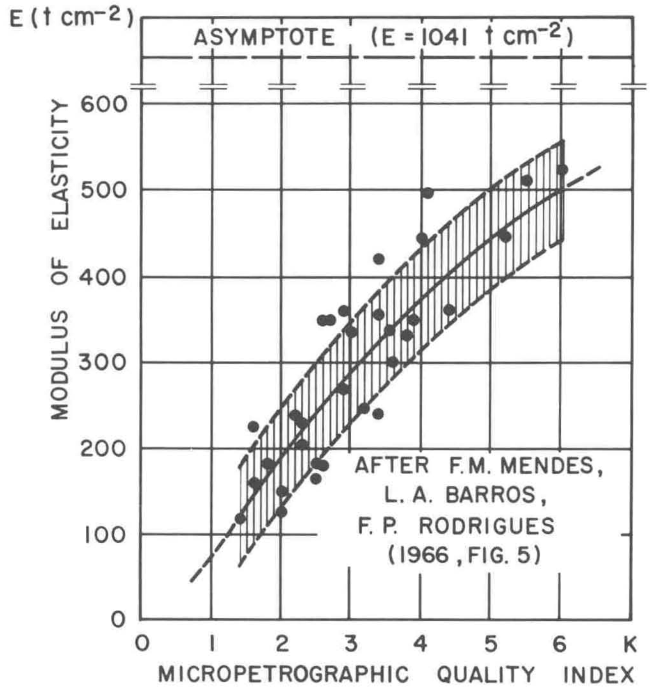


Figure 9

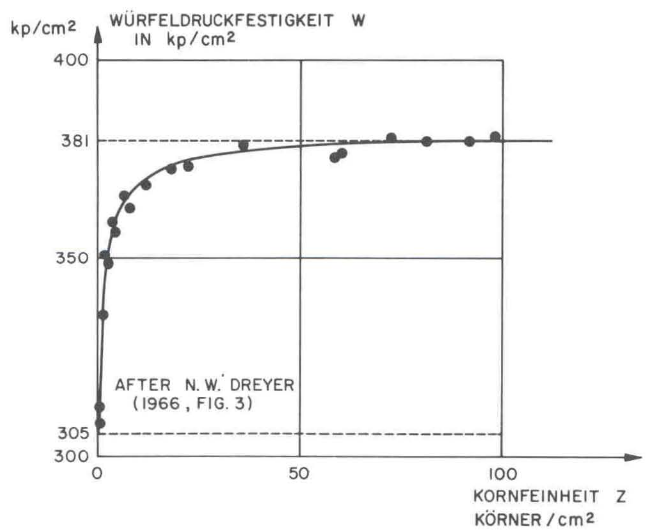


Figure 10

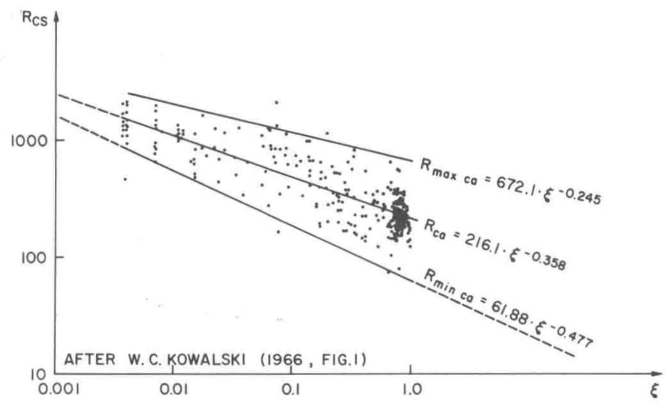


Figure 11

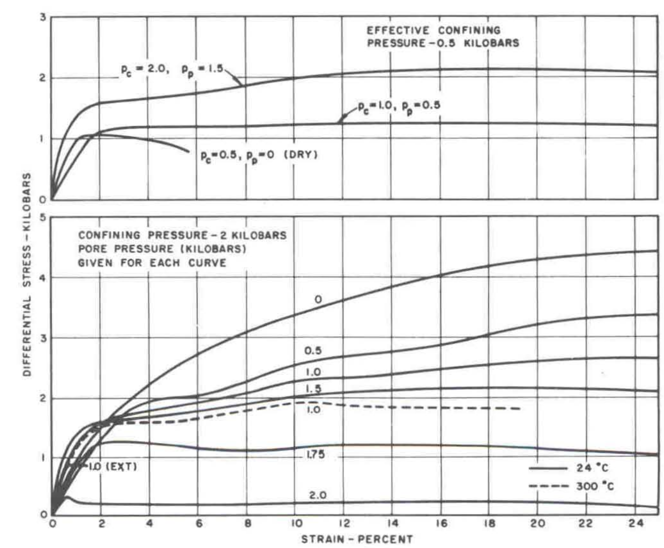


Figure 12

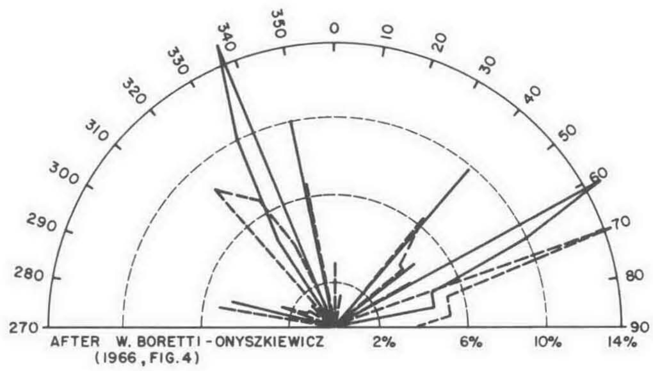


Figure 13

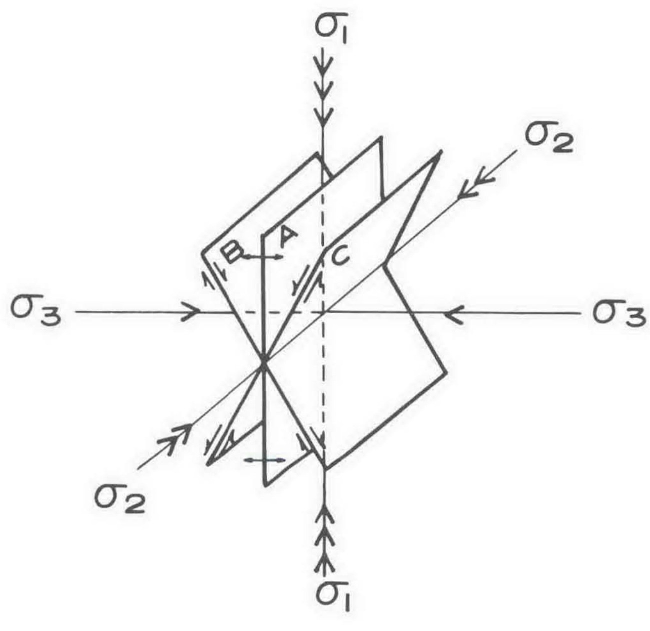


Figure 14

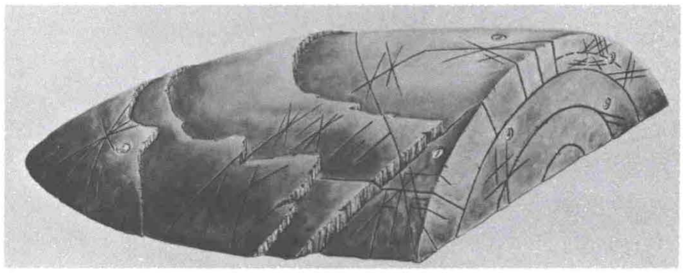


Figure 15

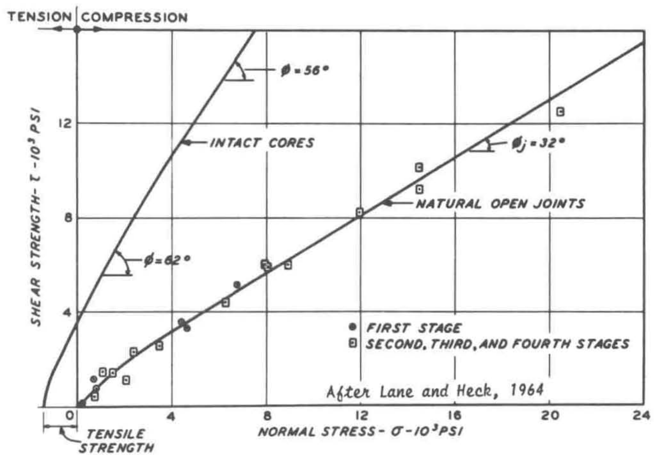


Figure 16

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