Fred A. Donath

Some Information Squeezed Out of Rock

Experimental rock deformation provides a key to the interpretation of dynamic geologic environments

Nearly everyone is aware that a ductile metal such as copper can be hammered or stretched and permanently deformed without fracturing. However, anyone who has ever raised a hammer to a rock knows that if he persists, and the hammer does not break first, the rock will eventually shatter into many pieces. Under ordinary experience rock is brittle. Thus, it comes as a surprise to many to learn that under certain conditions rocks too can be very ductile. Ordinary experience simply is not adequate to predict the behavior of rock under conditions found within the earth. The fact that rock does show marked differences in behavior in natural deformation provides the geologist with an unusual opportunity to learn more about the conditions of deformation—if only he can relate these differences in some systematic manner to environmental and rock factors.

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His primary interest has been the analysis of geological deformation based on the principles of mechanics and on laboratory and field investigations. For a number of years he has carried on an extensive experimental program, first at Columbia University and subsequently at the University of Illinois, to evaluate the effects of anisotropy, pressure, temperature, strain rate, and previous strain history on the deformational behavior of rocks. Address: Department of Geology, University of Illinois, Urbana, Illinois 61801.

From their studies of natural deformation geologists have long appreciated the ability of a rock to fracture in some instances and to flow in others. Moreover, since the work of F. D. Adams in the early 1900s, they have known that the pressure confining the rock has been a principal factor in effecting the change from fracture to flow. Adams deformed rock specimens under confining pressure by placing them in the center of a hole bored through a solid steel cylinder (Fig. 1). The central portion of the steel cylinder was turned down on a lathe so as to leave a thin wall opposite the specimen. By inserting a piston into each end and forcing them into the cylinder against the specimen with a loading press, he caused the specimen to deform and bulge against the restraining wall of the cylinder. The greater the deformation that was induced, the greater the restraint of the cylinder wall.

Thus, although the confining pressure varied continuously from initial atmospheric pressure to some unknown pressure at the end of the test, Adams demonstrated forcefully, if qualitatively, that rocks subjected to sufficient confining pressure will flow. Much more sophisticated work has been done since the early work of Adams, and the change from brittle to ductile behavior has been clearly demonstrated to be a function of confining pressure and temperature (see Paterson 1958; Heard 1960). Similarly, the strength of rock has been shown to be greatly affected by confining pressure, temperature, and strain rate (see Handin and Hager 1957, 1958; Griggs et al. 1960; Heard 1963).

The desire to predict the behavior of rock under conditions that exist within the earth underlies most of the work done in experimental rock deformation. Prediction is possible only when

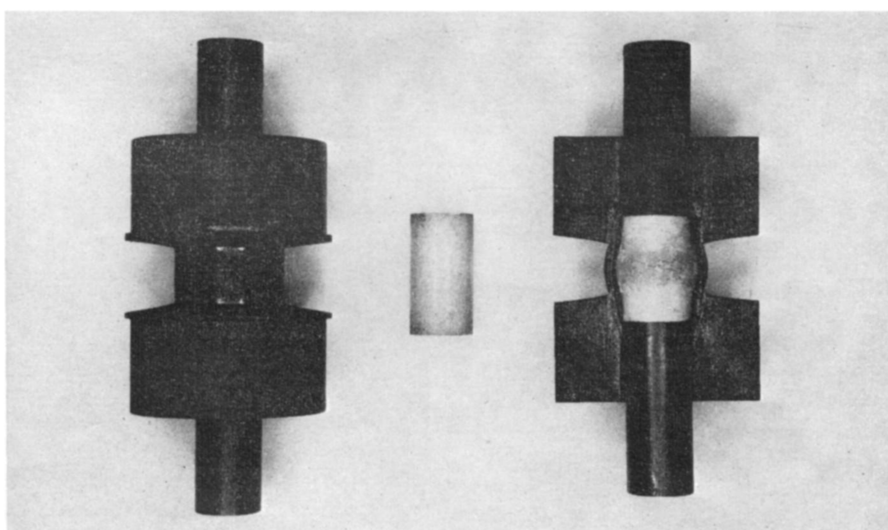


Fig. 1. Rock deformation apparatus used by F. D. Adams. Cylindrical specimen of rock is placed inside solid steel cylinder and compressed between steel pistons. Confining pressure is produced by restraint of cylinder wall in turned-down portion of cylinder. Undeformed and deformed specimens of marble are shown in the figure. The specimen diameter is $^{13}/_{16}$-inch.

the dynamics of deformation is well understood, and this can best be accomplished through controlled laboratory experiments in which the effects of important variables are studied systematically. The significance of such work clearly extends beyond geological and geophysical applications, for the predictable behavior of rock under different physical conditions is important, if not vital, to the success of many engineering operations. Studies in experimental rock deformation need not be directed solely at prediction, however, for they quite naturally provide a basis for the valid interpretation of deformational features and relationships found in geologic environments, and a means of determining the conditions that existed locally during the development of these features.

Geologic deformation has long been treated by geologists in a descriptive manner. The contortions of rock layers called folds, the localized offset (or faults) produced by fracture or flow, and other geologic structures have been studied and classified in great detail with increasingly refined techniques, but generally little attempt has been made to relate the observations to the mechanisms of deformation or to specific environmental or rock factors which permitted the mechanisms to operate. When the geologist looks at the effects of geologic deformation he must infer from the relationships he observes what has happened to produce a particular structure. The correctness of his inferences obviously depends on his understanding of the processes involved. Perhaps he can infer that a foliation or layering has been an important slip surface during geologic deformation (Fig. 2), but he may have no idea of the range of conditions under which such slip could occur. Possibly he can see that the rocks have been quite ductile and have flowed (Fig. 3), but he may not know whether the ductility reflects high pressure, high temperature, or other factors.

Fortunately, insight into these and similar problems can be provided through experimental rock deformation. Recognition of the intimate relationship between deformational geometry and the mechanisms of deformation, largely through an appreciation from experimental work of the effects of environmental and rock factors on deformation, has led in recent years to a more dynamic approach to the subject (see e.g. Donath 1962, 1963, 1967; Donath and Parker 1964; Friedman 1964). A comprehensive survey of dynamic structural geology would be too ambitious for an article of this length. Hopefully, the illustrations presented, which are based largely on previously unpublished data, will be adequate to provide an appreciation for the potential of the dynamic approach.

Fracture versus flow

All nonrecoverable geologic deformation consists of fracture or flow or a combination of the two, and is caused by the stress difference (i.e. differential stress) between a directed stress and the confining pressure in the deformational environment. Fracture

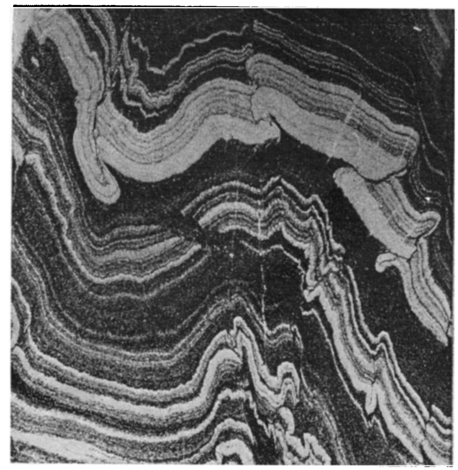


Fig. 3. Folded sandy dolostone laminae in slate of the West Castleton formation near West Castleton, Vermont. The disharmonic character of the folding reflects the presence of ductility contrast between layers in a layered sequence of generally high mean ductility. (Photographed from a thin section by B. Voight; approximately natural scale.)

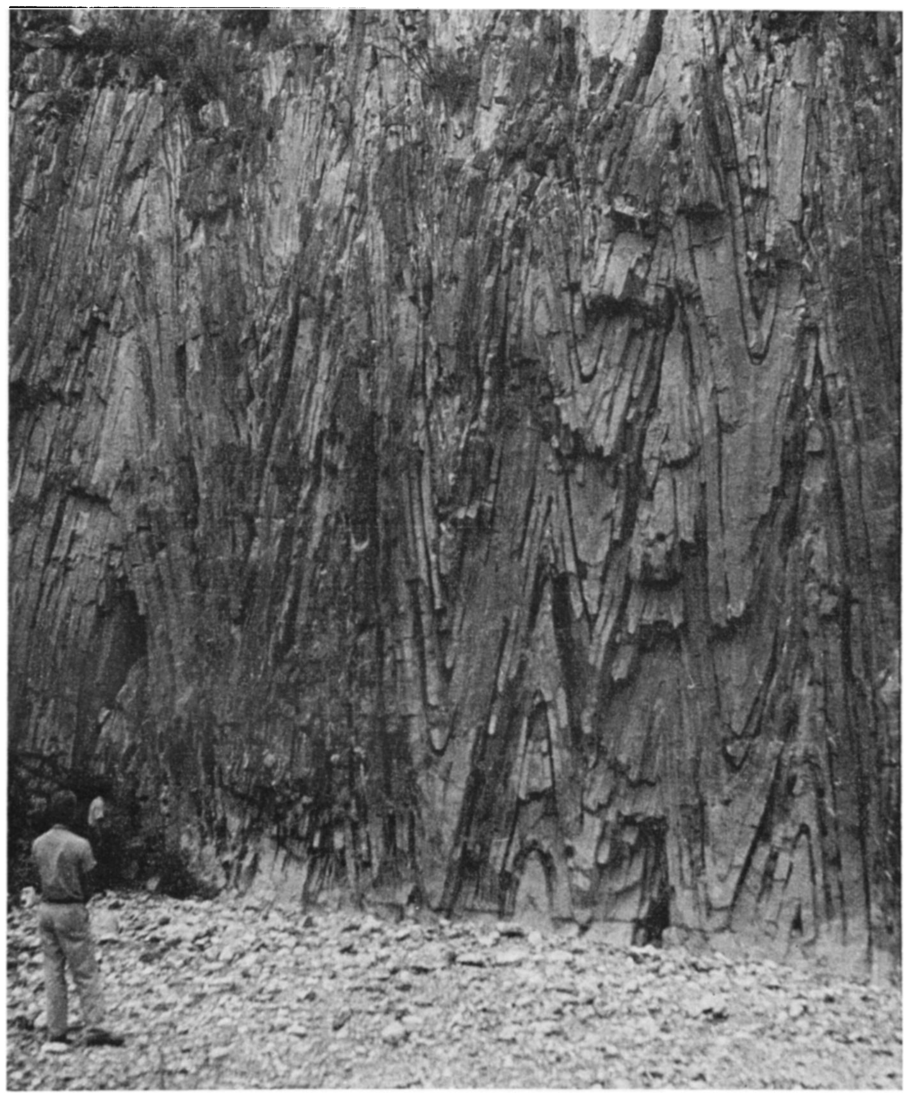


Fig. 2. Folding of thin-bedded limestones in the Cuesta del Cura formation near San Cristobal, Mexico. The bedding has been an important factor in the deformation, with slip having occurred between the limestone layers followed and accompanied by flow within the layers. (Photo by Emily H. Vokes)

is deformation that occurs with loss of cohesion; material which loses cohesion separates into two or more pieces, commonly with rapid release of stored elastic energy (as in shallow earthquakes). Flow is deformation, not instantly recoverable, that occurs without loss of cohesion; material may flow homogeneously, or discontinuously along a narrow zone (called a ductile fault).

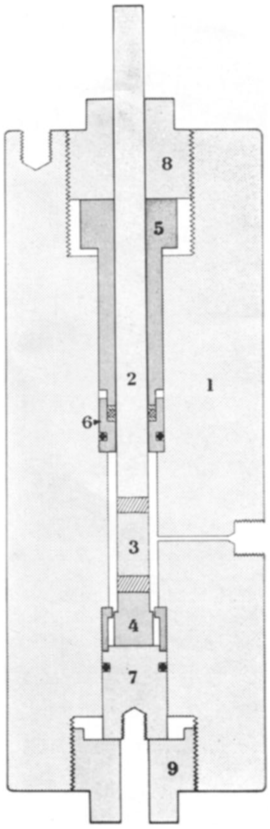
In experimental work the deformation is commonly expressed quantitatively by the amount of shortening of a cylindrical specimen relative to its orginal length (i.e. by longitudinal strain), or descriptively by the mode of deformation, which is the macroscopic expression of the microscopic and submicroscopic processes of deformation. A given rock subjected to differential stress may fracture, fault with or without loss of cohesion, or flow, depending on the specific conditions under which the differential stress acts. The amount of permanent deformation that a rock can undergo without fracturing or faulting, hence its mode, is dependent upon its ductility. A rock is considered to be brittle if the total strain before fracture or faulting is less than 5 percent, moderately ductile if between 5 and 10 percent, and ductile if it is greater than 10 percent.

Experimental apparatus used to deform rocks must provide a means of applying both confining pressure and a directed (axial) stress. An apparatus in the author's laboratory is shown in Figure 4, and a schematic drawing of a typical pressure vessel is shown in Figure 5. The test specimen is placed between an anvil and a piston, jacketed with a suitable thin-wall metal or plastic tubing, and inserted in the pressure vessel. Pressure seals and end plugs are put into place and the vessel is then positioned between the platens of a loading press. Confining pressure is produced by pumping a liquid into the vessel under pressure, and axial stress is created by loading the piston with the press. The confining pressure may be thought of as the pressure of the overburden; a pressure of 250 bars (1 bar equals 14.5 psi) would be approximately equivalent to the pressure produced by the overlying rocks at a depth of one kilometer in the earth. The differential stress would represent the stress difference caused by tectonic stresses acting within the earth. The shortening of the specimen, from which the longitudinal strain is calculated, is determined from the displacement of the piston into the pressure vessel. The applied axial stress is calculated from the measured axial load and the cross-sectional area of the specimen. Corrections must be made for elastic distortion of the apparatus and for changes in cross-sectional area of the specimen with continuing deformation in order to get true stress and longitudinal strain. The deformation is most commonly represented as a curve of differential stress (axial stress minus confining pressure) versus longitudinal strain.

Fig. 4. High pressure rock deformation apparatus. The apparatus shown in the lower photograph consists of four 50-ton tie-bar presses which, although mounted in a common frame, operate independently of one another. Two pressure vessels are in position on the right side of the apparatus and a third, which is inside a high-temperature furnace, is partly exposed behind the sliding-door shields. Pumps for creating confining pressure and for activating the rams are located on the floor beneath the apparatus. Pressure recorder-controllers and other instrumentation are shown in the upper photograph. The smaller cabinet at the left houses a PDP-8/S computer and Vidar data acquisition system for automatic selection and recording of test data.

The stress-strain curves of Figure 6 illustrate the change in behavior of a limestone as it is deformed under increasingly higher confining pressure. All of the curves show an initial nearly linear increase in sustained differential stress with increasing strain. This portion of the curve represents essentially elastic, or recoverable, deformation. Further strain causes permanent deformation, or yielding. At 200 bars confining pressure, the pressure equivalent of about 0.8 kilometer of overburden, yielding is followed by an abrupt drop in the sustained differential stress to a level that remains more or less constant with continuing strain. The decrease in resistance to differential stress reflects a loss of cohesion in the limestone during the development of a shear fracture, and the subhorizontal portion of the curve represents the differential stress required to produce continued displacement along this fracture.

The character of the stress-strain curves can be seen to change systematically with increasing confining



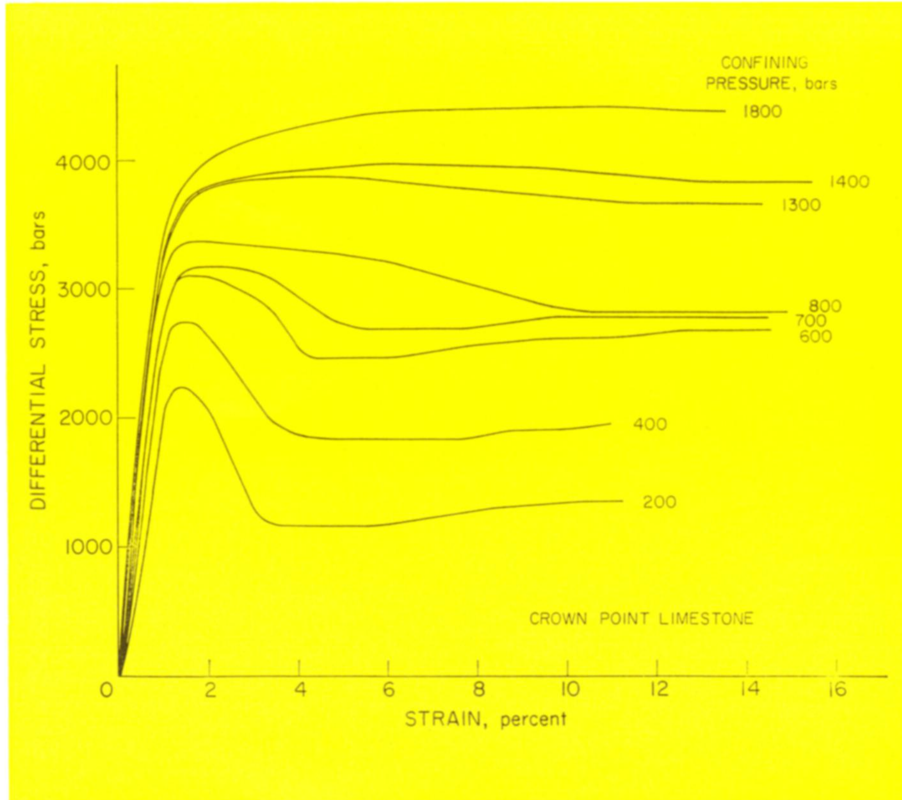


Fig. 6. Effect of increased confining pressure on sustained differential stress (strength) of Crown Point limestone. Curves of differential stress versus longitudinal strain show the increase in strength caused by increased pressure and a change in character as the deformation changes from brittle at low confining pressure to ductile at high pressure.

Fig. 5. Schematic drawing of a typical pressure vessel used for experimental rock deformation. The cylindrical specimen (3), commonly ½-inch diameter by 1-inch length, is placed between a piston (2) and an anvil (4). Pressure seals (6 and 7) prevent escape of the confining pressure medium from the pressure vessel. The specimen jacket—which holds the piston, specimen, anvil, and spacers together and prevents the pressure medium from entering pore spaces in the specimen—is not shown in the drawing.

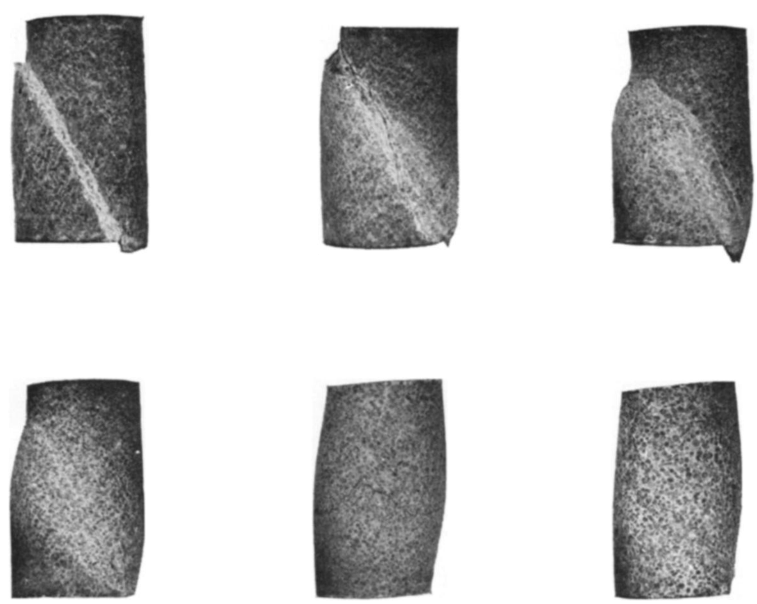


Fig. 7. Specimens of Crown Point limestone deformed to approximately the same total strain (about 15 percent) at different confining pressures. Increased confining pressure causes a transition in deformational mode from shear fracture at 200 bars (*top left*) to a well-defined ductile fault at 700 bars (*top right*) and at 900 bars (*bottom left*) to incipient ductile faults at 1400 and 1800 bars (*bottom center* and *right*). A shear zone developed in the specimen deformed at 600 bars (*top center*). Specimens were initially ½″ diameter by 1″ length.

pressure, and, at 1800 bars, yielding is followed by a nearly constant sustained differential stress with no indication of a stress drop. Examination of the specimens associated with these curves shows that a complete transition in the mode of deformation exists, from shear fracture at 200 bars to incipient ductile faulting at 1800 bars (Fig. 7). There is, as well, a change in the mode of deformation with increasing strain. The five specimens shown in Figure 8 were each deformed at 600 bars confining pressure, but to increasingly greater total strains. The transition from homogeneous deformation to a well developed shear zone characterized by loss of cohesion is clearly seen.

Deformational mode fields

The systematic variation of deformational mode with confining pressure and total strain permits a "field" to be defined for each mode of deformation, as shown in Figure 9 for the Crown Point limestone. Two fracture modes are recognized: extension fracture, characterized by initial displacement perpendicular to the fracture surface and oriented perpendicular to the direction of least compression (or maximum tension), and shear fracture, characterized by initial displacement parallel to the fracture surface and oriented at approximately 30 degress to the direction of maximum compression. Flow on a macroscopic scale can be homogeneous or discontinuous (e.g. faulting). Thus, faulting can reflect either fracture or flow, and can therefore occur in either brittle or ductile material. Localized offset along a shear fracture or shear zone constitutes brittle faulting, whereas ductile faulting is localized offset produced by a velocity discontinuity in flow. Shear fracture generally implies failure restricted to a single surface, but brittle faulting includes all faulting that involves permanent loss of cohesion whether restricted to a single surface (shear fracture) or distributed throughout a zone. Although a mode is not *sensu stricto* brittle or ductile, clarity favors direct correlation between material behavior and effect—hence, brittle fault and ductile fault. Characteristic modes of deformation in homogeneous, isotropic rocks are, therefore, extension fracture, brittle fault, ductile fault, and homogeneous flow.

A deformational mode field (DMF) diagram like that of Figure 9 is a potentially valuable tool in the interpretation of natural deformational environments. If, for example, the mode of deformation and the strain can be determined in the naturally deformed rock, reasonable limits for the effective confining pressure during deformation in the natural environment might be ascertained. Whether a DMF diagram might be successfully applied in this manner depends upon: (1) whether an accurate and definitive diagram can be constructed from the experimental data for a particular rock; and (2) whether the natural counterparts of the experimental modes can be recognized and thus correlated. If the mode of deformation is affected by temperature or

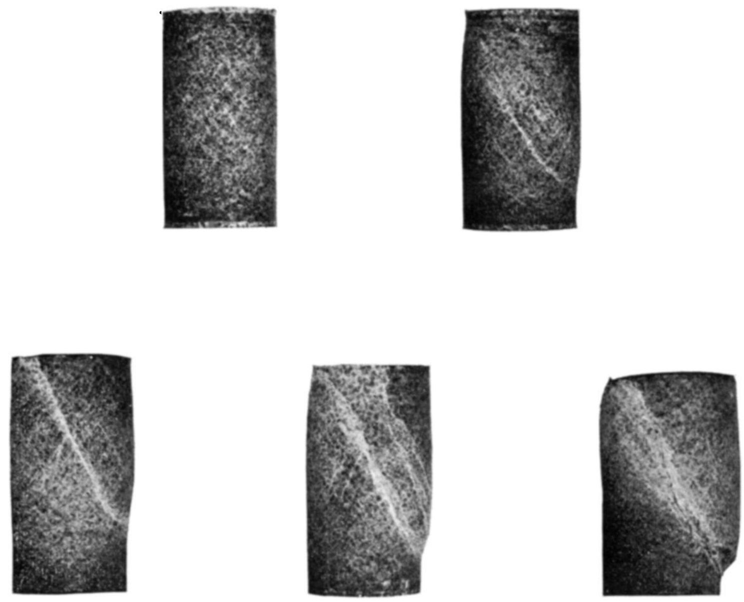
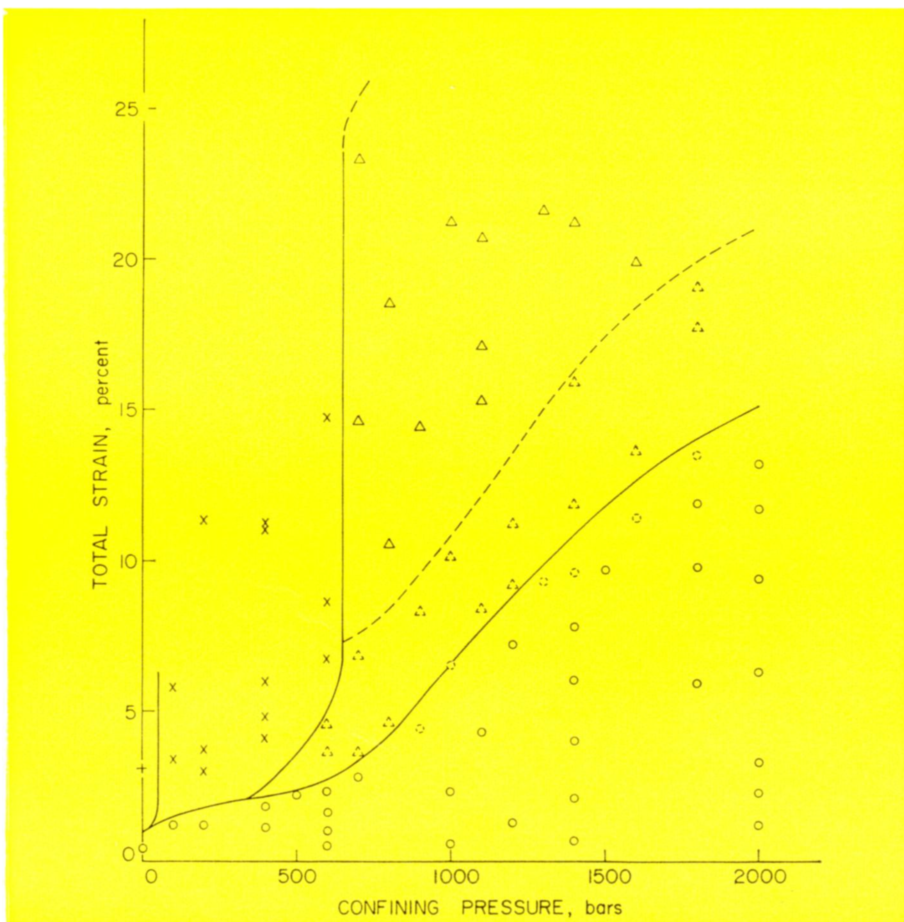


Fig. 8. Specimens of Crown Point limestone deformed at 600 bars confining pressure to different total strains. The development of a shear zone is seen as the strain increases from 2.3 percent (*top left*) to 14.7 percent (*bottom right*). Intermediate specimens have undergone strains of 4.5 (*top right*), 6.7 (*bottom left*), and 8.6 (*bottom center*) percent, respectively. Specimens were initially ½″ diameter by 1″ length.

Fig. 9. Deformational mode field diagram for Crown Point limestone deformed at room temperature and a strain rate of 10^{-5} per second (.001% per second). The mode of deformation observed for a specimen deformed at a given confining pressure and to a particular total strain is indicated by a specific symbol: extension fracture by (+); brittle fault (shear fracture or shear zone) by (×); ductile fault by (△); and homogeneous deformation by (○). Dashed boundary line indicates strain at which ductile faults are macroscopically recognizable. Compare Fig. 12.



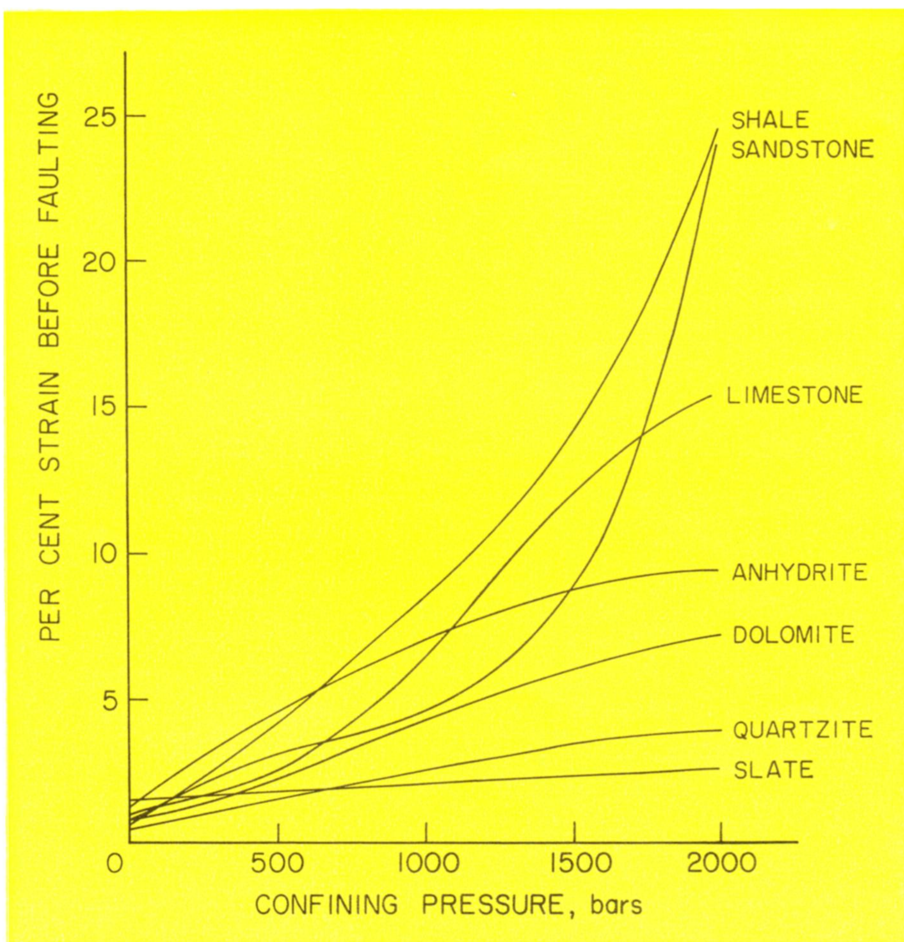
strain rate, as would generally be true for limestone, additional diagrams would be required for different temperatures and strain rates in order to define the fields in the n-dimensional space of the important parameters. However, there is no need to attempt this unless the fields in confining pressure–total strain space and the modes themselves can be accurately defined.

The boundary in Figure 9 that separates the field of homogeneous deformation from the fields of fracture and faulting is a measure of ductility, as it indicates the percent strain the rock can undergo before fracture or faulting. The largely vertical boundary at 650 bars separating the field of brittle faulting from that of ductile faulting is equally significant, for it indicates the minimum confining pressure under which deformation without loss of cohesion (i.e. flow) can occur. Thus, it can be thought of as defining the fracture to flow transition. Although the Crown Point limestone deforms at room temperature and at a strain rate of 10^{-5} per second without total loss of cohesion at confining pressures of 650 bars and higher, by our previous definition the rock is not considered "ductile" under these test conditions until the pressure exceeds 1300 bars, at which pressure it flows 10 percent without fracturing or faulting. Between 850 and 1300 bars confining pressure the rock shows moderately ductile behavior, and below 850 bars it faults before undergoing 5 percent strain.

The fracture-flow boundary is determined solely on the basis of whether or not cohesion has been lost, and it appears that the boundary can be fixed quite accurately. Positioning of the ductility boundary is generally more difficult, as one must ascertain at what percent strain fracture or faulting is initiated. For fracture, this is readily determined from the stress-strain curve; however, the onset of ductile faulting is not always easily recognized in the stress-strain relationships because a complete gradation exists from homogeneous flow to ductile faulting with increasing total strain (see Figs. 6 and 8). The boundary shown in Figure 9 is based on microscopic evidence, which allows it to be placed rather precisely. In some instances the percent strain to the ultimate strength is the most practical measure of ductility, and it is possibly the next most accurate alternative. The ultimate strength is the maximum differential stress that the rock can sustain under the conditions of deformation, and it is reflected as a true maximum in the stress-strain curve. The assumption here is that faulting is initiated at the point (strain) at which the specimen begins to weaken. However, commonly, the ultimate strength is reached between one to four percent strain before initiation of faulting, if faulting occurs, although in some rocks faulting may be initiated even before the ultimate strength is reached. By whichever method the ductility boundary is defined, marked differences in the curves are obtained for different rocks, as shown in Figure 10.

Because it is possible to define a deformational mode field diagram with reasonable ease and accuracy by experimental means, thus relating experimental modes of deformation to environmental conditions, one needs only to be able to correlate experimental modes with natural deformation in order to infer something about the conditions under which the natural deformation has occurred. Representative thin sections of the modes of deformation in the Crown Point limestone have been intensively studied for this purpose; several are

Fig. 10. Ductility (percent strain before fracture or faulting) of several common rocks as a function of confining pressure. The ductility is seen to increase with increasing confining pressure, but to differing degrees. (Data for several rocks are from Handin and Hager, 1957)



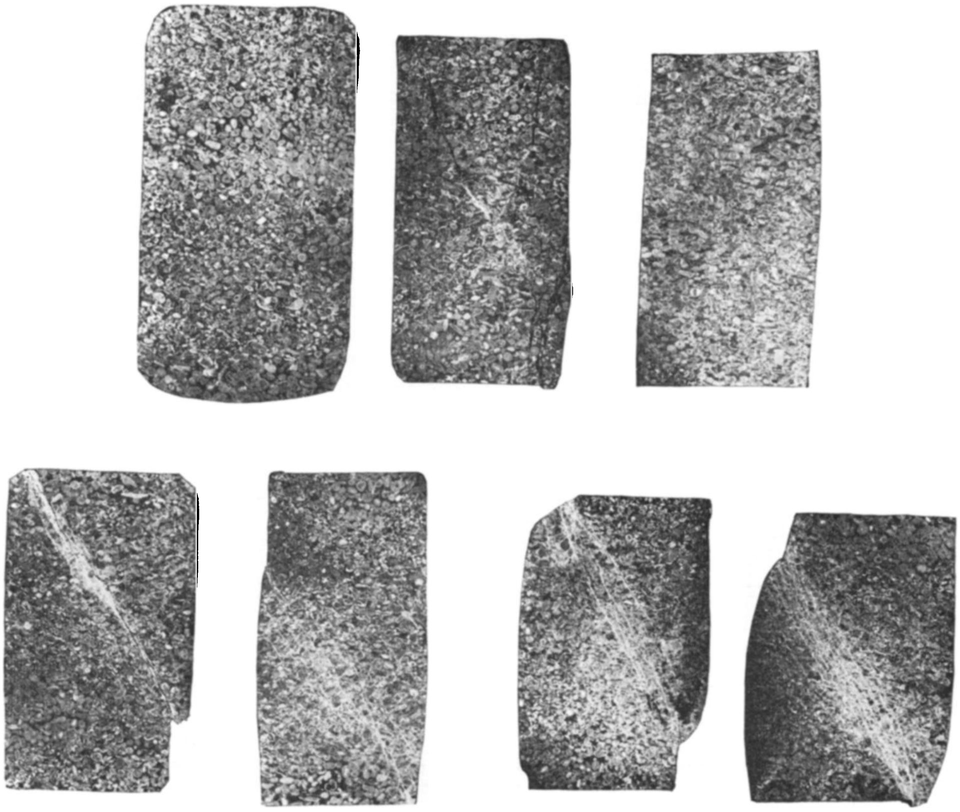


Fig. 11. Thin sections of typical modes of deformation in Crown Point limestone: undeformed fabric (*top left*); extension fracture, 1 bar confining pressure, 3.1% strain (*top center*); homogeneous flow, 1800 bars, 9.8% (*top right*); shear fracture, 200 bars, 11.3% (*bottom left*); shear zone, 600 bars, 6.7% (*bottom center left*); ductile faults, 800 bars, 18.5%, and 1000 bars, 21.2% (*bottom right*). The two specimens at bottom left lost cohesion during faulting; the two specimens at bottom right faulted without total loss of cohesion even though the total strain exceeded 18 percent. Specimens were initially ½″ diameter by 1″ length.

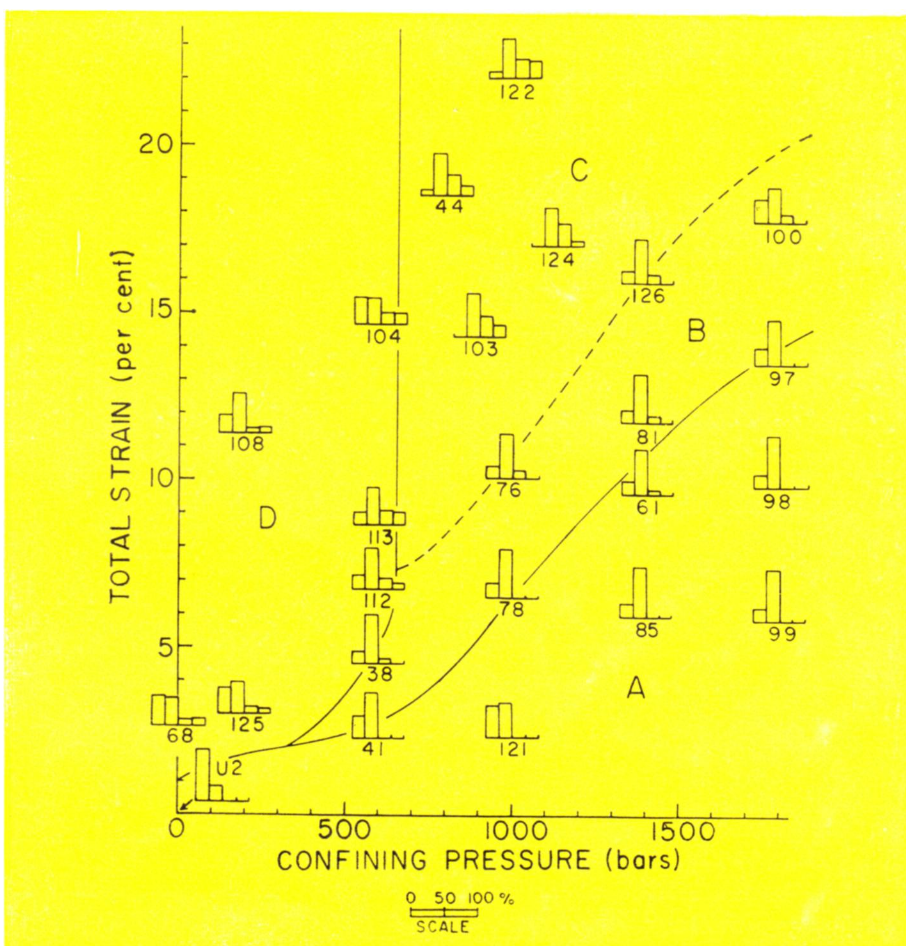


Fig. 12. Deformational mode field diagram for Crown Point limestone with modes defined by microscopic criteria. Bar graphs represent the relative contribution to the total fabric of four end-member categories (*left to right*): (1) undeformed grains; (2) grains with intracrystalline lamellae; (3) grains with fractures; (4) slip surfaces or faults extending across several grains. At high strains some completely twinned (category 2) components appear to be undeformed (category 1). The center of the base line of each bar graph indicates the amount of total strain and confining pressure. Field A is homogeneous deformation; field B is ductile faulting recognizable only by microscopic criteria; field C is ductile faulting recognizable macroscopically; and field D is brittle faulting. (From D. G. Tobin, Ph.D. thesis, 1966, Columbia University)

shown in Figure 11. It was found through multivariate analysis of features in the sections that the experimental modes of deformation could be represented in terms of four end-member categories: (1) undeformed grains, (2) grains with intracrystalline lamellae, (3) grains with fractures, and (4) slip surfaces extending across several grains. Figure 12 shows the deformational mode fields as defined by microscopic criteria related to the four end-member categories. It will be noted from Figure 12 that the onset of ductile faulting is indicated microscopically by the appearance of fractures within grains. Presumably, one could study a naturally deformed rock in like manner and determine the relative contribution to the macroscopic effect of undeformed grains, intracrystalline gliding, fracture within grains, and faulting or slip across the grains. The results could then be correlated with an experimental counterpart.

Suppose that two different rocks have been naturally deformed in the same environment under identical conditions and that their deformational mode fields are experimentally defined in the total strain–confining pressure plane as shown in Figure 13. Suppose further that one (A) shows ductile faulting whereas the other (B) shows a shear fracture mode. For these conditions reasonable limits could be placed on the confining pressure under which deformation occurred: the pressure must have exceeded 500 bars for ductile faulting to have developed in rock A, but must have been lower than 700 bars for fracture to have taken place in B. Similarly, if the strain is independently determined to be slightly greater than 10 percent and rock A shows homogeneous flow whereas rock B shows ductile faulting, the pressure must have been between 1000 and 1400 bars. This is obviously an oversimplified example; we would have to consider also the effects of strain rate, temperature, and solutions. Fortunately, however, it is commonly possible to place certain restrictions on the position in an n-dimensional DMF diagram. Other techniques can be used to determine paleotemperature, for example, and one can commonly determine the amount of strain in the rocks.

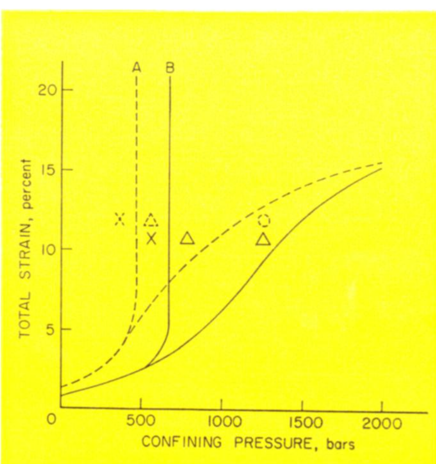


Fig. 13. Deformational mode field (DMF) diagrams for two different rocks. The coexistence of different modes of deformation in the same deformational environment may be useful in determining the conditions of deformation (see text). Brittle faults represented by (×), ductile faults by (△), and homogeneous deformation by (○).

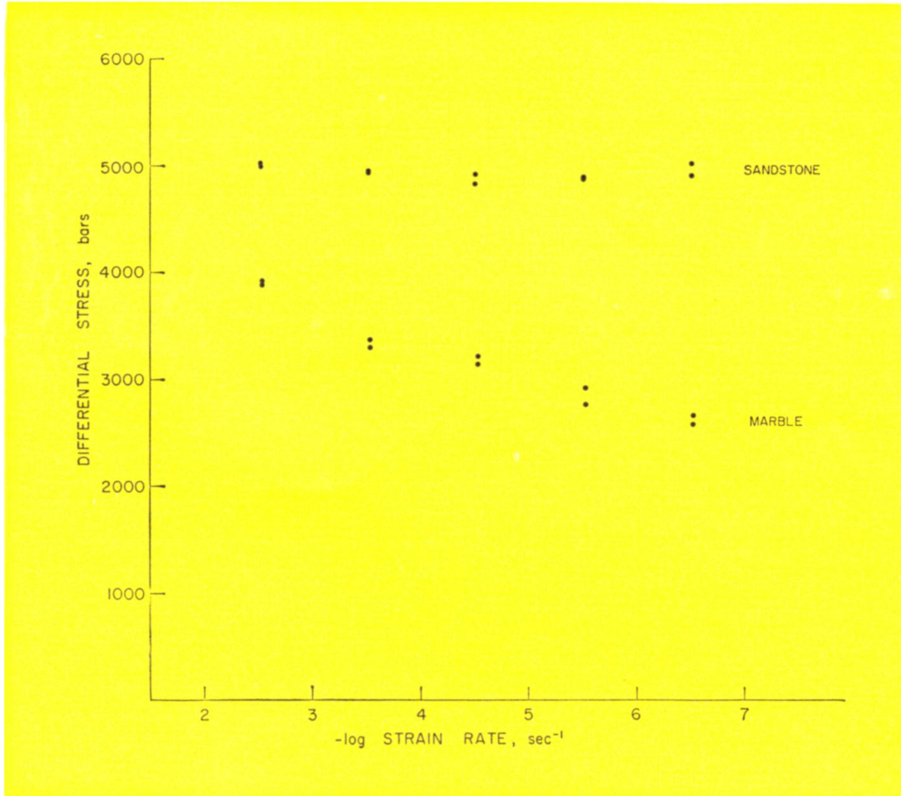


Fig. 14. Effect of decreasing strain rate on the strength of sandstone and marble deformed at 2000 bars confining pressure. Although there is a marked decrease in strength of the marble, the strength of the sandstone is unaffected over the same range of strain rate. The values indicated are for the differential stress sustained just beyond the knee of the stress-strain curve, at 2% for the marble and 3.5% for the sandstone.

Factors affecting ductility and strength

In addition to the obvious effects of confining pressure, already discussed, marked contrasts in the deformational behavior of rock can occur as a consequence of increased temperature, the presence of solutions, or deformation extended over periods of thousands or millions of years. Inasmuch as it is impossible to duplicate in the laboratory the extremely slow strain rates that may obtain in natural deformation, one must evaluate the effects over a limited range and extrapolate to geologically significant rates. Interestingly, it appears that whether strain rate is an important consideration or not depends upon the specific mechanisms of deformation. The effects of strain rate can sometimes be dramatic, as in marble. Heard (1963) concluded from his experimental study of large changes in strain rate in Yule marble that the strength would drop from about 3700 bars at room temperature and a strain rate of 4.0×10^{-1} per second to 10^{-3} bars at 500°C and at an extrapolated "geologic" strain rate of 10^{-14} per second—that is, the rock would be essentially incapable of sustaining any differential stress and would flow continuously!

The data in Figure 14 for marble are consistent with Heard's observations, and show a marked decrease in the sustained differential stress at 2 percent strain for this rock over the range of strain rate indicated. In contrast, the sandstone shows no significant effect under the same conditions of testing. The different responses reflect different microscopic mechanisms of flow. In the marble, flow was effected by intracrystalline gliding; in the sandstone, cataclasis was the principal mechanism of flow. Cataclasis involves fracturing across and around individual grains, rotation of grains, and sliding between grains. Although microscopic fracture is a significant process in cataclasis, individual fractures are minute and are distributed in such a way that cohesion of the rock as a whole is not lost. Experimental data for several other rocks support the observation that the strength and ductility of cataclastically deformed rock is independent of the rate of deformation at rates slower than 10^{-3} per second; the effect of strain rate is absent even in limestones and marbles when they are deformed under conditions favoring cataclasis. These results suggest that laboratory strain rates are adequate to evaluate the deformational behavior of rock deformed cataclastically under geologic strain rates.

Where rate is known to be important, as in intracrystalline gliding, it may be possible to place limits on both temperatures and strain rates by determining the predominating flow mechanisms microscopically. A study reported by Raleigh (1969) shows considerable promise in this regard.

Increased temperature generally reduces the strength and increases the ductility of rock (see Griggs et al. 1960; Handin and Hager 1958). In a few instances higher temperature may raise the ultimate strength by (1) increasing the ductility in strain-hardening rocks—i.e. in rocks deforming largely by intracrystalline gliding and requiring an additional increment of differential stress for each additional increment of strain—provided that the increased ductility is not accompanied by annealing of the grains, or by (2) favoring a change in mechanism of deformation that permits greater strains and higher stresses. The effect of increased temperature would thus generally be to raise somewhat the ductility boundary in the DMF diagram and to shift the fracture-flow boundary to a lower confining pressure. For conditions existing in the upper part of the earth's crust, however, any shift of these boundaries caused by a temperature increase would be less significant than the changes in deformational behavior induced by the corresponding change in pressure with depth, since pressure effects always override temperature effects for depths less than about 10 kilometers. Moreover, because temperature can commonly be determined by other techniques, or estimated within reasonable limits from petrologic observations, this information can be used to help delimit the part of DMF space that applies.

Solutions present in the pore spaces of rock may have a purely mechanical effect, normally decreasing ductility as well as strength, or they may promote recrystallization of certain mineral phases and thus enhance ductility. The mechanical effects of pore solutions are readily evaluated, as the behavior of a given rock is simply a function of the effective pressure under which it is deformed, and the effective pressure is equal to the confining pressure minus the pore pressure. Thus, if the pore pressure equals the confining pressure, the effective pressure is zero. The relationship appears to hold for all rocks provided the rate of deformation is less than some critical value. This critical value depends on the permeability of the rock, viscosity of the pore fluid, and on certain geometric factors, and is about 10^{-7} per second in rocks of low permeability saturated with water (Brace 1969).

In some instances solutions clearly have important chemical as well as mechanical effects. They may contribute significantly to the lowering of strength or to the enhancement of ductility. Certain effects appear to be a function of ionic mobility rather than solubility of the rock in the pore fluids. Griggs (1940) found, for example, that the creep rate in gypsum (hydrous calcium sulphate) subjected to constant differential stress was greater in a calcium chloride solution than in distilled water, even though the solubility of gypsum in the solution was 45 percent lower than in water. If solutions have been influential in natural deformation, this might be revealed by the evidence of recrystallization or related effects.

In addition to certain inherent rock factors, such as composition and grain size, and the environmental factors (pressure, temperature, etc.), two other factors can strongly affect the deformational behavior of rock and may be useful in dynamic interpretation. These are the previous strain history of the rock and the presence of planar anisotropy.

Effects of previous strain history

Inasmuch as naturally deformed rocks are not uncommonly subjected to more than one period of deformation, it is useful to evaluate the possible effects of repeated deformations. Moreover, because the principal stress directions may change between deformations the effects should be evaluated for repeated deformations of both constant stress orientation and of changed orientation. The most practical way of doing this experimentally is by deforming large cylindrical cores of rock to some desired percent strain, constituting the "prestrain," and then coring the center of the deformed specimen to obtain a smaller cylindrical sample for subsequent additional deformation.

Two parent specimens, one with a coaxial daughter specimen and one with a transverse daughter specimen, are shown in Figure 15. The stress-strain curves for these specimens, given in Figure 16, show several noteworthy relationships. The coaxial specimen, which has been subjected to 12.6 percent longitudinal prestrain, is appreciably stronger than the previously undeformed specimen and the knee of its stress-strain curve is more clearly defined at yielding. In sharp contrast, the transverse core, which has been subjected to 12.5 percent longitudinal prestrain, is considerably weaker than the parent specimen and the knee of its stress-strain curve is completely obliterated. Interestingly, the transverse specimen can sustain a higher differential stress than the parent specimen at strains greater than 7 percent; its strength approaches that of the coaxial specimen at 16 percent. These results suggest that much higher stresses would be required in natural deformation for repeated deformation under constant stress orientation, but that a change in the orientation of the principal stresses would favor additional deformation initially at rather low levels of differential stress.

The increase in sustained differential stress with increasing prestrain for coaxial cores is quite significant, and the relationship suggests a possible means of evaluating the amount of

Fig. 15. Deformed parent specimens of Salem limestone with subsequently further deformed daughter cores. The parent specimens have both been shortened about 12.5%; the daughter cores were subsequently deformed an additional 30%. Each specimen was deformed at 1000 bars confining pressure, and all showed homogeneous flow. Parent specimens were initially 1″ diameter by 2″ length.

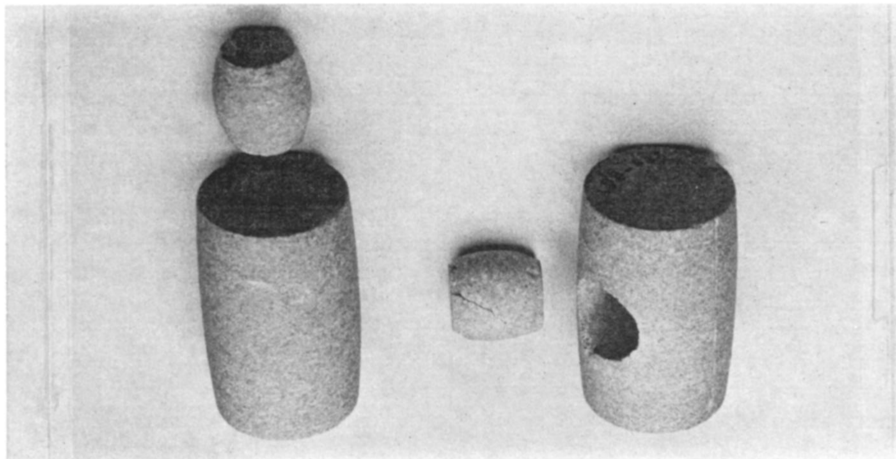


Fig. 16. Stress-strain curves for coaxial and transverse cores from Salem limestone specimens that had been prestrained, compared with a previously undeformed specimen. (After E. Karp, Ph.D. thesis, 1966, Columbia University)

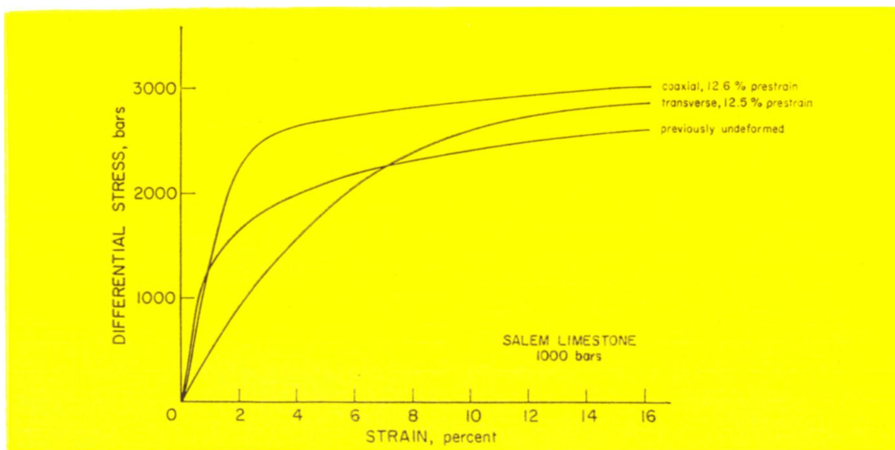


Fig. 17. Stress-strain curves for coaxially deformed Salem limestone showing effect of magnitude of prestrain on the yield stress and strength. (After E. Karp, Ph.D. thesis, 1966, Columbia University)

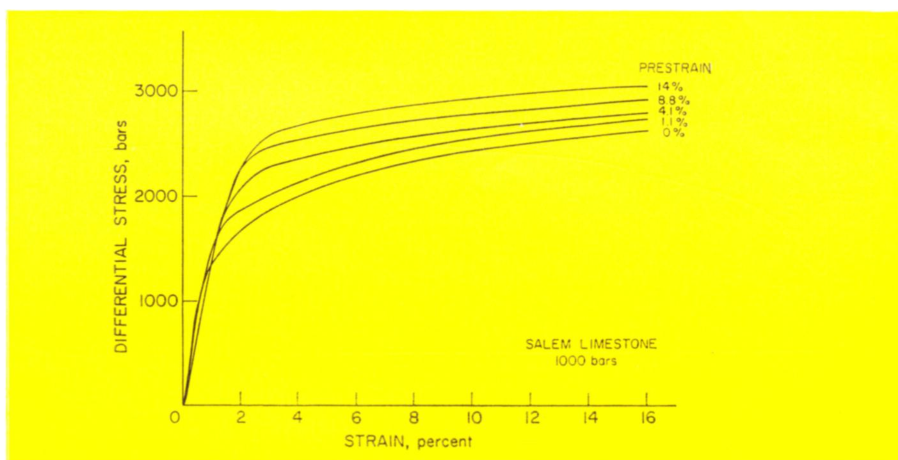
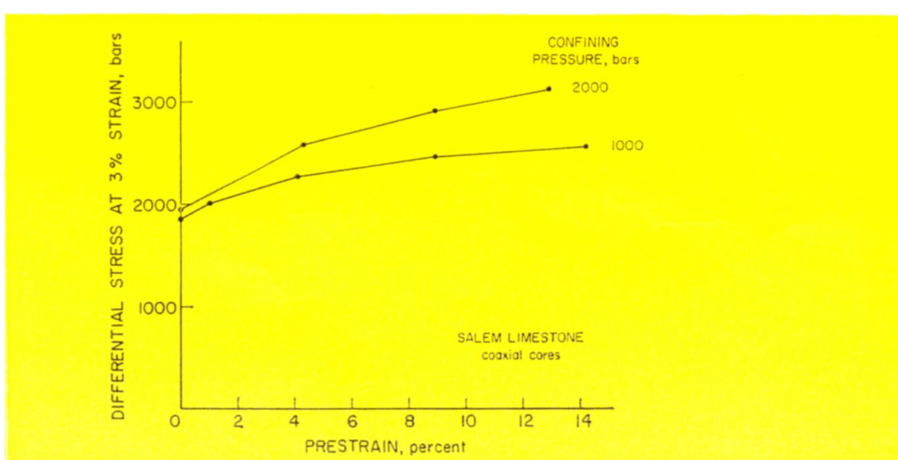


Fig. 18. Effect of prestrain on the sustained differential stress at 3% strain for coaxial daughter cores of Salem limestone deformed at 1000 and 2000 bars confining pressure. Parent specimens were deformed at 1000 bars pressure. (After E. Karp, Ph.D. thesis, 1966, Columbia University)



previous strain that a strain-hardening rock has been subjected to in natural deformation. The stress-strain curves for several coaxial cores prestrained to differing amounts at 1000 bars confining pressure are shown in Figure 17, and the differential stress at 3 percent strain is plotted as a function of prestrain in Figure 18 for coaxial daughter cores deformed at two different confining pressures. The individual stress-strain curves shown in Figure 17 are significantly different for prestrain up to 14 percent, and reflect the increase in stress required for yielding. The effects are even greater for specimens prestrained at 1000 bars and then further deformed under 2000 bars pressure (see Fig. 18).

Suppose one were to take two samples of limestone from the folded sequence shown in Figure 2, one from the limb (steeply dipping straight portion) of the fold and the other from the hinge (area of maximum curvature), as well as a sample from the same rock unit in an area where the rock has not been deformed. If the three samples were then deformed under identical conditions in the laboratory, one could expect to get three different stress-strain curves, much like those of Figure 17. The differences would reflect the differing amounts of strain produced by natural deformation. The individual curves would be analogous to separate stress-strain cycles that collectively define a stress-strain curve for a continuously deformed virgin specimen, like the relationships shown in Figure 19. The individual curves could be shifted parallel to the strain axis until they fit the curve for the previously undeformed specimen; their intercepts with the strain axis would indicate the amount of previous strain. Again, we are dealing with an oversimplified illustration. For the previously deformed specimens one would likely have to evaluate the effects for each of three orthogonal orientations (preferably coinciding with the principal strain directions), and possibly at several different pressures. Nevertheless, even crude results could provide valuable insights into certain aspects of natural deformation.

The observed strength differences among prestrained limestone samples reflect largely the changes in orientation of intracrystalline slip surfaces caused by deformation. Grains in which the slip surfaces are most favorably oriented with respect to the principal stress directions deform first; increasingly higher stresses are required to cause continued deformation as these surfaces rotate (with increasing strain) and to initiate intracrystalline gliding in the less favorably oriented grains. One consequence of this, in addition to the strain-hardening effects already discussed, is the development of a preferred orientation of grains with continuing strain.

Although coaxial deformation produces equal lateral strains about the axis of compression, transverse specimens show marked ellipticity. Cross sections of the coaxial and transverse specimens of Figures 15 and 16 are shown in Figure 20. The elliptical cross section of the transverse specimen apparently reflects a weak anisotropy induced by intracrystalline deformation in the parent specimen. Of significance is the observation that the long axis of the ellipse is parallel to the direction of maximum compression in the parent specimen.

Thus, it appears that for rocks which

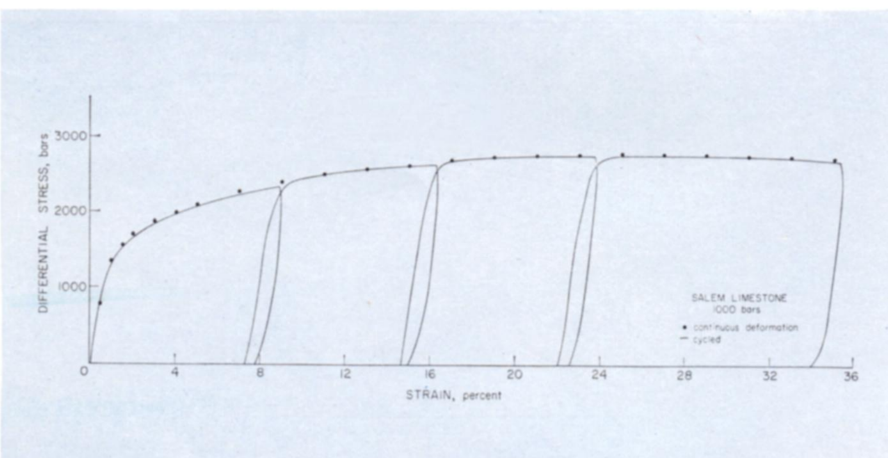


Fig. 19. Stress-strain curves for continuously deformed and for cycled (loaded and unloaded) Salem limestone specimens deformed at 1000 bars confining pressure. (After E. Karp, Ph.D. thesis, 1966, Columbia University)

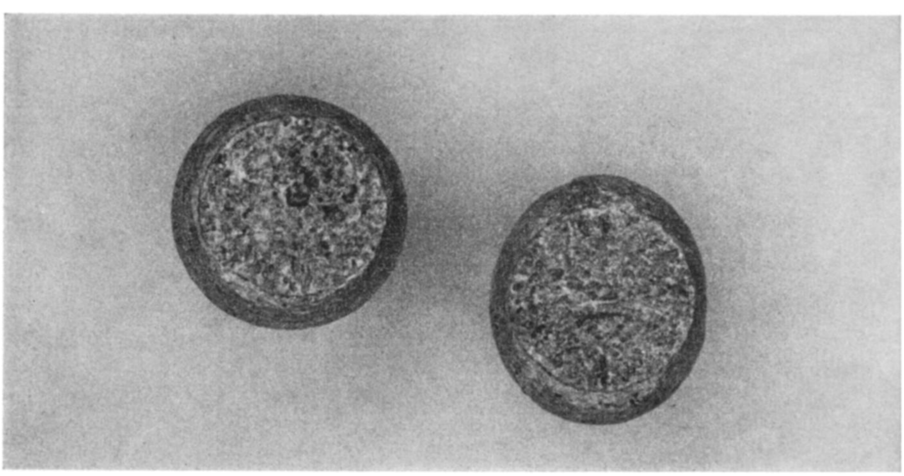
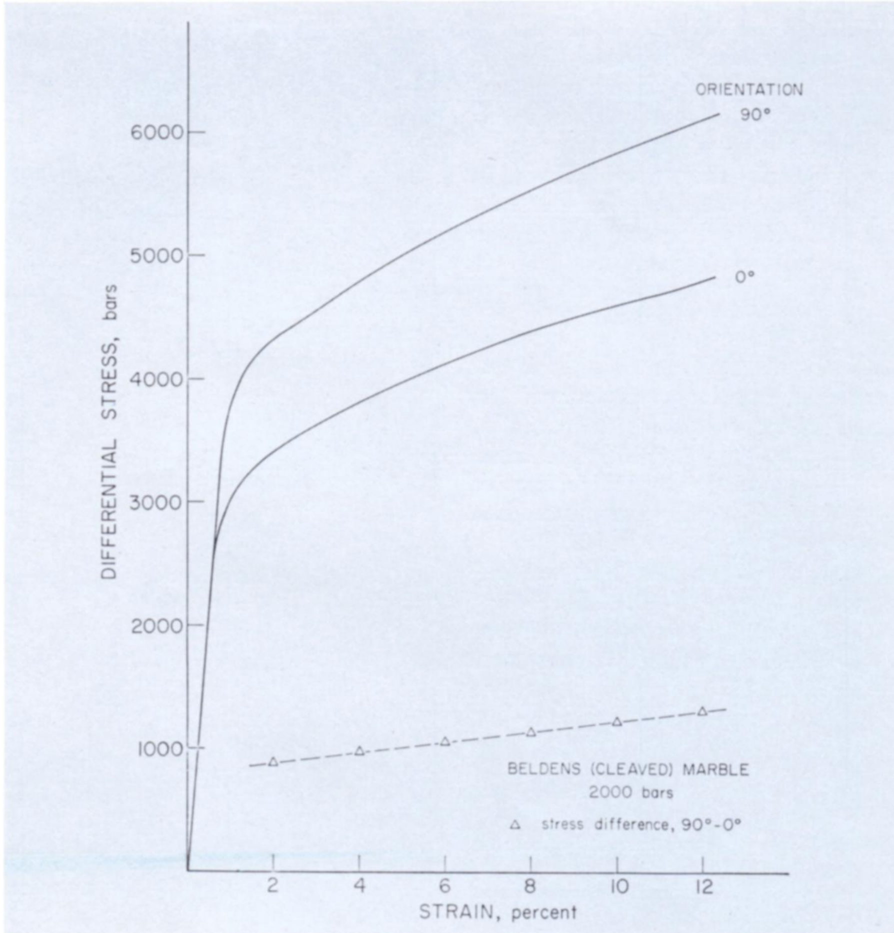


Fig. 20. Cross-sections of coaxial (*left*) and transverse (*right*) cores of Salem limestone deformed to approximately 30% strain after the parent specimens had undergone a longitudinal strain of 12.5%. Lateral strains in the coaxial core are equal, giving rise to a circular cross-section. The long axis of the elliptical cross-section in the transverse core is parallel to the direction of maximum compression in the parent specimen.

Fig. 21. Stress-strain curves for specimens of Beldens marble compressed at 90° and at 0° to cleavage in the marble under a confining pressure of 2000 bars. The 90-degree orientation is significantly stronger initially and becomes increasingly stronger with increasing total strain.



deform largely by intracrystalline gliding, we may have in this phenomenon a technique for determining the direction of maximum compression in natural deformation. One would need only to deform three orthogonal specimens from a single block of naturally deformed rock under identical conditions in the laboratory and analyze the distribution of lateral strain in each in order to determine this direction. The determination of the principal stress directions in naturally deformed rock is of major importance in dynamic structural geology.

Natural deformation of limestone, particularly at high temperature, can produce marble with a well-developed foliation or cleavage, along which the rock tends to split rather easily when struck with a hammer. The stress-strain curves for two experimentally deformed specimens of such marble are shown in Figure 21; one has been compressed perpendicular to the cleavage (90°) and one parallel to the cleavage (0°), both under 2000 bars confining pressure. The difference in sustained differential stress between the two specimens with increasing strain, as a consequence of the differences in strain hardening, is also indicated. The rock is 99.9 percent calcite; approximately half of the grains are roughly equant and half are elongate with a length to width ratio of 2:1. The parallel orientation of the elongate grains produces the foliation.

Both specimens were taken from the same block of marble; hence, they had been subjected to identical conditions during natural deformation. The cross section of the 90-degree specimen is circular, whereas that of the 0-degree specimen is elliptical with the major axis of the ellipse perpendicular to the cleavage. Moreover, the 90-degree specimen is appreciably stronger and shows sig-

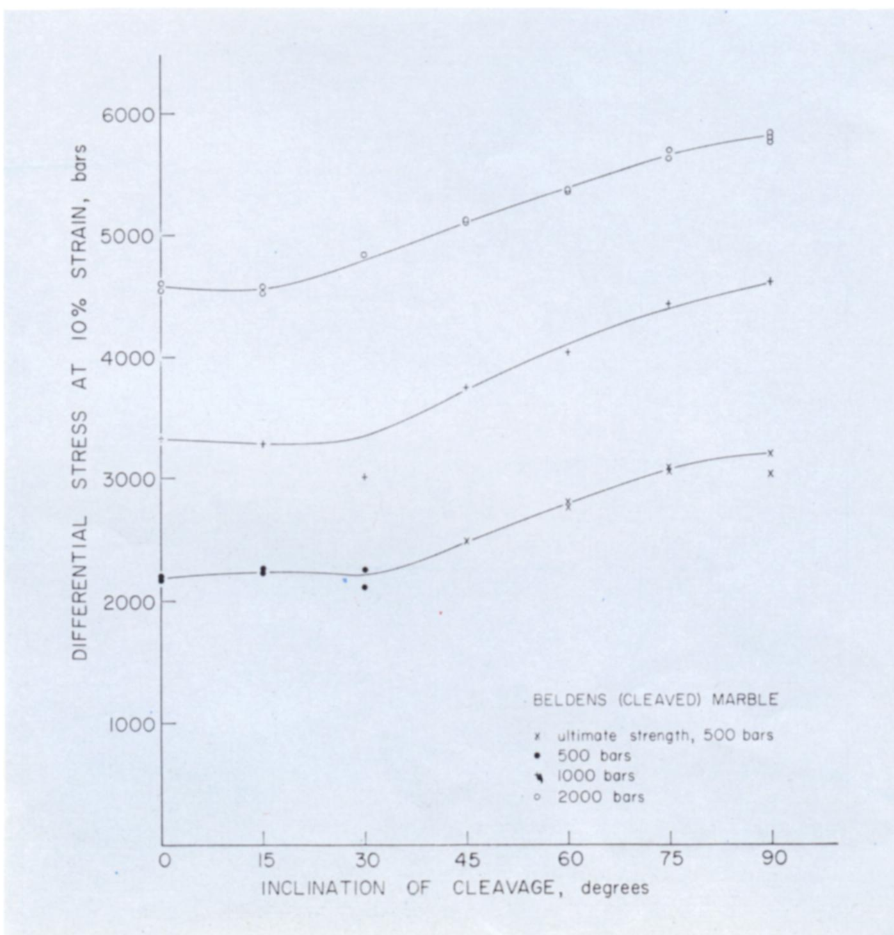


Fig. 22. Strength variation in Beldens marble as a function of the orientation of cleavage with respect to the direction of maximum compression. Curves indicate the general trends at each confining pressure.

nificantly more strain-hardening than the 0-degree specimen. Both lines of evidence indicate that the direction of maximum compression during natural deformation was perpendicular to the cleavage, a conclusion that is supported by field relationships in the area from which the marble was collected and by related work on the development of cleavage.

The sustained differential stress (or strength) of the marble increases systematically between the 0- and 90-degree orientations, as shown in Figure 22 for specimens deformed at three different confining pressures. The 30-degree specimen deformed at 1000 bars is weaker than what one might predict from the general trends because the specimen had an exceptionally low yield stress. This is not unusual for this orientation and can be particularly striking in rocks such as slate or phyllite.

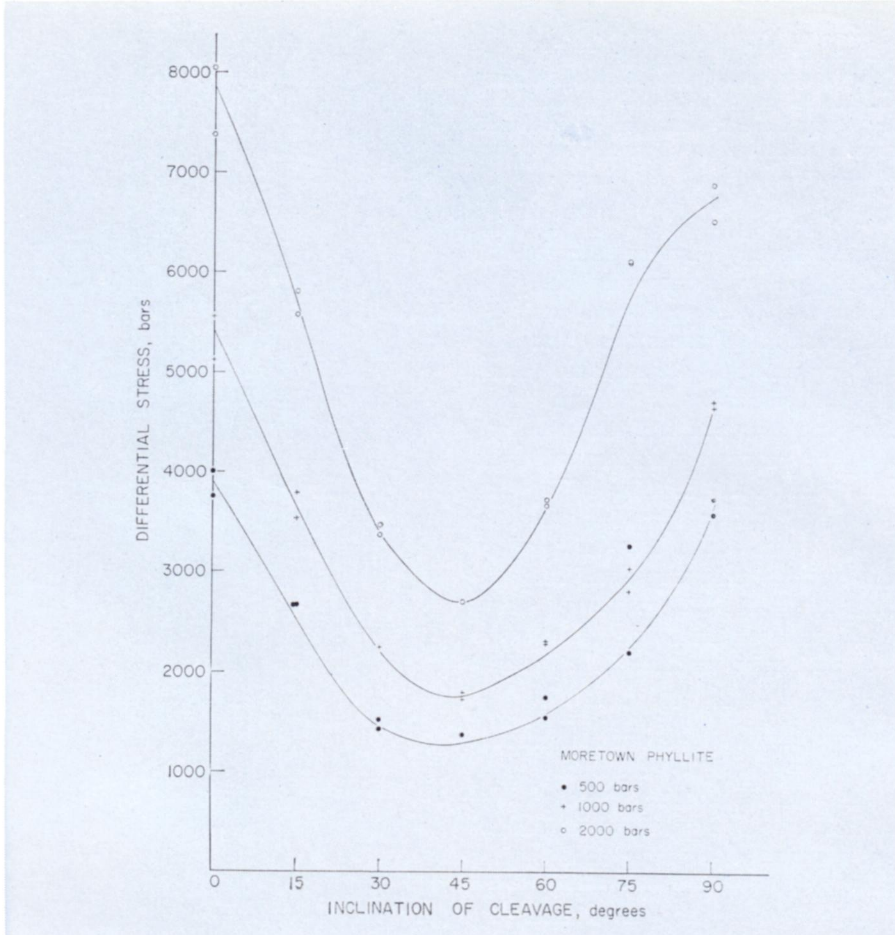


Fig. 23. Strength variation in Moretown phyllite as a function of cleavage orientation. Curves indicate the general trends at each confining pressure.

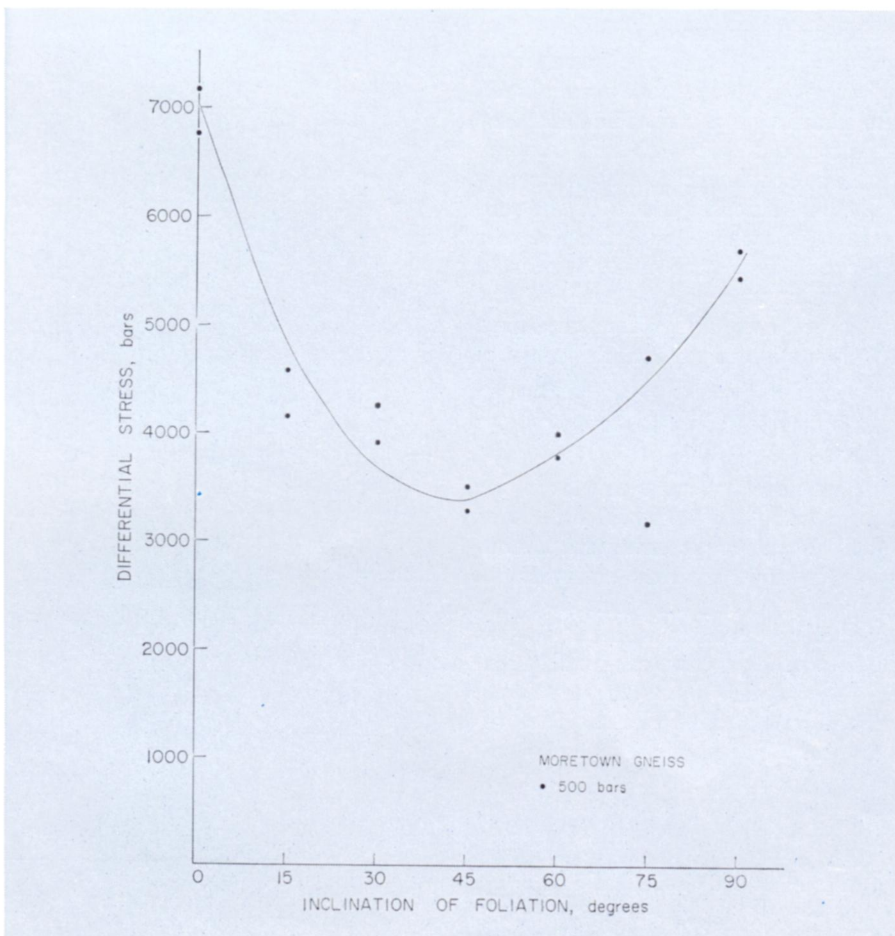


Fig. 24. Strength variation in Moretown gneiss at 500 bars confining pressure for different inclinations of the foliation to the direction of maximum compression. Curve represents general trends.

Effects of planar anisotropy

The conditions of homogeneity and isotropy generally assumed in theoretical analyses and commonly sought for experimental work are seldom realized in materials of the earth's crust. Most rocks in the upper crust are characterized by some type of foliation and thus are distinctly anisotropic. The presence of anisotropy can be expected to have a pronounced effect on both the strength and deformational behavior of rock.

The strength variation in a phyllite at three confining pressures is shown in Figure 23 as a function of the inclination of its foliation (cleavage) to the direction of maximum compression. The phyllite is composed of slightly over 30 percent platy minerals (muscovite, chlorite) and about 60 percent very fine quartz grains that are elongate parallel to the foliation defined by the platy minerals. The strength variation curves of Figure 23 are concave upward and roughly parabolic in form, much like curves obtained previously for slate (Donath 1961, 1964). As with slate, the curves are shifted upward with increasing confining pressure, indicating increases in strength with higher pressure. There are two interesting differences between the results for phyllite and those for slate, however: (1) the 0-degree orientation in phyllite is stronger than the 90-degree orientation, whereas the reverse is true for slate; and (2) the 45-degree orientation is weakest in phyllite, whereas the 30-degree orientation in slate is weakest.

That the 0-degree orientation in phyllite is strongest is not too surprising in view of the fact that there are more large mineral grains (higher granulosity) aligned parallel to the foliation in phyllite, as compared with slate. It is apparently easier to break across these grains in the 90-degree orientation than it is to break the grains longitudinally in the 0-degree orientation. If the degree of granulosity is responsible for this, then the effect should be even more pronounced in a schistose gneiss, which has a greater proportion of non-platy minerals. The strength variation in a schistose gneiss consisting of about 15 percent platy minerals (muscovite, chlorite) and 80 percent fine quartz grains, elongate parallel to the foliation defined by the platy minerals, is shown in Figure 24 for specimens deformed at 500 bars confining pressure. Although the strength of the gneiss is appreciably greater than that of the phyllite at 500 bars for each orientation, the strength differences among the 15- to 90-degree orientations are remarkably similar. Yet, the strength difference between the 0-degree and 15-degree orientation of the gneiss is much larger than the corresponding difference between the 0-degree and 15-degree orientation of phyllite, as we have anticipated.

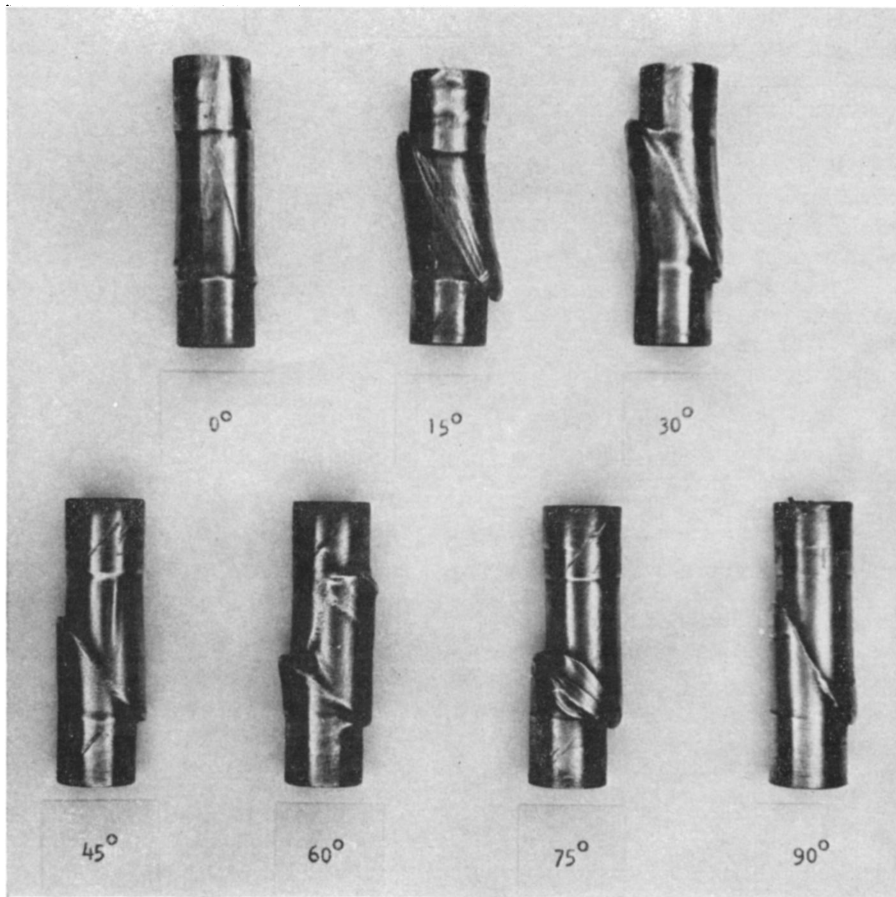


Fig. 25. Jacketed specimens of Martinsburg slate showing effects of anisotropy (cleavage) on mode of deformation at 500 bars confining pressure. (From Donath 1964)

Like the phyllite, the gneiss is weakest in the 45-degree orientation. Inasmuch as a plane oriented at 45 degrees to the direction of maximum principal stress has the highest possible shearing stress on it for any given state of stress, the observation that this is the weakest orientation is not too surprising. Perhaps more surprising is the greater weakness of the 30-degree orientation in slate. Presumably, this latter relationship is caused by "internal friction" within the rock which creates the most favorable conditions for failure on planes that have relatively high shear stress but somewhat lower normal stresses acting across them; generally, such planes are oriented more nearly at 30 degrees to the direction of maximum compression.

It was mentioned previously that the determination of principal stress directions in naturally deformed rock is of major importance in dynamic structural geology. Shear fractures are among the most useful common deformational features for this purpose (see Donath 1962a, 1963). Their relationship to the principal stresses is well established, and they develop on all scales from microscopic to regional. An ideal shear fracture forms parallel to the direction of intermediate principal stress and at about 30 degrees to the direction of the maximum principal stress. Any factor that affects the ideal relationship between a shear fracture and the principal stress directions is certainly of concern to the structural geologist.

Figure 25 shows a set of slate specimens that were experimentally deformed at 500 bars confining pressure. The deformation is clearly recorded in the thin-walled copper jackets that still enclose the specimens. (The jackets are used to prevent access of the pressure medium to pore spaces in the specimen during testing.) Fracture has occurred parallel to cleavage for inclinations of cleavage of 15, 30, 45, and 60 degrees to the direction of maximum compression; the cleavage has strongly affected the fracture orientation in the 0- and 75-degree orientations, as well. In the 90-degree specimen the shear fracture is oriented at 30 degrees to the direction of maximum compression.

The results from tests on slate specimens run at six different confining pressures are summarized in Figure 26. The inclination of the fault (commonly a shear fracture) is plotted against the inclination of the cleavage with respect to the direction of maximum compression. The points along the diagonal line represent failure that occurred parallel to cleavage. This is seen to be the situation for the 15-, 30-, and most 45-degree specimens. It is clear that the 0-, 60-, and 75-degree orientations are strongly affected. Only the 90-degree orientation is free of the effects of anisotropy; faults for that orientation are inclined at approximately 30 degrees to the direction of maximum compression, as would be expected in homogeneous, isotropic material. The information squeezed out of the rocks in this study warns us that whenever strong planar anisotropy is present in naturally deformed rock, we must be extremely cautious about inferring the principal stress directions from fracture relations.

The tendency for slip to occur along the foliation in anisotropic rock provides another possible means of evaluating environmental conditions in natural deformation. In general, any mechanism that operates in both experimental and natural deformation could be useful for such purposes provided that one is able to determine the range of conditions under which it can operate and that the range is sufficiently narrow as to be meaningful. In the least, one might be able to place some limits on the conditions that existed within the natural environment.

In west-central Vermont the rocks have been subjected to two obvious periods of deformation. The earlier one was characterized by flow and ductile behavior, with extensive development of cleavage in the folded rocks. Superposed on this was a second deformation of less ductile nature in which the previously formed cleavage was an important slip surface. Folding of the cleaved limestones and marbles in this second deformation was effected by slip along discrete cleavage planes, much like the flexing of a telephone book, rather than by extensive flow related to pressure gradients as in the earlier deformation.

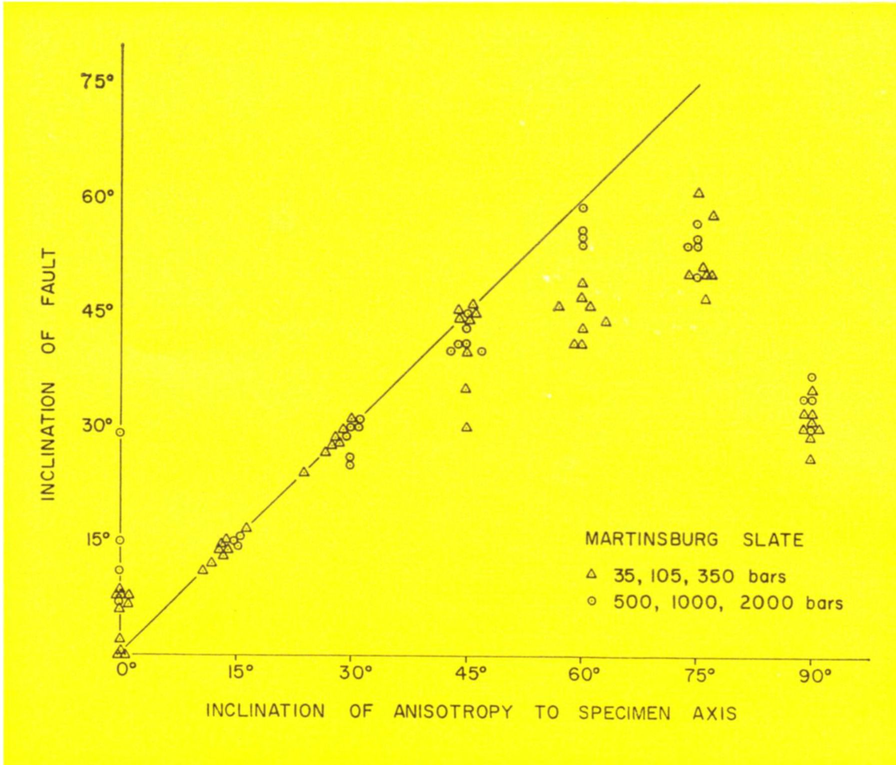


Fig. 26. Effect of cleavage on the angle of faulting in Martinsburg slate deformed at several confining pressures. (From Donath 1964)

It seemed reasonable to assume that the maximum confining pressure that would permit slip to occur along the cleavage in specimens deformed in the laboratory could be considered to be equivalent to the maximum confining pressure that could have existed in the natural environment during the later deformation. Accordingly, specimens were cored from a single block of cleaved limestone at 30 degrees to the cleavage, the orientation that most favors slip along the cleavage in this rock. The specimens were then deformed at confining pressures from 100 bars to 1200 bars in 100-bar increments in order to observe any changes in the mode of deformation caused by pressure change. Slip occurred along the cleavage in specimens deformed at pressures up to 400 bars; at 500 bars and above, failure was by flow subparallel to the cleavage. The maximum confining pressure of 400 to 500 bars suggested by these results agrees closely with pressure calculations based on the estimated thickness of the overlying rocks at the time of deformation.

The effects of strain rate and temperature do not have to be evaluated in order to establish the maximum pressure that could have existed. Increased temperature or decreased strain rate, which would represent more accurately the natural conditions, would both contribute to increased ductility of the limestone, and would thus require that deformation take place at lower pressures for slip along the cleavage to occur. A temperature determination utilizing oxygen isotope techniques indicated the temperature at the time of deformation to be about 400°C. This would suggest that the actual confining pressure was somewhat less than the maximum inferred from the pressure-mode data.

It is interesting to note that even at 1200 bars confining pressure the cleaved limestone did not exhibit symmetrical flow; the direction of flow was still strongly influenced by the anisotropy (cleavage).

Kink bands as dynamic indicators

In strongly anisotropic rocks the modes of deformation may differ significantly from those discussed for homogeneous, isotropic rock. One of the most interesting modes in anisotropic rock is the kink band. Kink bands are not only ubiquitous in naturally deformed slates and quite common in many other rocks with strong planar anisotropy, but they can be one of the more valuable tools in dynamic structural analysis.

Several specimens of the 15-degree orientation of slate which have been deformed at different confining pressures are shown in Figure 27. The numerous light gray parallel lines on the surfaces of the specimens—especially obvious in the specimens deformed at 800, 1000, and 1400 bars—are powdery material called gouge, which reflects the cataclasis that has occurred along discrete cleavage surfaces as a consequence of gliding on the cleavage. Each of the specimens shows a marked change in orientation of the cleavage from the initial 15-degree inclination to the compression axis. This reorientation is confined between two parallel inclined boundaries that are rather closely spaced in the 1800-bar specimen, rather widely spaced in the 1000-bar specimen, and located at the very ends of the 800-bar specimen. These boundaries, across which the attitude of the cleavage changes abruptly, are called kink planes; the area between them constitutes a kink band.

Some general characteristics of kink bands in experimentally deformed specimens of slate are seen in Figure 27. First, as already mentioned, there is evidence that slip has occurred along discrete cleavage surfaces throughout the specimen as a whole. Second, the gouge produced by cataclasis accompanying gliding is considerably more concentrated within the kink band, indicating that there was more activity along the slip surfaces within the kink band than outside it. Third, the width of the kink band can vary widely, although it may vary systematically; the width of the kink bands illustrated in Figure 27 decreases with increasing confining

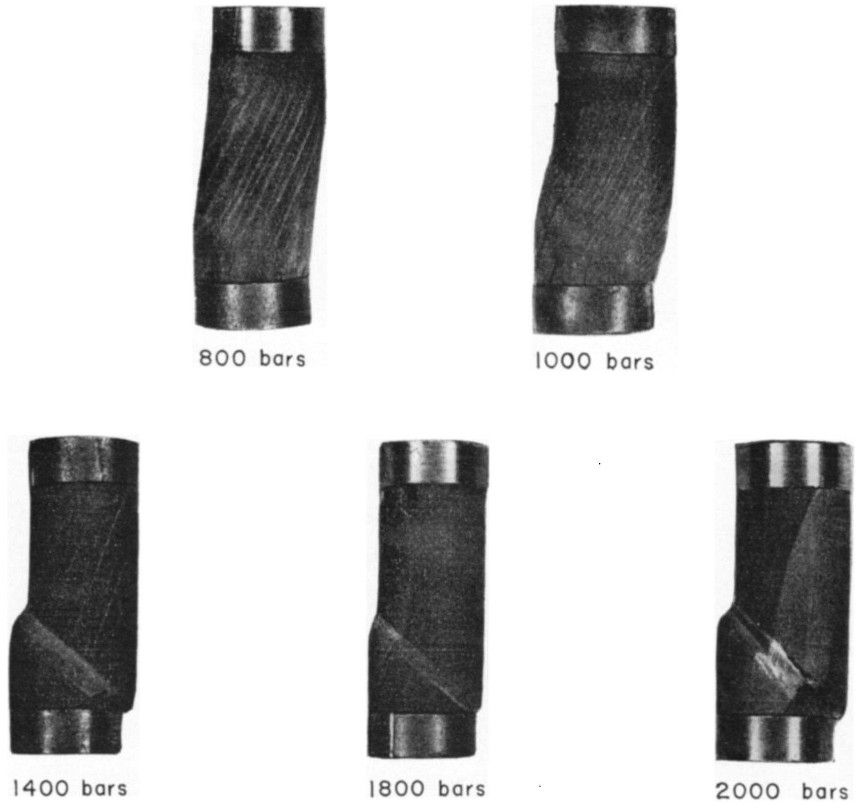


Fig. 27. Kink bands developed in 15-degree orientation of Martinsburg slate at different confining pressures. Specimens were initially ½″ diameter by 1″ length. (From Donath 1968)

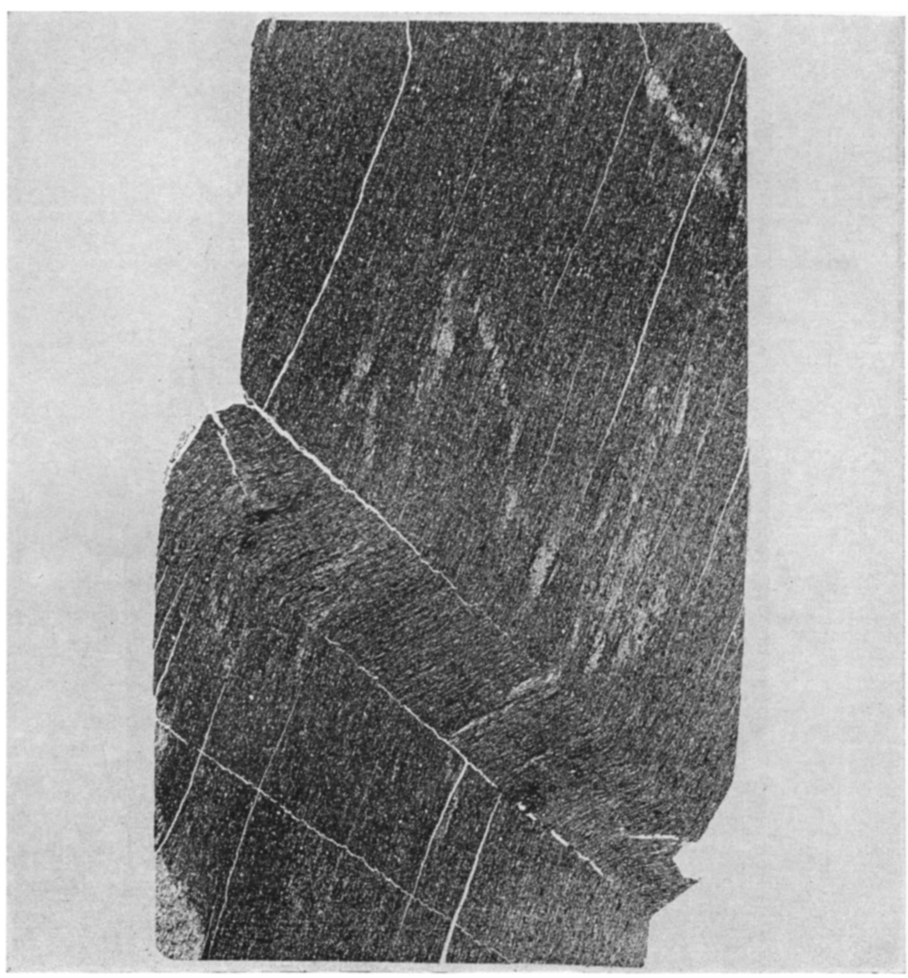


Fig. 28. Thin section of a kink band developed in the 15-degree orientation of Martinsburg slate experimentally deformed at 1600 bars confining pressure. The specimen was subjected to 8.7% total strain.

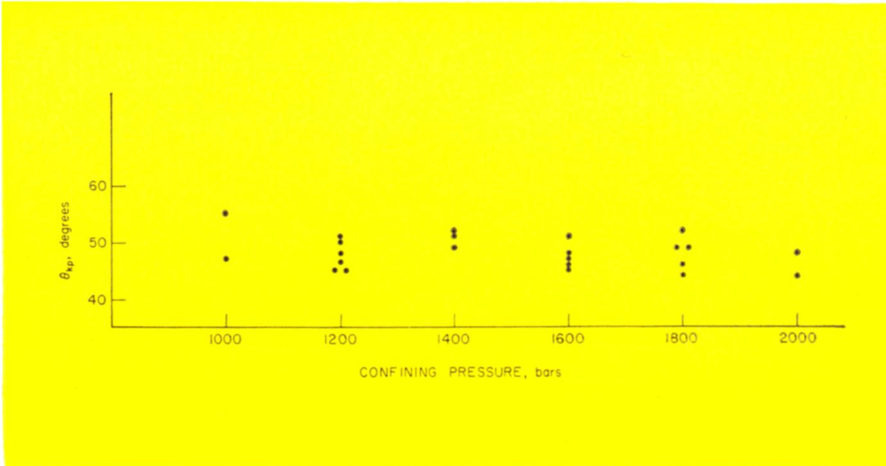


Fig. 29. Kink-plane inclination to the direction of maximum compression versus confining pressure for the 15-degree orientation in Martinsburg slate. (From Donath 1968)

Fig. 30. Natural kink band in Orwell limestone, Vermont. (Photo by G. W. Crosby)

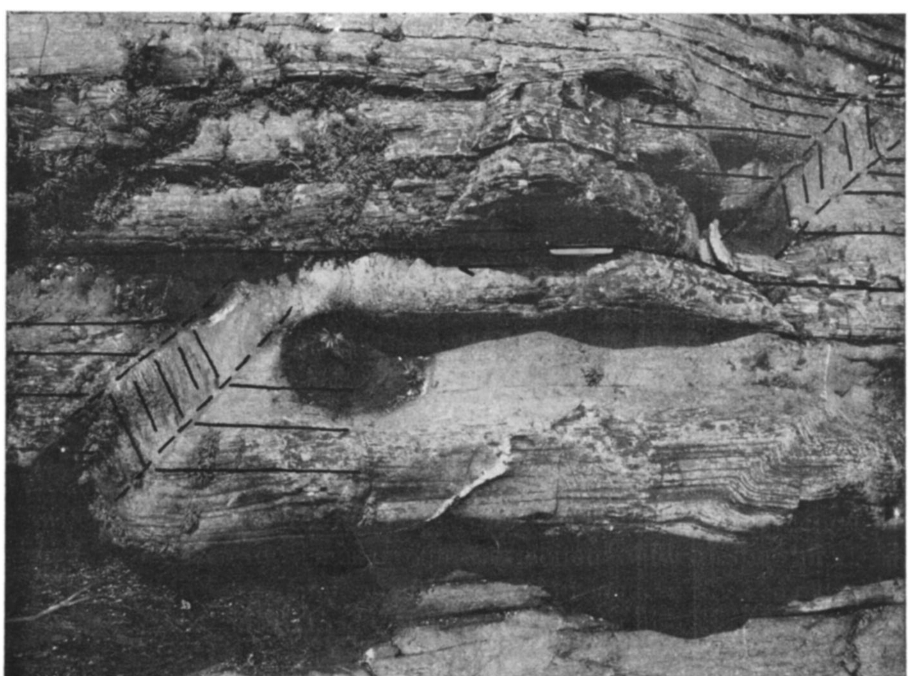


Fig. 31. Faulted kink band in slates of the Ards Peninsula, Northern Ireland. Knife placed along fault is 3″ long.

pressure. (Although it is not readily apparent from the photograph, two kink bands and a fault along one of the kink planes are present in the specimen deformed at 2000 bars.) Fourth, the inclination of the cleavage within the kink band may differ from specimen to specimen (compare, for example, the 1000-bar specimen with the 1400-bar specimen). And fifth, the angle between the kink planes and the axis of compression appears to be nearly constant regardless of kink-band width or confining pressure. The thin section illustrated in Figure 28 shows in more detail the character of a kink band developed at 1600 bars confining pressure.

The development of kink bands in experimentally deformed slate reflects the operation of several mechanisms—namely, gliding on cleavage accompanied by cataclasis, definition of kink planes along planes of high shear stress, and rotation of the foliation segments between the kink planes, with slip, cataclasis, and dilatation all occurring within the kink band (Donath 1968). Understanding the phenomenon allows one to interpret natural kink band relationships and thus determine the sense (relative direction) of slip along the foliation, the amount of shortening (or strain) produced by the kink banding, and—possibly the most useful application to dynamic structural geology—the direction of maximum compression.

The kink-plane inclination in experimentally deformed slate appears to be independent of confining pressure and, hence, other factors affecting ductility (see Fig. 29), initial inclination of the anisotropy, amount of rotation of anisotropy during the kinking process, and strain rate. In the slate specimens the kink planes are inclined at approximately 47 degrees to the direction of maximum compression. Since it might be inferred that, if no unusual boundary conditions exist, the intermediate principal stress would lie in the slip plane perpendicular to the direction of slip, the orientation of all three principal stresses can be determined from kink band relationships.

From the relationships seen in the kink band shown in Figure 30, we can say that each layer of limestone has moved to the left relative to the layer below, and that the direction of maximum compression lay somewhere within the acute angle formed by the kink-band boundaries and the layering, presumably at an angle of about 45 degrees to the kink-band boundaries. A geologist who did not recognize the phenomenon as a kink band, but considered it instead to be a subsidiary fold on the limb of a major fold, would infer that the movements were of the opposite sense—and he would be looking in the wrong place for the crest of the fold! As for determining the direction of maximum compression, it would be very difficult without the kink band relationship, inasmuch as slip along planar anisotropy can occur at any inclination to the direction of maximum compression up to about 60 degrees (see Fig. 26).

Other insights into the deformational history of an area are sometimes gained from the study of kink band relationships. The kink band shown in Figure 31, for example, tells us that the rocks have been subjected to at least two stages of deformation, and that the sense of relative displacement was reversed between the two stages. This conclusion can be drawn because each layer must have moved to the left relative to the layer below for the kink band to form; yet, the portion of the kink band at the top of the photograph has been displaced to the right relative to the portion at the bottom, subsequent to the formation of the kink band. Apparently, even kink bands have their faults!

One additional relationship observed among experimentally produced kink bands in slate holds promise as a natural pressure indicator. It was mentioned above that the width of the kink band (more correctly, the length of the foliation segments between kink planes) in Figure 27 decreases with increasing pressure. Data obtained from these specimens and from several others cored from the same block of slate are shown in Figure 32. The relationship is so nearly linear between 1200 and 2000 bars confining pressure that one could determine from measurement of the foliation segment length alone the confining pressure under which the deformation occurred, to within 50 bars! Obviously, if such a relationship held in natural deformation, it could be extremely valuable as a means of evaluating the pressure conditions of natural deformation. Unfortunately, even before the effects of temperature and strain rate could be considered, it was found that the slightest difference in cohesion in the slate could destroy the relationship. Nevertheless, a general pressure relationship might be established if the effects of differences in cohesion and other complicating factors can be recognized and evaluated. This is only one of many challenging studies that are awaiting further investigation and which show promise of contributing greatly to our understanding of natural rock deformation.

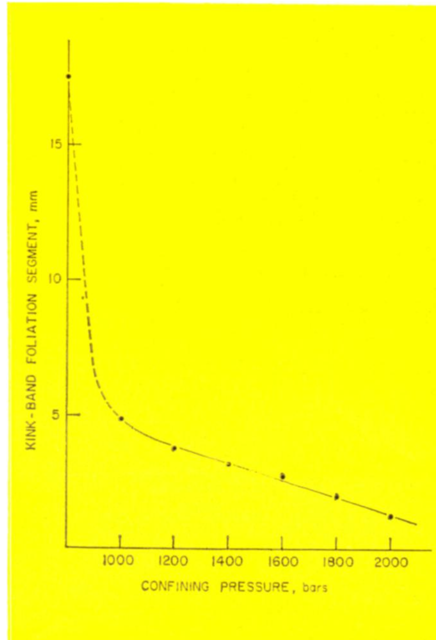


Fig. 32. Kink-band foliation-segment length versus confining pressure for 15-degree orientation in Martinsburg slate. (From Donath 1968)

Acknowledgments

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