FAULT ZONES, GOUGE AND MECHANICAL PROPERTIES OF CLAYS UNDER HIGH PRESSURE

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ABSTRACT

Clay minerals are found to be a significant component of all the fault zones, young or old, at surface or at depth down to 2 kilometers. These clays are stable up to 200 to 400 degrees under 1-4 kb of confining pressure. Thus, there are reasons to believe that in the part of the fault zone which earthquakes are found to occur we may encounter clay gouge.

Mechanically, the clays are quite weak compared either to rocks or to frictional contacts of rocks. The peak strength under 1-4 kb confining pressure of montmorillonite, a common gouge mineral, is 0.25 to 0.35 kb; its residual strength is 10-20% lower.

INTRODUCTION

Not being able to observe an active fault at the whole depth range, we have to use a conbination of methods to decipher the possible types of materials in the fault zones of interest; such information is necessary for the understanding of the mechanics of faulting. The methods at our disposal to investigate the fault zone materials are the same ones used by geologists and geophysicists to learn about the interior of the earth, taking into consideration that the temperature, the amount of fluid circulation, the presence of shear stress, the mechanical comninution actions, etc. are factors peculiar to fault zones.

Available to direct observation are outcrops of fault zones of different ages (the time since a fault ceases to be active), types, total displacement, in different country rocks, and in different stages of denudation. By surface or

drill-hole sampling of an active fault, we may look directly at the top portion of a fault regime; to date, the deepest part of the San Andreas fault we have seen is a little over 1000 ft (300 m) and plans are afoot to reach one kilometer in the near future; glimpses of San Jacinto at several hundred meters below the surface have been obtained during a tunneling project (Thompson, 1966); a tunnel that will cut through the active Bocono fault in Venezuela about 1 km under the surface is underway. Surface observation and sampling of fault zones has evidently to be interpreted with care, as weathering processes may alter the materials so much that they are not representative of the fault zone. Shallow drill hole or trench sampling may reach beyond the severely weathered zone but effects of circulating ground water still has to be considered. For both surface and drill holle (even down to a depth of a couple of kilometers) one may ask whether such materials are products generated at depths where earthquakes are taking place or art they intrinsically shallow depth fault materials. In regions that are undergoing rapid uplifting and erosion, the exposed zone may well represent the deeper part of the zone. Especially in a wide zone such as San Andreas in central California, we may also be seeing fault zone materials of different ages next to each other; that is to say, some part of the zone may be products of fault activity at depth. As we shall see later, the fault zone materials are probably weak and may flow; they could be squeezed upward in the fault zone (also see Wallace, 1949). It should also be remarked here that while surface observation of active strike-slip faults can conceivably reveal the nature of fault zone, the surface features of an active thrust may tell us only what is going on at the surface as the lip of the thrust is continuously covering new ground. Even along strike-slip faults, for example, the San Jacinto fault, the near surface trench across the active trace reveals a very limited zone with individual fracture planes and small clay filled gashes (Clark et al., 1972), at depth in the San Jacinto tunnel, the fault zone is extensive enough to form water barrier and is filled with rock fragments, sand and clays (Thompson, 1966).

As we have found out that in surface outcrops and in drill holes of active fault zones a significant part of the intrafault materials is what is commonly referred to as gouge, a mixture of fine grained sand, silt and particles of sizes less than 2 microns, i.e., clays of various kinds. Their ubiquitous and sometimes very dominant presence in outcrops and their particular mechanical behavior demand that we know whether they can exist at greater depth where earthquakes are generated. Unfortunately, we cannot find much clue about the T, P conditions of the mineral assemblages directly as these minerals are stable within the temperature range of 25°C to maybe 400°C, the range of interest. However, we can find to what extent these minerals can exist at the depth of 3-15 kilometers given the temperature, pressure and the chemical conditions. One can also use the familiar geophysical method of comparing the velocity and density of surface or drill-hole samples under pressure with those obtained through seismic and gravity measurements over the fault zone (Wang et al., 1978); although not terribly specific, we may estimate whether the materials in the fault zone are consistent with the kind of materials we find near the surface. Recently acquired reflection profiles across San Andreas seem even capable of showing gross structure inside the fault zone (COCORP; Long et al., 1979). Of no less significance is the fact that we know certain behaviors of active faults, e.g., creep, sudden stress drops, and the placement of fault trace within a fault zone, and we have some estimates of the order of magnitude of tectonic

stresses; whatever exists at depth, should also be able to reproduce these behaviors and should have strengths that are consistent with the estimated tectonic stresses available to generate earthquakes.

Knowledge of active fault zones can also be augmented by observing at surface, in deep tunnels or drill holes, fault zones that have ceased to be active and has since been uplifted and eroded so that they are now exposed. These fault zones are especially useful in knowing what might exist beyond the depth at which the shallow mineralogy in active fault zone are in equilibrium. A different sort of interpretation problem is involved here. For example, during the long period of quiescence, the fault zone is subjected to the lithostatic stress with little or no shear stress; thus the zone is under "consolidation", the water would have diffused out from the fault zone, carrying with it some of the more mobile ions; it is also possible that if the gouge was in a saturated state, its mobility may result in the depletion of gouge with respect to the breccia. The post faulting burial may also lead to prograde metamorphism. For mylonites that were generated at depth where metamorphism takes place, it would be necessary to know what changes the original fault rocks had gone through in order to reconstruct the fault zone. It is obvious the older fault zones can be used as analogues to the active fault zones after the extraneous factors are removed.

In this paper, we shall first describe the mineralogy and macroscopic appearances of fault zones of several ages that we have studied so far, discuss the stability field of the various clay minerals detected in the fault zone, report on the initial results of triaxial mechanical experiments on whole clay samples and finally make a conjecture on the processes involved in fault zones, their formation, the subsequent actions and the generation of earthquakes. Obviously, we are only dealing with fault zones at the depth range of 0 to about 12 km, beyond which cataclastics may be replaced by mylonites and flow may be the dominant mode of deformation (e.g., Sibson, 1977).

FIELD OBSERVATION AND SAMPLING OF FAULT ZONES

Fault zones in three areas of quite different ages, and in different tectonic settings are studied with respect to their mineralogical compositions and mode of deformation. (1) The two normal faults in southeastern Adirondacks were probably active in late Paleozoic (Fisher et al., 1971) with Pre-Cambrian metamorphic rocks, anorthosites and Paleozoic dolomite in the vicinity. The deformations in these fault zones are in the brittle range; cataclasis is prominantly seen but neomineralization in the fault zone is an important aspect. (2) Several normal and strike-slip faults are encountered in the silver-lead-zinc mines of the Coeur d'Aleue district of northern Idaho (Hobbs et al., 1965). These faults were probably active during Cretaceous to early Tertiary; many of these faults form the fracture systems that control the placement of mineral veins. (3) In the San Andreas system, we have already described the fault gouge mineralogy of several surface outcrop (Wu et al., 1975; Wu, 1978), here we are examining the fault geometry, the mineralogy of several down-hole samples (down to 428 meters) in Dry Valley, California, the mineralogy in one outcrop in southern California that has not been described before and some reflection data across San Andreas.

I. TWO NORMAL FAULTS IN SOUTHEASTERN ADIRONDACKS, NY

Figure 1 shows the location of these fault zones and the geology of the immediate vicinity. The total displacements along these faults are not well known; they are probably in the range of five hundred meters to a few kilometers.

1) The Schroon River fault. This fault has a mapped length of over fifteen kilometers. An easily accessible and good exposure is seen on highway 9 in Warrensburg. The fault zone there has a width of about 20 meters with a "meta-anorthosite" and a "granitic charnockitic gneiss" on the west and the east sides of the fault zone, respectively.

Within a few tens of meters of the bocndaries of the fault zone, it is clear that the density of joints is higher than farther out, and the orientations of the joints become auite varied. Some of the joints near the fault zone have irregular Surfaces and are filled with a fine-grained greenish gouge. The gouge were evidently not generated in the joints, as the surfaces are still quite intact, with slickensides on fresh rock. The presence of the slicdensides in the gouge and on the joint surfaces indicate that movements along the joints have taken place.

Going into the fault zone, we see breccias of various sites; some of them are still quite angular, although they seem to have undergone substantial rotation and translation. The breccia are surrounded by gouge (Figure 2). It is quite evident that the gouge was not generated by grinding up these angular breccia, rather the angular breccia may have been plucked off from the nasi and mixed in with gouge. The breccia are mainly gneissic ones; meta-anorthosite breccia are rarely seen.

The original mineralogy of the wall rocks and the production of alteration in the fault zone are summarized in Table 1. Figure 3a shows an example of unaltered gneiss. Figure 3b shows the result of alteration of the ferromagnesium minerals; the long grains and chlorite streaks seem to be aligned parallel to the horizontal axis of the picture -- the chlorite seems to have healed a sheared rock. Figure 3c shows a sanple of what appears in hand specimen as dark green "gouge"; here chlorite dominates and a definite flow texture is obvious.

2) The I-87 fault. This fault also has a mapped length exceeding 15 kilometers. An excellent outcrop is seen along highway I-87 in southeastern Adirondacks (Figure 1). The width of the fault zone is about 30 meters. Here the Ordovician Little Falls dolomite and a Pre-Cambrian metagranite are on the west and east side of the fault zone, respectively.

The metagranite is very similar to the granitic gneiss at the last site (Table 1), and the alteration products are essentially the same. The dolomite breccia is limited to within a few meters of the western boundary of the zone. The fairly angular dolomite breccia is embedded in a fine grained matrix -- showing no sign of having cndergone cataclasis in hand specimen -- that is composed of dolomite, quartz and feldspar (Figure 4). Dolomite grains in the matrix seem to fill in all the spaces around the quartz and feldspar grains.

In one part ofthe fault zone, angular breccia of metagranites from apparently different points of the fault wall (different color, grain size and texture) are next to each other with a small amount of gouge among them to fill in the space. The breccia have not been rounded much. It appears as though the gouge, relatively incompetent had carried them about in the fault zone for some distance without rounding the breccia and much of the gouge had subsequently been squeezed out of the system.

DISCUSSION

In the fault zone alteration process a large amount of chlorite is produced. This implies that a good amount of water has to be added to the system to produce them. This presence of ample water can conceivably cause law effective stress in the fault zone. Since the chlorite-rich "gouge" is quite pervasive, they could control the behavior of faulting after the fault zone is formed. Whether chlorite was the mineral in the gouge at the time of activity or is it the result of alteration after the fault had ceased to be active cannot be determined.

II. FAULTS IN DEEP MINES OF NORTHERN IDAHO

Hobbs et al. (1965) have mapped in some detail the faults in Coeur d'Aleue district of northern Idaho (Figure 5). The main fault of the region, the Osburn fault, is a right-lateral strike-slip fault that trends approximately N 70° W and has an estimated total displacement of 25 kilometers. The width of the fault in surface and mine exposure ranges from about 30 meters to 270 meters (Hobbs et al., ibid.). During the Summer of 1978, we could gain access to only one closely timbered tunnel across the Osburn fault; many previously open tunnels across the fault have been damaged due to squeezing ground. Sampling has been carried out across the Osburn as well as a number of smaller faults ringing from less than a meter to about 10 meters in width. The clay mineral contents of the wall rock, breccia and the gouge have been determined. In the sampling process, we have also observed the variations in the fault zone geometry and the internal structures.

A. Fault zone geometry and internal structures.

Exposures of a fault at different depths along the dip and sometimes at several points along the strike allow us to see the fault in three dimensions. It is quite noticeable that the fault surfaces are seldom flat planes; the width may increase or decrease by a factor of 2 over a distance of a few fault widths. All the faults that we can observe (Table 2) in detail are relatively small (1-10 meters) and bounded by quartzite or argillite; the general features are quite similar; they include (1) sheared zone, (2) detached blocks, (3) tightly folded alternating layers of gouge and often crumbly quartzite or argillite, and (4) large lens of gouge. Figure 6 shows the Polaris fault zone at three levels in the Galena Mine. The zones are usually quite wet and the gouge is soft and easily deformed in hand.

B. Clay mineralogy of the wall rock and fault zone materials.

Table 2 lists the fault zone, sample descriptions and mineralogy of the less than 2 microns fraction.

The gouge samples contain chlorite and illite in all cases; these are actually the dominant minerals of the Belt series rocks of this region, and although the relative amounts of these minerals in the Belt series vary considerably, often the gouge is found to be relatively richer in chlorite. Of more interest is the presence of smectites, the expandable clays, in several fault zones. The wall rocks are completely free of such minerals in most cases.

III. SAN ANDREAS FAULT ZONE

Detailed maps in the San Andreas fault zone often reveal that the zone is variable in width. As shown in Figure 7, the faults on the two sides of a valley are sometimes parallel but more often than not curved or seemingly composed of segments. The section from Pearblossom to San Bernardino, however, shows a number of branches nearby. But as these are surface observations, we are by no means certain that the configuration of the fault zone at depth will correspond to that suggested by the surface trace. It is interesting to observe however, the curvature of the bounding faults seem to be related to topography; in the Point Reyes section, the divergent faults at two ends of the Olema Valley are perhaps the cause for the formation of topography lows there. Also, a COCORP seismic reflection profile across the San Andreas fault zone hear Parkfield indicate that the fault zone at depth is about 2 kilometers wide coinciding with the width defined by the surface trace of the two faults.

The gouge in the San Andreas fault zone has been studied previously (Wu et al., 1975; Wu, 1978). Clay mineral assenblages somewhat different from those in Idaho have been found in this zone. It is quite often thought that along the "great bend" of the San Andreas in southern California, there is little clay gouge (e.g., Anderson and Osborne, 1978). The lack of extensive gouge materials in some sections may be due to the predominance of thrust motion there; it could easily be seen that at the surface outcrop of a young thrust, the front is progressively advanced; the hanging wall will break up but no gouge is generated there. As shown in a tunnel crossing the Sierra Madre fault, a thrust, northeast of Los Angeles, the fault zone broadens at depths and substantial amount of intrafault materials, not unlike those seen in the deep mines of Idaho, are found (Proctor, 1970). West of Lake Elizabeth in the west part of the "great bend", a very substantial fault zone is exposed. It consists of severaly brecciated granite, altered in most cases, with gouge in fractures and also as quite massive lens. The outcrop is extensive enough to create a badland topography there. The gouge is found to be composed mainly of montmorillonite and some illite and chlorite (Table 3).

A recent drill hole in Dry Valley Just inside the San Andre,as fault zone at depths down to 280 meters. The side-wall samples are all clayey in appearance. Their densities, general description and less than 2 microns compositions are shown in Table 4. The mineralogy is not radically different from that in the surface crops of San Andreas or Hayward fault (Wu, 1978). The apparent decrease of montmorillonite with increasing depth is most probably not related to the stability of this mineral,, but due to the change of the pH of the solution (see next section).

The COCORP seismic reflection profile mentioned move indicate that the fault zone extends all the way through Moho, at a depth of 25 to 30 km; while the zone above 12 kilometers shows chaotic reflections, the zone below that level is more seismically transparent, i.e., more homogeneous (Long et al., 1979). Other seismic profiles of San Andreas under the Pacific north of Point Arena, California, also indicate a width of about 2 km to a depth of at least 4 km.

CLAY MINERAL STABILITY AT DEPTHS

The T, P stability range for each important clay mineral group is discussed below. Primary emphasis is placed on experimental data, although evidence from field studies is also incorporated into the discussion where appropriate. Special attention should be called to the importance of the extensiveness of Velde's experimental studies and to his important review of the literature in this field (Velde, 1977). It should be remarked here that the stability field of different minerals at the same temperature and pressure is controlled by the pH values (Roberson and Wu, in preparation, 1979).

Kaolinite

Several investigators (Perry and Hower, 1970; Muffler and White, 1969 ; Dunoyer de Segonzac, 1965) have observed the disappearance of Kaolinite with depth. Upper stability tenperatures can vary widely from 100°C to over 200°C . Several reasons have been suggested to account for the disappearance of Kaolinite with depth including conbination with Mg (Muffler and White, 1965), reaction with other phases to form illite and chlorite (Velde, 1969), or reaction with quartz to produce pyrophyllite.

Kaolinite has also been observed as a common constituent of hydrothermal deposits which have been interpreted to have formed at depths of several kilometers (Keller, 1963; Keller and Hanson, 1968; Lowell and Guilbert, 1970).

Kaolinite, a common constituent of many soil types, appears to be thermally stable from 25°C, 1 atm, to 100° to 250°C (+) at 1-2 kb. Its upper stability is primarily related to the chemistry of the host rock and reacting pore solution. Velde (1977) has pointed out that kaolinite, during epi-metamorphism, is incorporated into other phases due to a displacement of the bulk composition of the silicate system in which it is found, either through an increase in total R^{+} ion content due to chemical reduction of ferric ion or by the increase in availability of R^{+2}

ion to the silicate minerals through the destabilization after dolomite. Furthermore, Velde has stated that the existence of kaolinite in a mineral assemblage will not generally indicate the important factors of silicate equilibria, that is, the ultimate stabilities of the clays present.

Illite

Illite (defined herein as a dioctahedral, aluminous, clay size mica) is reported to be the most abundant clay mineral in argillaceous sedimentary rocks (Grim, 1968). Studies of deeply buried sediments (Dunozer de Segonzac, 1965; Perry and Hower, 1970; Weaver and Beck, 1971) indicate that illite contents increase with increasing temperatures and pressures. Experimental studies of the muscovite-pyrophyllite system by Velde (1969) show that an illite-like phase is stable to temperatures over 400°C at 2 kb pressure. Muscovite, according to Velde (1969), can persist up to over 400°C at 2 kb pressure.

Chlorite

Chlorite is a common constituent of argillaceous sediments. Chlorite and illite are dominant clays reported in rocks of Paleozoic age. Authigenic chlorites formed at relatively low temperatures have been extensively studied by Hayes (1970). Chlorites are also found in deeply buried sedimentary rocks as well as in low-grade metamorphic rocks. Metamorphic chlorites vary widely in composition. Upper stabilities will vary depending on the composition, among other things. Generally, chlorites will persist up to temperatures between 450°C to 5000.

Expanding Clay Minerals

Expanding clay minerals, as discussed herein include fully expanding clays such as montmorillonite and vermiculite, as well as partially expanding mixed-layer clays such as illite/smectites, chlorites/vermiculites, etc. Expanding clays can be highly aluminous in conposition (dioctahedral) or poor in aluminum and rich in Mg and/or Fe (trioctahedral). It is important to distinguish between the dioctahedral and trioctahedral varieties because their thermal stabilities are quite different.

Table 5 is a condensed version of a summary prepared by Velde (1977) which shows reported thermal stabilities of fully expandable clays and mixed-layer phases. It is inportant to note that the values given in Table 6 may be somewhat higher than the true thermal stability values of these clays. This situation arises because it is difficult to establish if equilibrium has been attained in hydrothermal synthesis studies of systems such as these. Reaction rates are so slow in these systems that it may be impossible to attain equilibrium over a period of weeks or months (times used for most of the runs that have been reported). Fully expanded and mixed-layer dioctahedral clays are stable to much higher temperatures if octahedral Mg is present or if Na or Ca (rather than K) predominates as the interlayer cation.

Although it is possible that Ca and/or Na in the interlayer will effect a much higher upper stability limit for dioctahedral smectites, it is .also plausible that the run products observed in hydrothermal synthesis studies were not at equilibrium. Roberson and Lahann (1978) have shown that both Na and Ca greatly reduce the reaction rate of smectite/illite.

The most spectacularly high values representing upper stability limits for smectites are those for the trioctahedral smectites (hectorite and trioctahedral beidellite). Iiyama and Roy (1963) report that these trioctahedral smectites are stable up to 800°C at 1-2 kb pressure. Eberl, Whitney, and Khouri (1978) recently showed that K-saponite is still fully expanded after heating between 300°-485°C for 34 days at 2 kb. Although there may be sone question as to whether equilibrium was obtained in these studies, it is clear that trioctahedral smectites can be expected to persist to great depths. Wilson, et al. (1968) reported that trioctahedral smectites have been observed in metamorphosed carbonate rocks in Scotland.

Studies of Gulf Coast Tertiary sediments by Perry and Hower (1970) show that mixed-layer illite/smectite persists to depths of over 16,000 ft and temperatures of up to nearly 200°C. The percentage of smectite in these mixed-layer clays decreases monotonically with depth. The character of the mixed layering also changes with depth. Random mixed layering gives way to an allevardite-like ordered mixed-layer phase at about 100°C. Studies by Velde (1969) of the muscovite-pyrophyllite system, allow us to compare stability ranges of mixed-layer illite/smectite phases in a controlled, sinplified, synthetic system with those of a comparable natural system. Figure 22 from Velde (1977) shows the phases in the muscovite-pyrophyllite join. Although the sequence of change is very similar to that observed in the natural system, the temperatures observed by Velde are considerably higher than temperatures observed in the natural system. For example, the upper stability of mixed-layer illite/smectite in the Gulf Coast is shown to be about 200°C, Velde's studies would indicate an upper stability of about 400°C. The higher temperatures observed by Velde suggest that equilibrium was not obtained. Alternatively, the simplified chemical system studied by Velde may be sufficiently different from the natural system to cause major differences in thermal stabilities. Velde (1977) prepared a depth-temperature plot of natural mineral assemblages for fully expanding phases, random and ordered mixed-layer phases, and an ordered allevardite-type phase. The data for Velde's plot were obtained by several different investigations in studies of sediments of Tertiary age and younger.

It is clear from the data available to date that mixed-layer illite/smectite, with an ordered allevardite-type stacking (with approximately 30% smectite layers) can be expected to persist to at least 200°C. Whether or not this phase can be expected to occur at greater tenperatures and pressures cannot be answered at the present tine.

Trioctahedral smectites and/or mixed-layer trioctahedral smectite/chlorite are reported in rocks that have undergone heating at depth. Closer examination of fine grained Mg-rich metamorphic rocks may show that this occurrence is not at all unique.

MCHANICAL PROPERTIES OF CLAYS UNDER HIGH PRESSURES

Knowing that gouges have a high proportion of clays and such clays can exist at depths of up to perhaps 12 km, it is logical to find out the properties of conponent clays under high pressures. Although the behavior of clays under differential stresses are of great importance in soil mechanics, most of the tests are performed under a few bars of confining pressures under drained or un-drained conditions. Recently experiments were performed by Summers and Byerlee (1976) and Shimamoto (1977) with clays sandwiched in between surfaces in tri-axial experiments. They found a drastic lowering of frictional strength with montmorillonite and vermiculite gouge in particular. In Sumners and Byerlee's (1976) and Shimamoto's (1977) experiments with many types of simulated gouge, the strength was determined by the gouge-rock contact. For the clay gouge in Summers and Byerlee's experiments, it is possible that the clay water is actually under partially drained condition, as the water in the clay can escape into the surrounding rock.

Since gouge is often massive enough such that the properties of whole clay are important in determining the behavior of the fault. For this reason we are performing experiments on cylindrical sanples of clays without rock plugs under undrained conditions. The sanples are prepared from clay powder by placing pouches of clays under confining pressure of 2-4 kb for 24 hours or more. The compressed clay is hard enough that it can be machined with ease. Similar to experiments in soil mechanics, the clay samples so prepared can underqo large strain but are still able to bear stress; quite often an experiment is-terminated because the available travel of the piston is used up, rather than because the sample has lost strength completely. Details of the experiment are described in papers by Wang et al. (1979a, 1979b).

Figure 8 shows three examples of stress-strain curves for montmorillonite. In the figures, ε_z represents axial strain as sensed by foil strain gages attached to the cylindrical wall of the sanple, ε_u is that conputed from external displacement gages. Notice the close agreement of these two values when ε_z is less than about 15 percent; when ε_z exceeds 15 percent, the gage may be affected by excessive deformation underneath it. ε_θ denotes the lateral strain and ε_v the calculated volume strain $\frac{\Delta V}{V}$.

Figures 8 show clear peak strengths and with further strain, residual strength. The curves resemble those for overconsolidated clay in soil mechanics. As consolidation refers to the decrease of pore fluid under confining pressure the material is undersaturated or that the effective stress is higher than a saturated clay under the same condition. The explanation offered in soil mechanics for such phenomenon is that after peak strength is reached the grains in a certain region gradually rotate to follow the same general direction, and the residual strength is the frictional strength of sliding along this zone. Under normal consolidation, i.e., in saturation, the barrier for grain rotation is essentially removed. The residual strength for the two states are found in soil mechanics to be indistin-guishable. Both the residual or "ultimate" strength and the peak strength are likely to be of interest in the fault zone if the gouge controls the fault behavior, as the gouge has undergone extensive straining beforehand, however, the slipping surfaces do not have uniform orientation.

The montmorillonite results are quite remarkable in that the peak strength varies only slightly with confining pressure especially for $\sigma_r > 3$ kb as summarized in Figure 9. This phenomenon may be related to the closing or pore space, and the rise in pore pressure of the water, therefore the near constancy of effective stress. The maximum shear strengths in this case are only 0.25 to 0.35 kb and are much smaller than those needed to activate a stick-slip with rock-on-rock contact (Byerlee, 1978), or stick-slip experiments with thin clay gouge layers (Summers and Byerlee, 1976). The residual shear strength is even smaller. Wang et al. (1979b) have also shown that illite, kaolinite and chlorite are somewhat stronger, of the order 0.6-0.9 kb.

Further tests are anticipated to investigate the influence of pore pressure, mineralogy, and temperature.

DISCUSSION

The formation of fault zones could involve several processes. In this section, we shall provide a conceptual sequence of the events in a fault zone based on field observations, rock mechanics, and intuition. Only the shallow part, i.e., above 12 to 15 km, of the zone concerns us here, assuming that flow rather than sudden slips is taking place below the level of 12 to 15 km (Sibson, 1977).

Major fault zones probably form by joining lesser weaknesses. As a result, the fault is likely to be irregular in shape to start with. The smaller irregularities or asperities will be broken off first, as the stress concentration there will be high; the broken rock will be caught up in the fault zone, and be further reduced in size as displacement along the fault continues. The amount of the intrafault materials should thus be related to the volume of asperities broken off. There may also be several parallel or subparallel weaknesses in the area, these may have movements on them initially, until a dominant fault forms, then the activities on these subsidiary faults would decrease.

Although the larger asperities may last for a time, repeated loading and rapid unloading and the stress concentrations on them may cause the formation of new joints. Jointed blocks near the fault zone could easily be "plucked" from the wall to go into the fault zone. The jointed rocks in the vicinity of the fault and the fault zone itself would form an excellent water conduit as well as reservoir.

The finely ground rocks have increased free energy and will soon be able to transform to new minerals that are consistent with the local T, P conditions, pH of pore fluid and overall chemistry. These are most probably clay minerals of various kinds. The large blocks caught up in the fault zone may also be altered to some extent.

The fault zone resulted from the processes described above is a complex system with subsidiary faults and densely jointed rocks nearly, a heterogeneous mixture of fault zone materials and the zone varies in width from place to place. Water may move around the fault zone under the influence of the stress-induced pressure gradients and continuous reaction of water with the fault breccia will further alter the mineralogy of the fault zone.

Based on available triaxial clay test data described in this report, we expect that a fault zone rich in expandable clays would have a shear strength of about 200 bars. With other clays, the strength will be in the range of 600-1000 bars. We should hasten to add here that temperature is one of the most inportant factors in the mechanical properties of clays. If higher temperature will weaken the clays or even lead to dehydration, then the numbers mentioned will be lowered correspondingly. The problem then involves the mechanisms for raising temperature and the rehydration of clays.

Curvature could evidently cause strength variation in the fault zone as well. If we have two asperities approaching each other (Figure 10), then the fault zone materials or even the asperities may nave to be sheared through new planes before the fault can move with enough displacement to relieve the accumulated strain. It is often wondered whether the strongest point or the weakest will be the starting point, the hypocenter, of a large earthquake. Obviously, the strongest point has to yield before sufficient displacement can be achieved. On the other hand, if the motion at the weaker points will cause the strongest point to yield, a large earthquake can also result from it.

With the strength differences at different points in the fault zone, the strain field will not be uniform after a major displacement event. Such non-uniformity and the time-dependent behavior of the materials will then create aftershocks.

CONCLUSION

Clay gouge has been found in fault zones of very different ages and at different depths down to a depth of 2 km. Based on available clay stability studies, we can expect the gouge clays to be able to exist up to tenperatures around 400°C for several dioctahedral clays and to 800°C for expandable trioctahedral clays; these temperatures correspond to depths to 15km or deeper.

Clays are quite weak mechanically, if they do exist extensively in mature fault zones, then we would expect low (~ hundreds of bars) shear stresses in faulting. For montmorillonite, the shear strength is as low as 200 bars. With higher temperatures, the strengths of clays can be lowered considerably.

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Table 4. Mineralogy of Dry Valley drill holes in central California in the San Andreas fault zone.

Table 5. Thermal stabilities of fully expandable clay phases (Velde, 1977).

Table 1. Wall Rock and Fault Zone Mineralogy of Adirondacks Faults

Wall Rocks and their Mineralogy			Fault Zone Alteration Mineralogy
Schroon River Fault	1-87 Fault		
Meta-anorthosite	Dolomite		hornblende → chlorite
plagioclase			biotite → chlorite
(andesine-oligoclase) partly sericitized			garnet → chlorite
hornblende			plagioclase → sericite
orthopyroxene [trace)			→ calcite + qtz
apatite (trace)			+ chlorite
biotite			→stilpnomelene
Charnockitic Granitic Gneiss	Metagranite		
quartz	quartz		
plagioclase (andesine)	plagioclase (01igoclase)	60-70%	
microcline-perthite	microcline		
hornblende	hornblende		
opaques	biotite		
apatite	garnet		
biotite	chlorite		
zircon	zircon		
chlorite	apatite		
sericite			
calcite			
garnet			

70-80%

Table 2

Percentages of Clay Minerals in Deep Mines In Idaho

	% Chlorite	% Illite	% Montmorillonite
Osburn Fault (21 Samples; 13 Gouges & 8 Rocks)			
Gouge	10.46 ± 10.86	84.08 ± 8.17	5.46 ± 4.50
Rock	6.25 ± 8.17	93.75 ± 8.17	0
Polaris Fault (23 Samples; 10 Gouges & 13 Rocks)			
Gouge	14.0 ± 7.13	86 ± 7.13	0
Rock	8.42 ± 3.23	91.57 ± 3.23	0
Cate Fault (15 Samples; 10 Gouges & 5 Rocks)			
Gouge	23.90 ±	70.9 ± 8.25	4.30 ± 2.50
Rock	11.60 ± 5.03	84.4 ± 8.29	2±0
Silver Syndicate Fault (13 Samples; 7 Gouges & 6 Rocks)			
Gouge	20.71 ± 8.14	73.29 ± 9.8	4.5 ± 4.35
Rock	15.67 ± 10.25	84.33 ±10.25	0
Morning - Independence Fault (16 Samples; 14 Gouges & 2 Rocks)			
Gouge	32.07 ± 19.70	48.85 ±21.90	18.78 ± 1'5

Table 3

Sample	% Kaolinite and/or Chlorite	% Illite	% Montmorillonite & Mixed Layer
TP 1	22	0	78
TP 2	0	0	100
TP 3	0	0	100
TP 4	19	0	81
Tp 5	22	22	56
TP 6	0	0	100
TP 7	0	0	100
TP 8	0	0	100

125	5	4	31	22	38	2.29
275	8	4	39	25	24	2.53
530	23	18	31	28	0	2.38
575	19	21	15	19	26	2.41
620	20	26	29	23	2	2.46
650	18	30	27	20	5	2.41
740	12	23	15	19	31	2.39
800	15	31	20	32	2	2.48
875	19	20	11	21	29	2.51
905	27	26	18	15	14	2.49

TABLE 4

Densities and clay compositions of the less than 2 microns fractions of gouge samples.

Table 5

Thermal Stabilities of Fully Expandable Phases

Type	Elements in 2:1 lattice	Reference	1-2Kb Pressure temp, °C
K dioct	(AlSI)	Velde, 1969	230
K dioct	(MgAlSi)	Velde, 1973	400
Na dioct	(AlSi)	Sand, et al., 1957 Koizumi and Roy, 1958	350-450
Ca dioct	(AlSi)	Chatterjee, 1959 Hemley, et al., 1971	300-500
Mg trioct	(MgSi)	Esquevin, 1960 Velde, 1973	<250
Mg trioct	(MgAlSi)	Velde, 1973	430
Na tricoct beidellite	(MgAlSi)	Iiyama and Roy, 1963	550
Na tricoct hectorite	(MgAlSiNa)	Iiyama and Roy, 1963	800

Thermal Stabilities of Mixed Layered Phases

K dioct	(AlSi)	Velde, 1969	400
K dioct	(AlSiMg)	Velde, 1972	430
Na trioct beidellite	(MgAlSi)	Iiyama and Roy, 1963	780
Na tricoct hectorite	(MgAlSiNa)	Iiyama and Roy, 1963	800

Figure Captions

Figure 1. Faults in southeastern Adirondacks, locations and geology.

Figure 2. Breccia surrounded by gouge in the Schroon River fault zone (from Farrar, 1976).

Figure 3. a -- unfaulted and unaltered gneiss (Schroon River Fault) (crossed nicols).
b -- Ferromagnesium minerals altered, almost completely, to chlorite. Chlorite forms streaks, parallel to the horizontal axis of the picture; the horizontal direction also is the preferred direction of the long axes of mineral grains (crossed nicols).
c -- Chlorite is the predominant mineral in this "gouge." Appears as dark green fine-grained rock. Some quartz and plagioclase appear as well rounded or subangular grains.

Figure 4. Angular dolomite breccia (edge marked by x) embedded in a fine grained matrix of dolomite, quartz and plagioclase.

Figure 5. Major faults in the Coeur d'Aleue district of northern Idaho (Hobbs et al, 1965).

Figure 6. Internal structures of Polaris fault at three depth levels in the Galena mine of Coeur d'Aleue district in northern Idaho.

Figure 7. Maps of different sections of the San Andreas fault zone; the locations of the sections are shown in the index map.

Figure 8. Stress and strain diagram for montmorillonite at 2kb σ_r is the confining pressure, σ_z the axial stress, ε_z, the axial strain. u, the percentage axial strain calculated from an external displacement transducer, ε_θ the lateral strain and ε_v the volume strain.

Figure 9. The peak strengths of montmorillonite clay as a function of confining pressures. The samples were pre-compressed, undrained, at 2kb.

Figure 10. When two "asperities" on a curved fault move into each other, new breaks through the fault zone will have to be made and wall rocks will have to be sheared off for further displacement.

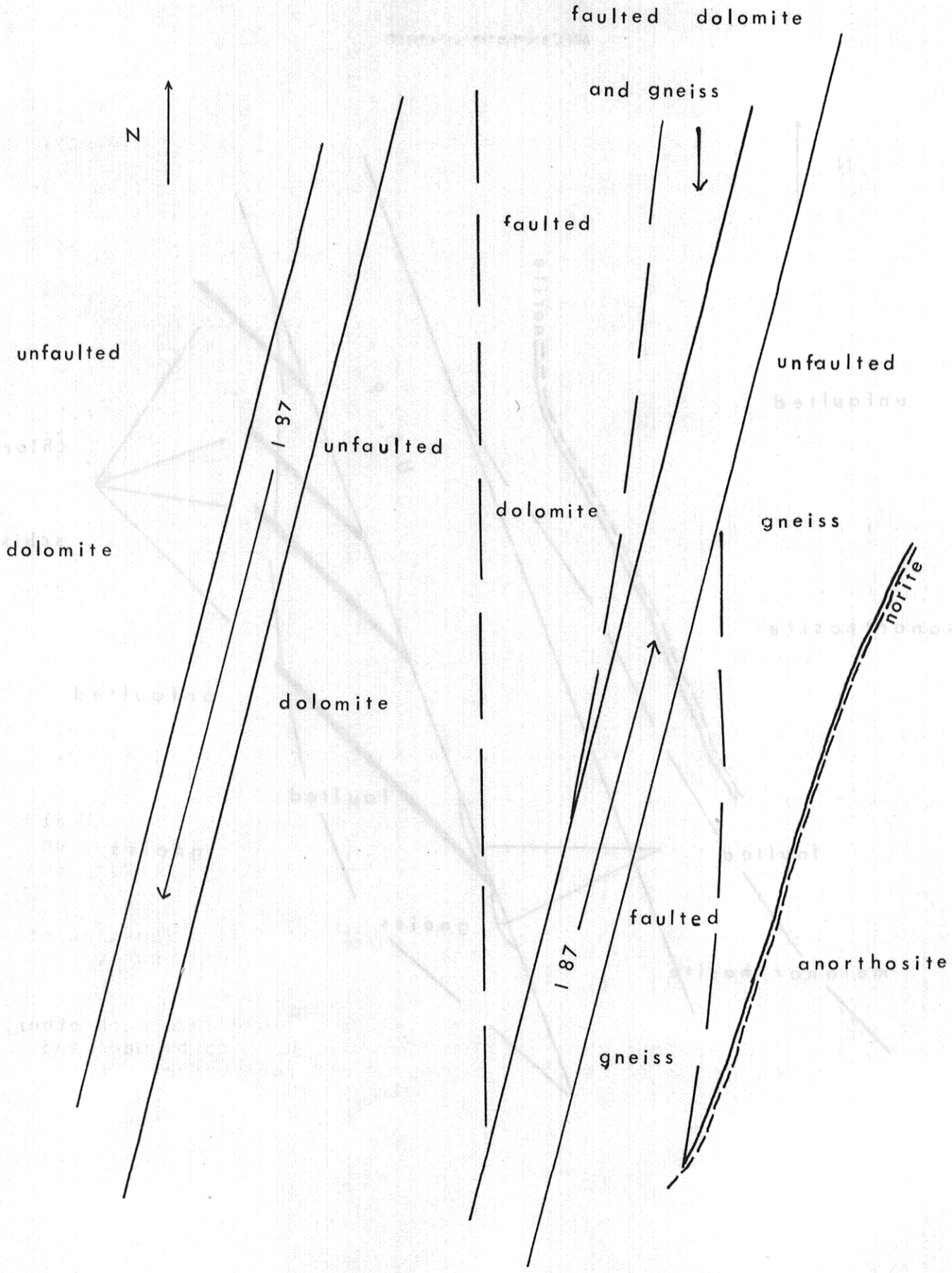
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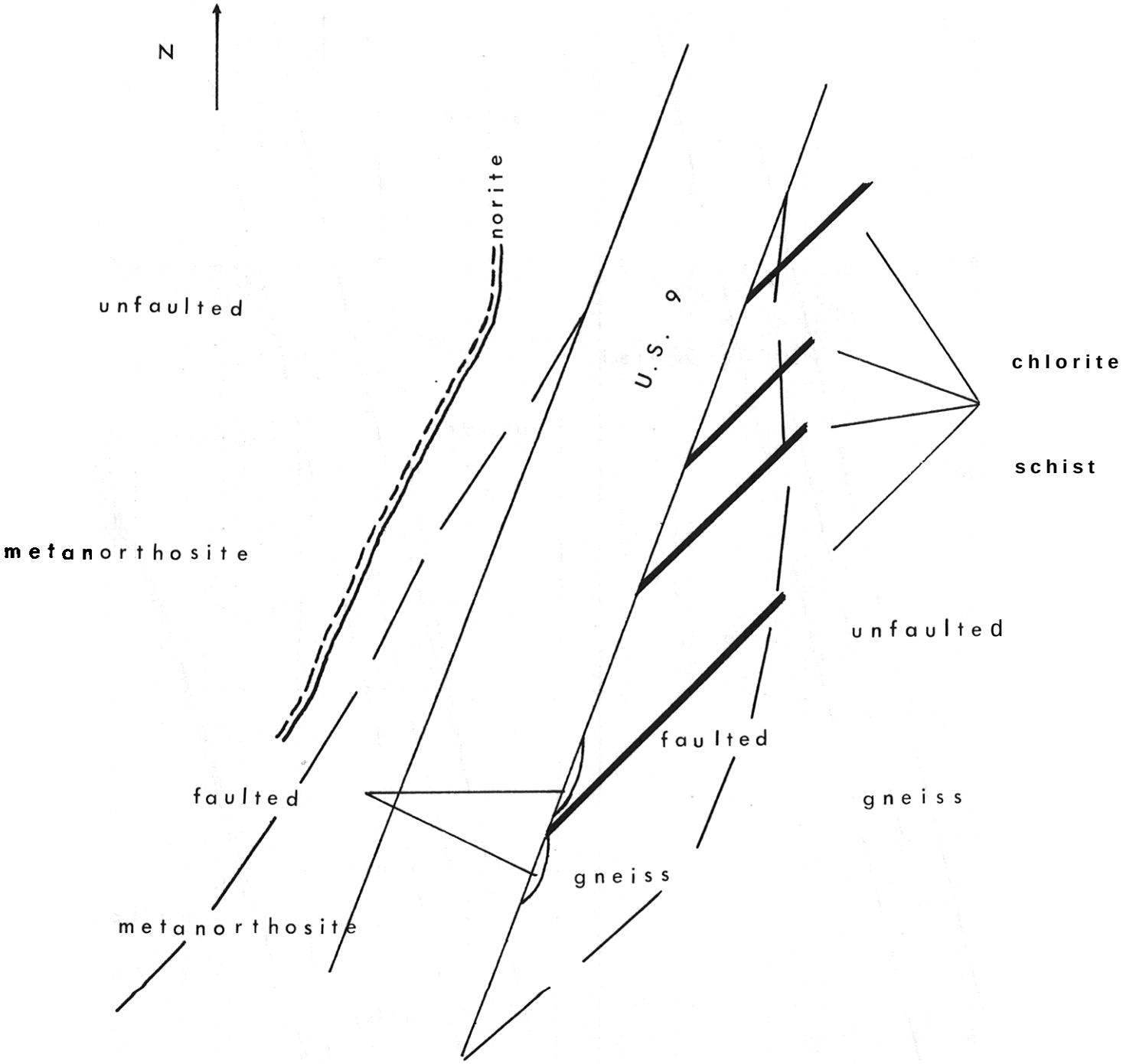
FIG 1a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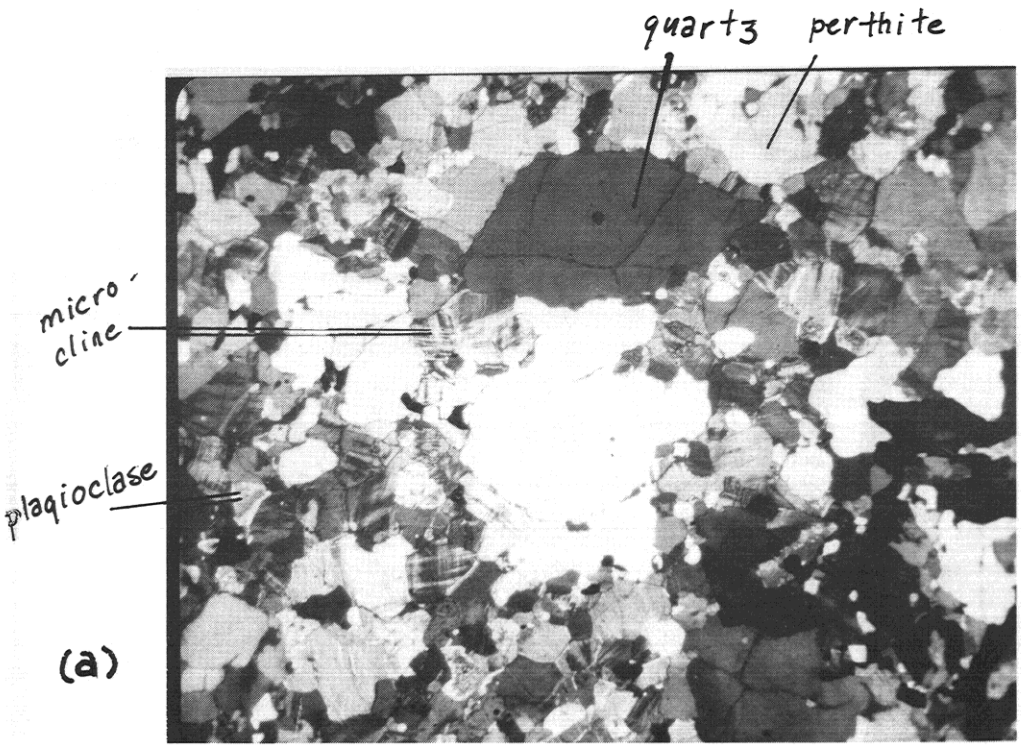
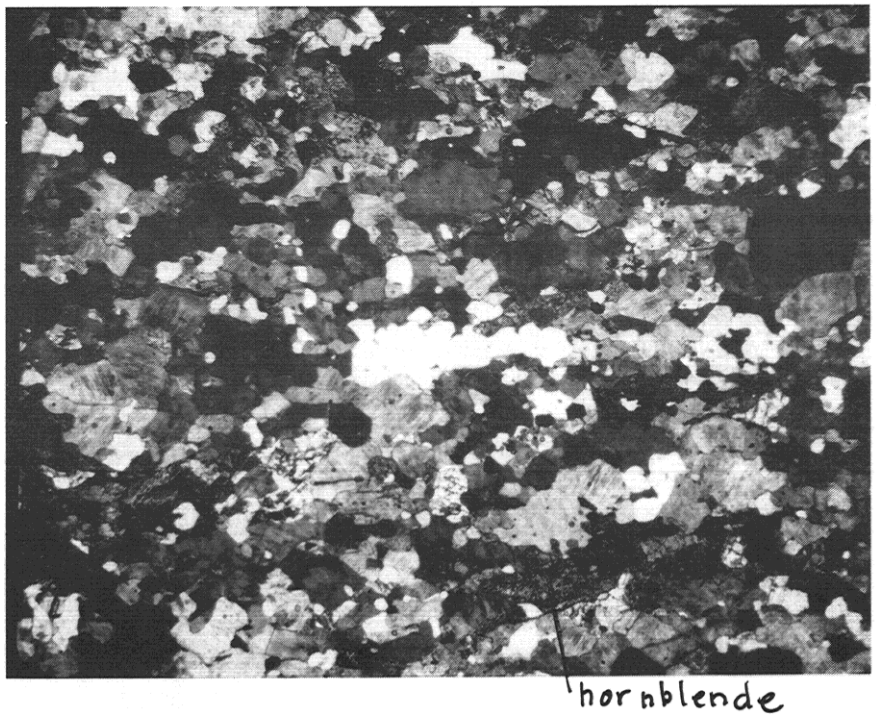
igure 65. Chlorite-calcite cemented breccia of charno

FIGURE 2





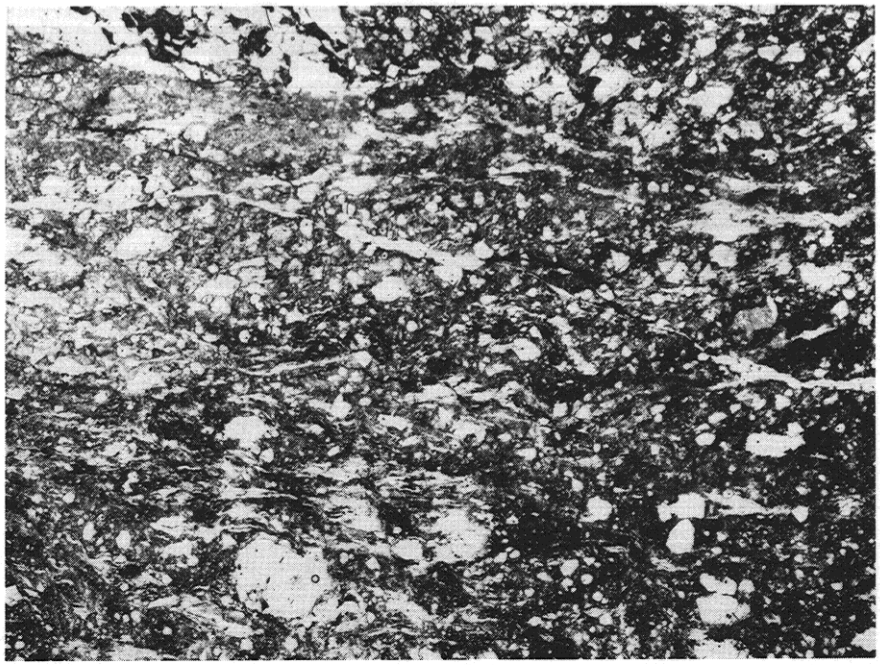


FIG 3

+ = edge of dol. clast

93, feldspar

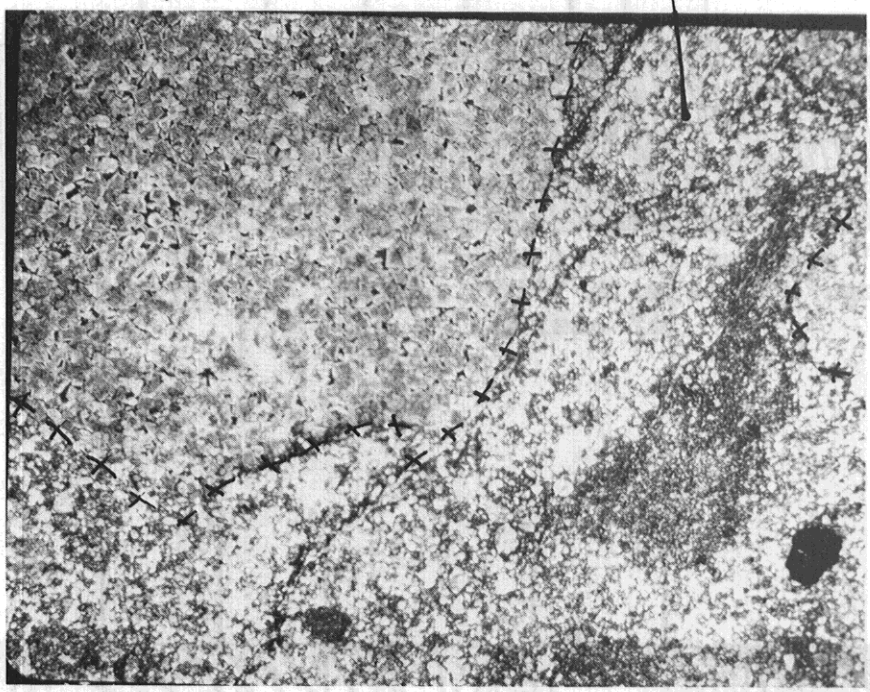


FIG 4

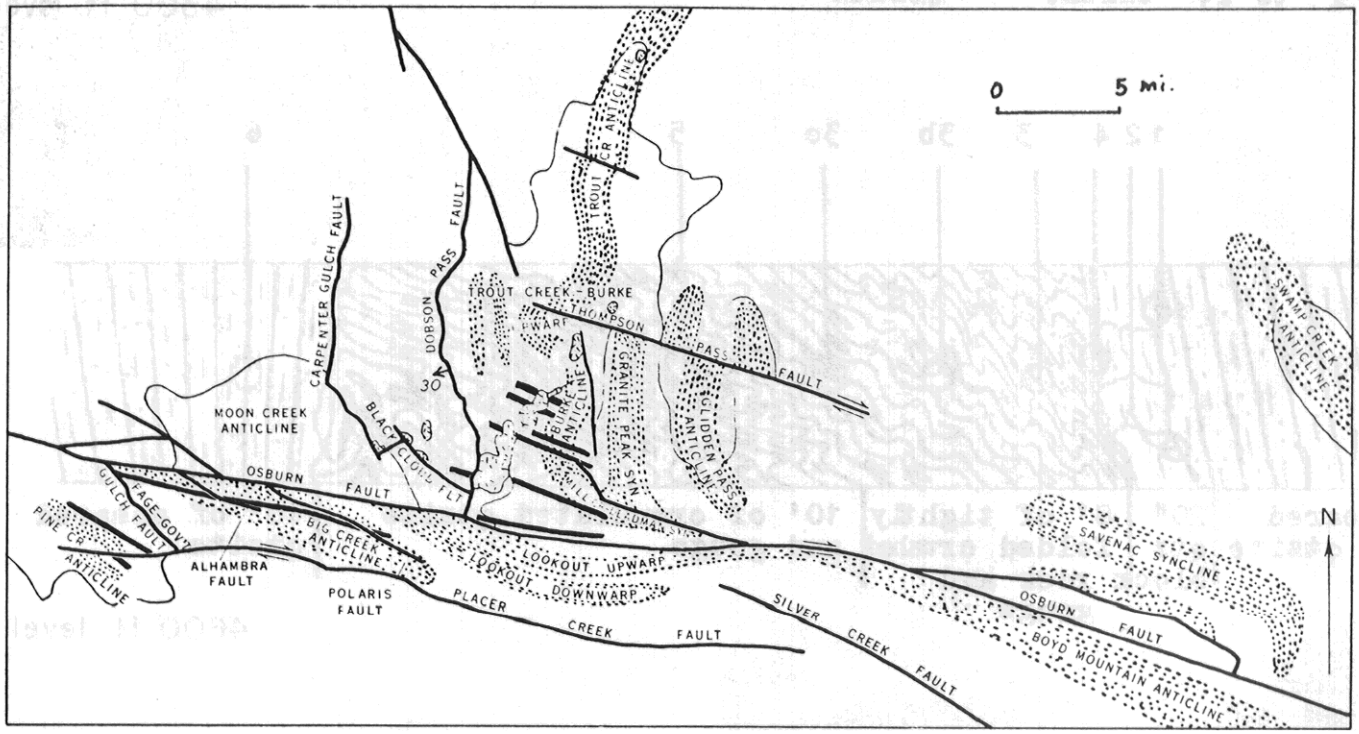


FIG 5

Galena Mine– Polaris Fault

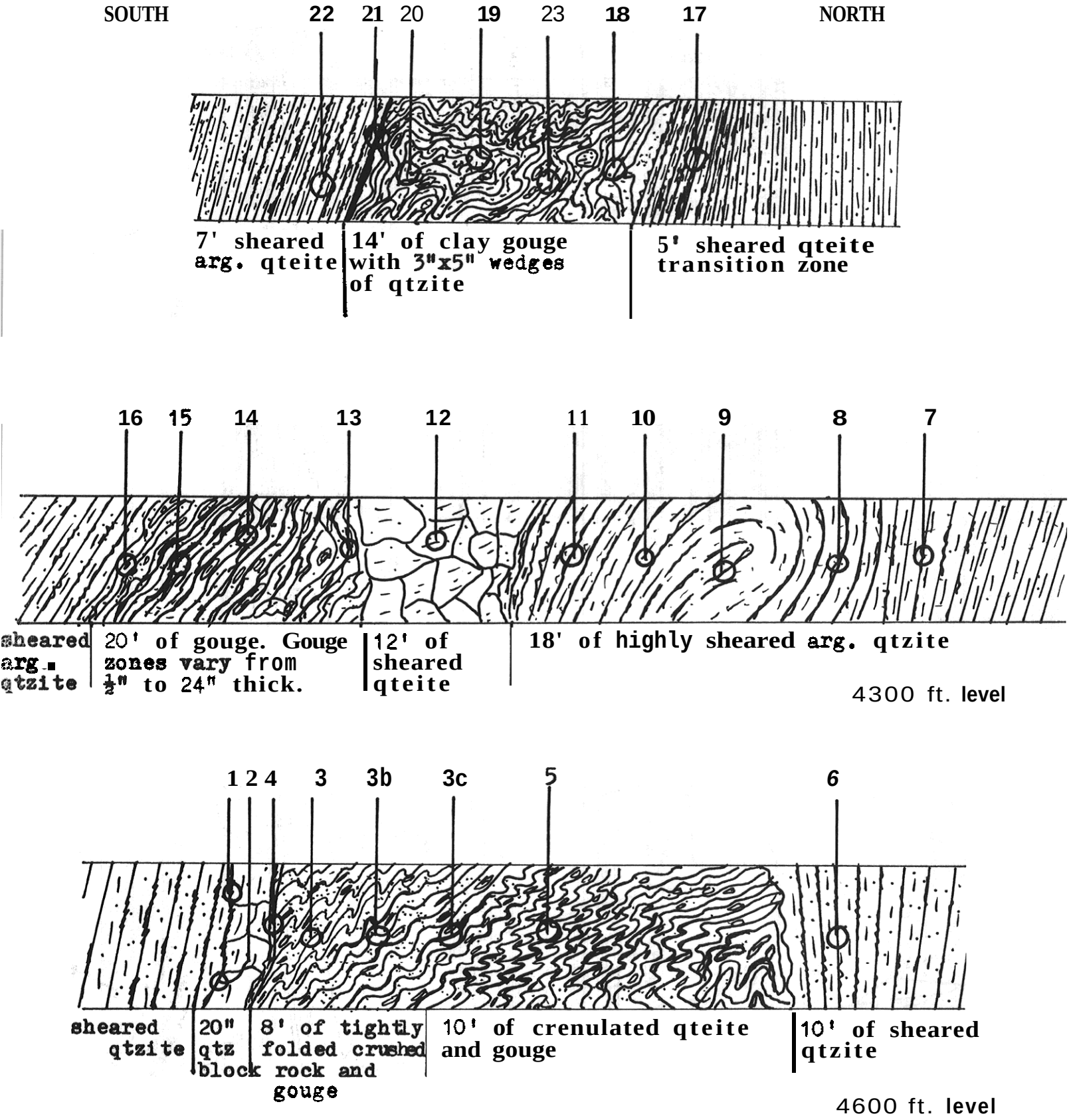
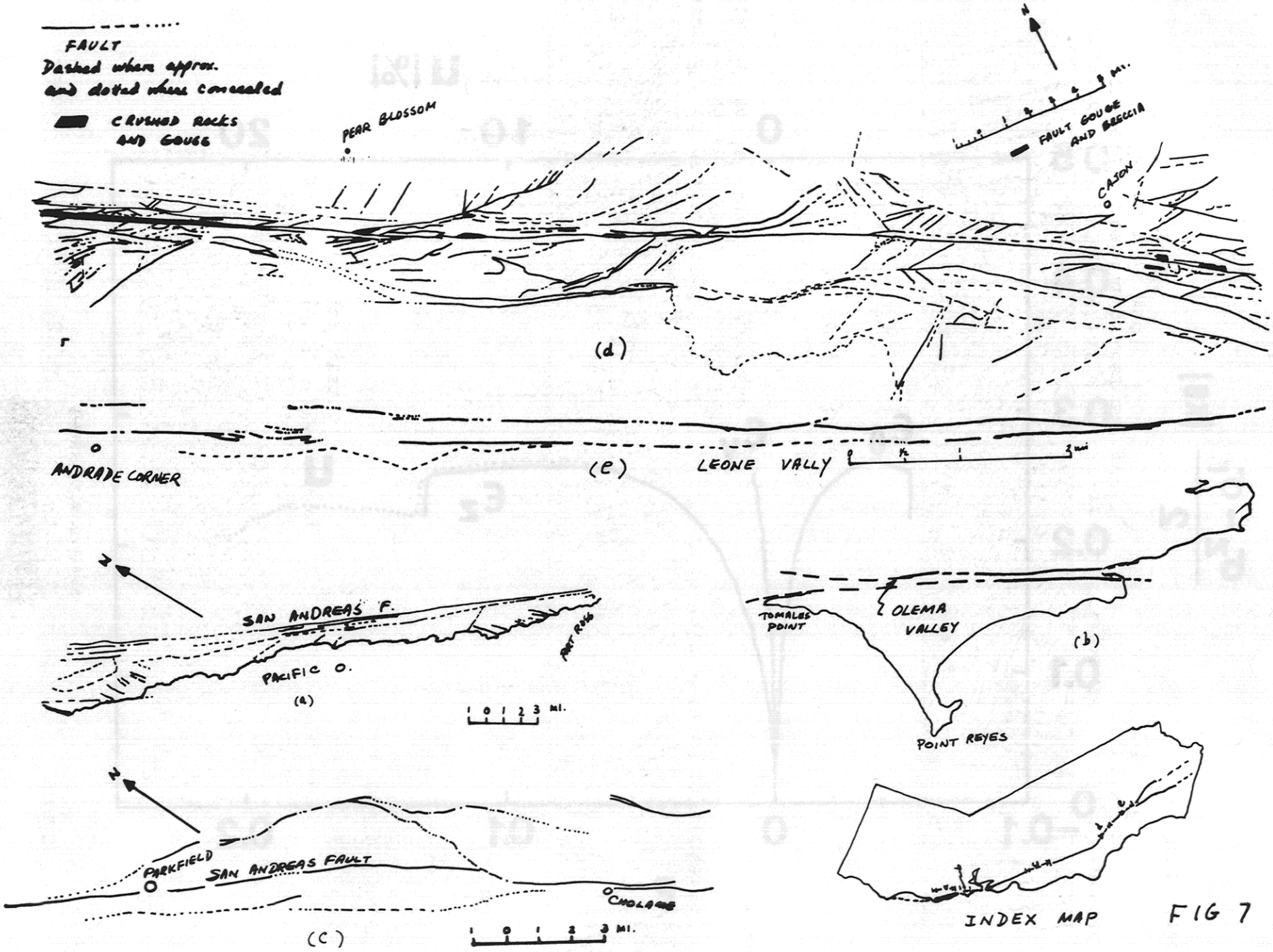


Figure 6


FAULT
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and dotted where concealed
CRUSHED ROCKS
AND GOUGE
PEAR BLOSSOM
FAULT GOUGE
AND BRECCIA
CAJON
(d)
ANDRADE CORNER
(e)
LEONE VALLY
SAN ANDREAS F.
PACIFIC O.
(a)
TOMALES
POINT
OLEMA
VALLEY
(b)
POINT REYES
PARKFIELD
SAN ANDREAS FAULT
CHOLAME
(c)
INDEX MAP

FIG 7

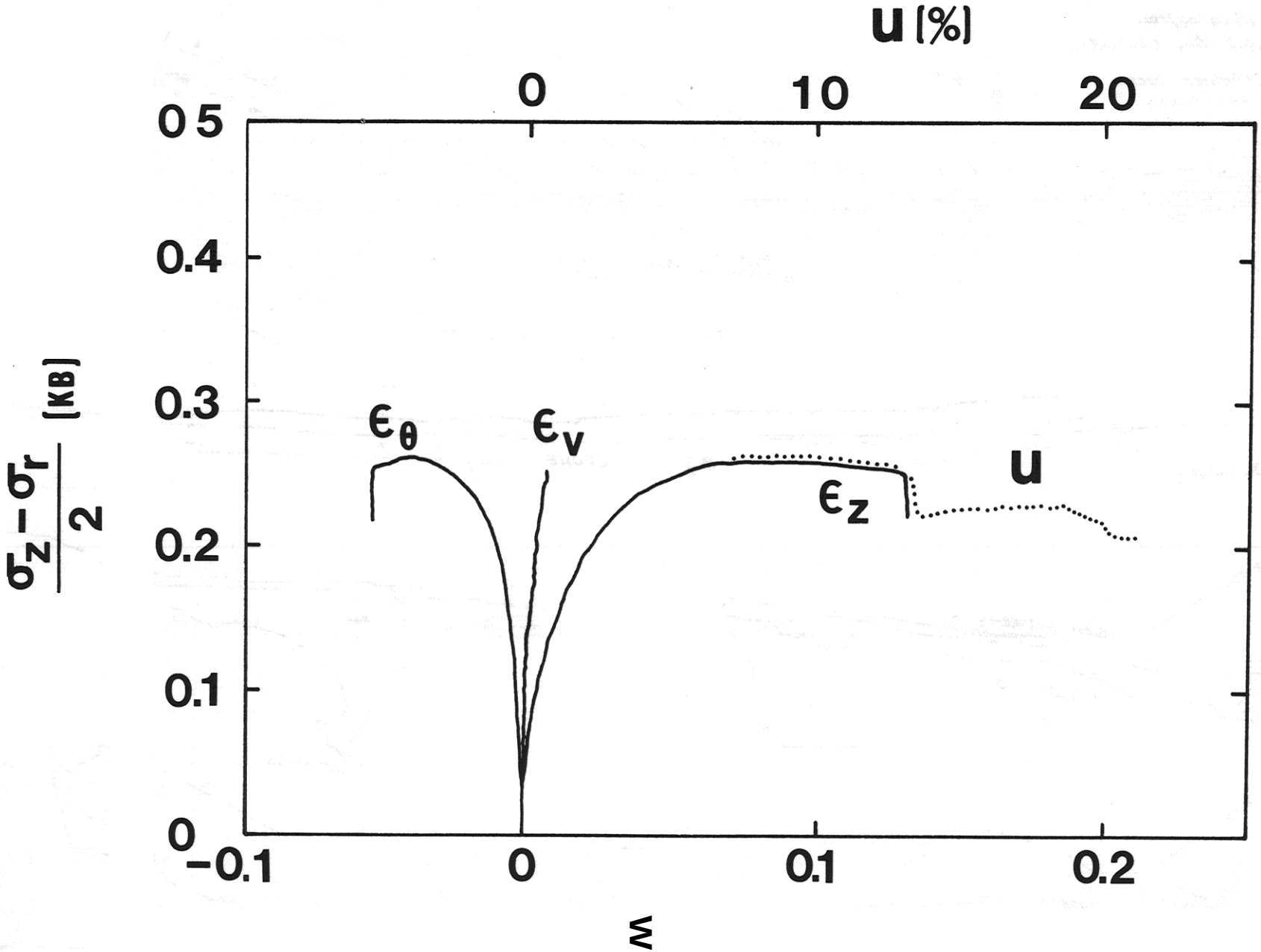
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Figure 8

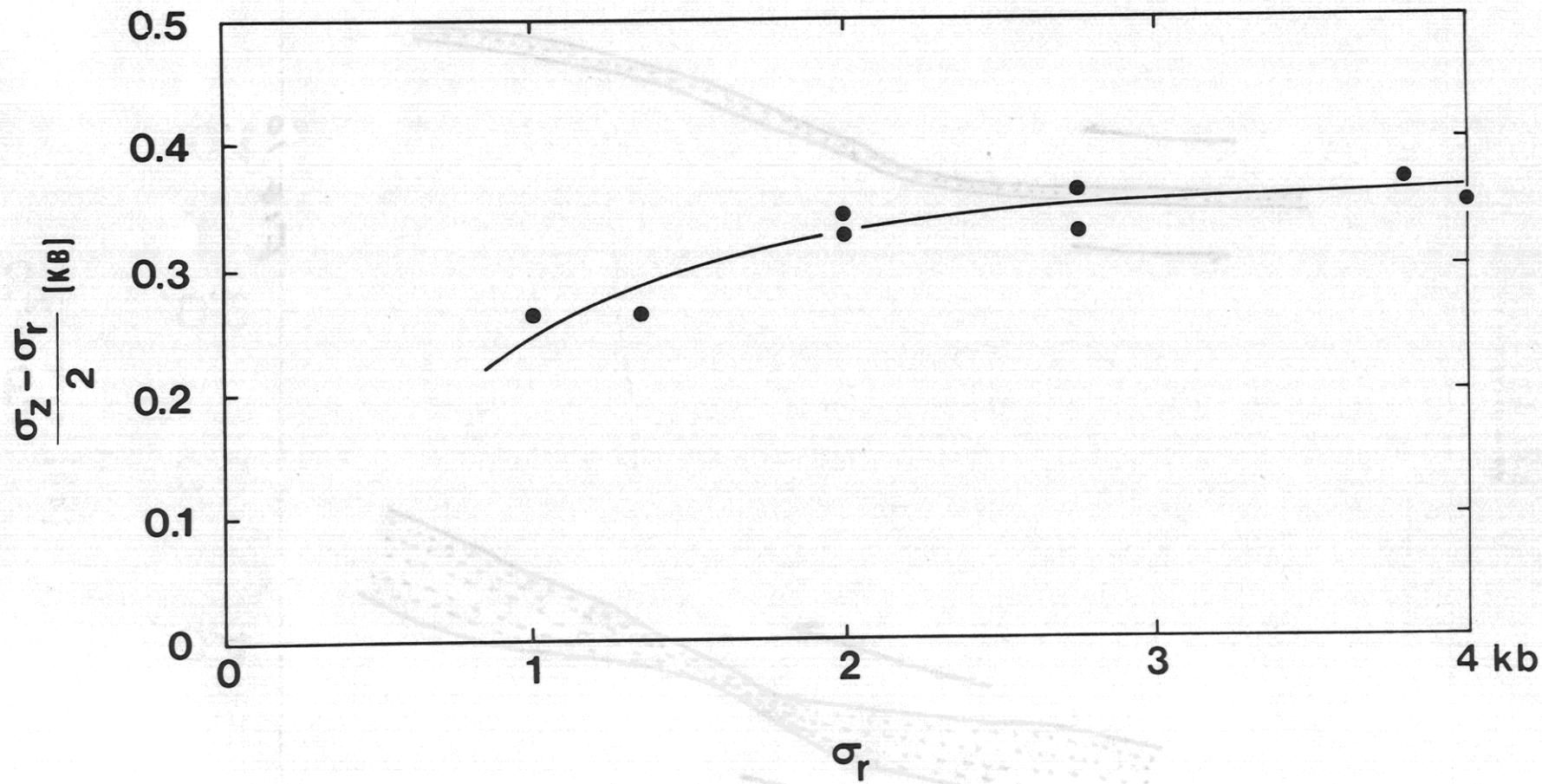

Fig. 9

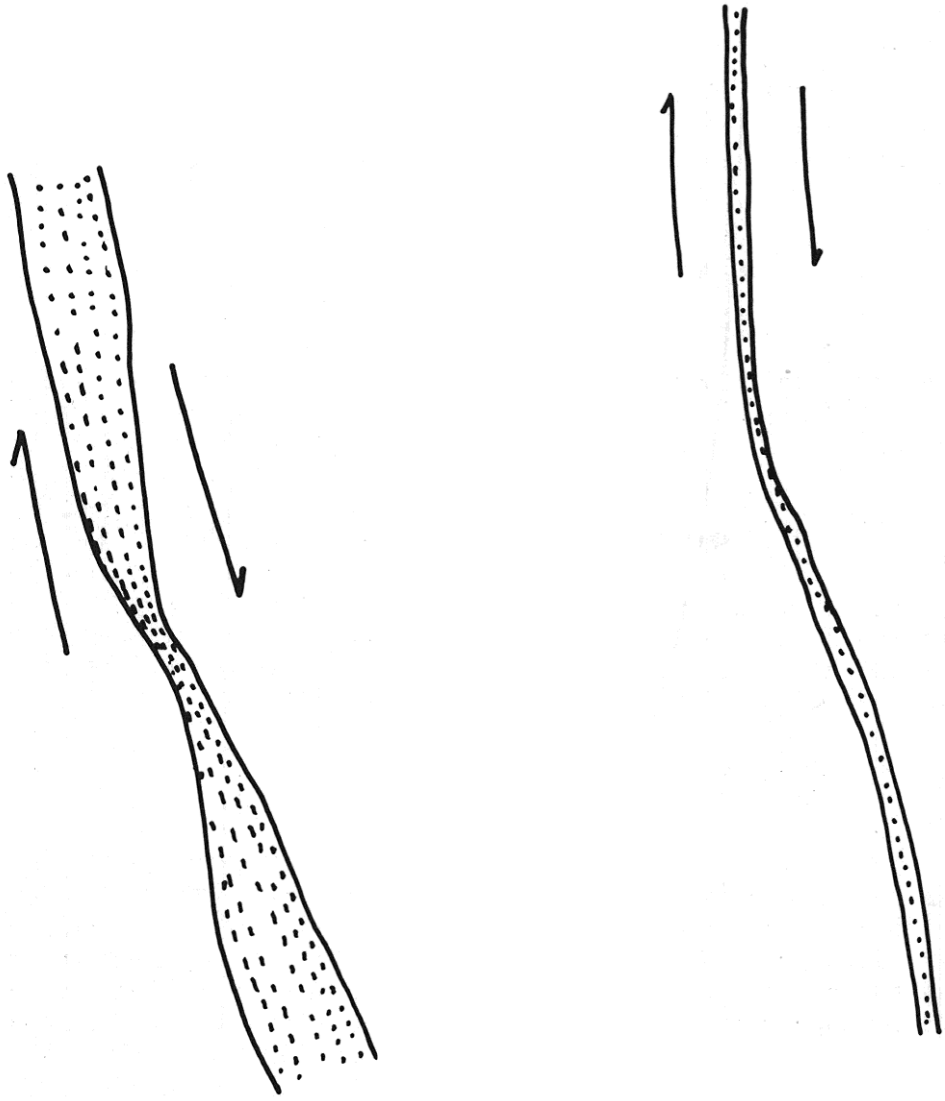

FIG 10