EVIDENCE FOR A VARIATION OF FRICTION ALONG NATURAL FAULT ZONES

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ABSTRACT

Within natural fault zones at least two mechanisms cause a variation of friction. The first is a variation of friction along smooth fault surfaces caused by either asperities with different normal stresses or changes in lithology. The second is the interlocking of asperities on irregular fault surfaces. These mechanisms are independent of the orientation of the fault zone. In the one case the rupture associated with repeated slip does not actually break through country rock, whereas country rock is sheared in the other. Because fault slip is dependent on friction and sensitive to variations, it is suggested that fault zones discussed in this paper contain mechanisms responsible for irregular propagation of earthquake ruptures or multiple event earthquakes.

INTRODUCTION

Seismologists have identified several large earthquakes that are multiple events characterized by the radiation of short-period ($T \leq 40$ sec) surface waves with a complicated pattern. Such a complicated wave pattern from the Guatemala earthquake of February 4, 1976 is believed to indicate at least 10 independent events along a 250 km long fault zone (Kanamori and Stewart, 1978). Each event occurred on a 10 km section of the fault with 14 to 40 km between sections. This multiple-shock sequence indicates that the earthquake rupture does not propagate smoothly but rather the rupture is momentarily slowed or stopped by zones of higher shear strength or friction. The same effect has been observed for the Anatolian fault zone, Turkey (Stewart, 1978), the 1923 Kanto earthquake (Imamura, 1937), the 1964 Alaskan earthquake (Wyss and Brune, 1967), and the 1940 Imperial Valley earthquake (Trifunac and Brune, 1970).

Multiple-events have been identified when many seconds separate events during large earthquakes. Although multiple-events have not been identified for smaller earthquakes, there is no physical reason why multiple-events should not occur on a smaller scale. In this latter case the radiation pattern is less distinct from a single event earthquake because the time between events is shorter. Previous studies do show that multiple-events are not restricted to one type of faulting or fault zone. Multiple-event earthquakes establish beyond a doubt that the rupture propagation is not smooth. One explanation for this behavior is that the shear strength or friction of a fault zone is not homogeneous.

Multiple-event earthquakes have been modeled by in-plane shear cracks with barriers or asperities (Das and Aki, 1977). The effect of breaking a barrier is seen on the amplitude spectrum as high frequency ripples propagating in all directions.

Do multiple-events represent the shearing of unfractured country rock, the reshearing of a healed fault zone, or shearing through a fault zone with a variable static coefficient of friction? In Laboratory experiments rock type, surface roughness, and gouge thickness are some parameters that affect the load necessary to induce slip. Do these parameters also affect natural faults and can they be recognized in outcrops? What can we learn about the variation of strength and friction within fault zones? Here I describe some outcrops that may be used to answer some of these questions.

FRICTION

Friction can have two meanings, depending on the scale of observation relative to the fault zone or sliding surface. The friction of a sliding surface is equivalent to the average shear stress to induce slip at a given normal pressure. The coefficient of friction is a constant in the functional relationship between normal stress and shear stress as determined from many experiments (Byerlee, 1978). Here friction is caused by the sum total of all mechanisms resisting slip. In contrast, the friction on the sliding surface can vary from point to point depending on the local mechanism resisting slip. The variation of friction from point to point on a sliding surface or within a fault zone is a likely parameter leading to multiple-event earthquakes.

Laboratory experiments show a variety of mechanisms for resistance to slip or friction (Engelder, 1976). Two common mechanisms include sliding on finely ground surfaces where surfaces only touch at the tips of asperities and a more general situation where irregular surfaces with large asperities interlock. In the first, frictional forces act only at the tip of asperities and sliding is permitted with little surface destruction. For the latter case, the asperities must be sheared before movement commences and friction is equivalent to the shear strength of asperities. Both mechanisms require that frictional forces develop locally rather than uniformly on the sliding surface. In the first case frictional resistance occurs only where contact is made, whereas in the latter case frictional resistance occurs both at points of contact (asperity on flat surface) as well as points of interlocking (asperity abutting asperity). Both simple mechanisms are modified with the generation of fault gouge that fills gaps between asperities on the sliding surface and may either increase the contact area and thus increase friction or decrease the friction by shielding large asperities and preventing interlocking. Examples of both types of frictional contacts are found on several scales.

A point to point variation of friction without asperity interlocking may conform to either of two models. In one model asperities touch the flat surface over a small fraction of the total fault surface. Here friction is high under the asperities and zero where there is no contact. The same picture is developed for surfaces where asperities are pressed against a flat surface, each with a different normal force. Here frictional resistance varies between points because friction is a function of normal stress. The second model involves a physical change in coefficient of friction from point to point. Here a change in lithology of the wall rock or change in fault gouge may be responsible for a variation in coefficient of friction along the fault. In this case the entire fault surface may be in contact with a constant normal stress across the fault zone.

VARIATION OF FRICTION ON "FLAT" SURFACES

On the microscopic scale or in-hand specimens slickenside surfaces show either a variation of friction caused by lithologic changes (asperities on a flat surface) or by asperity interlocking. Examples of the former are slickenside surfaces that are not the result of tectonic polishing but rather are a manifestation of pressure solution and redeposition (Durney and Ramsay, 1972; Rutter and Mainprice, 1978; Geiser, personal communication, 1979). In this case the fault surface appears to be striated yet there is never a clear beginning or end to the individual lineations. The lineations on the surface are not frictional wear grooves but rather fibers of quartz or calcite. The surfaces usually have a duller luster than a tectonically polished surface. In thin section cut normal to the surface, high points on asperities show a pressure solution, whereas low areas show a growth of minerals that sometimes form in long fibers forming the surface lineations (Fig. 1).

In this example slip results from diffusion controlled creep where point contacts do not slide in the usual frictional sense (Rutter and Mainprice, 1978). However, point contacts on asperities do resist slip and, thus, are the points of high friction along the fault zone.

On a much larger scale, exposures of the Muddy Mountain thrust fault show a variation of friction equivalent to asperities with contact areas of several km^2 (Fig. 2). Here the scale of variation of friction is too large to identify complete asperities within the area exposed below the thrust sheet. The presence of large asperities is based on the assumption that the outcrop within the Buffington Window is representative of a much larger area of unexposed fault surface. Outcrops below the fault contact show two distinct rocks: a well indurated eolian sandstone and a poorly cemented molasse. The molasse fills topographic depressions on an erosional surface that had developed on the eolian sandstone (Brock and Engelder, *1977).* The fault contact touches the indurated sandstone along several km of exposure but contacts the poorly cemented molasse along other lengths of exposure (Fig. 2). Where the fault contacts the eolian sandstone light cataclastic deformation is found up to 75 m below the contact and cataclasis pervades the sandstone within the first meter below the fault contact. Cataclasis obliterates the cross bedding near the fault contact. This cataclastic deformation is a manifestation of a high friction contact between the eolian sandstone and the thrust sheet. In contrast the poorly cemented molasse still displays a fluvial cross-bedding right up to the fault contact. The idea is that the fluvial cross beds in a poorly cemented molasse would be preserved only if the fault transmitted little if any frictional stresses to the rock below the contact. The areas of high friction are equivafent to asperities of several km^2 in area. Here it is not clear if the areas of high friction are indicative of areas of high normal stress or a fault zone with a higher coefficient of friction. Higher normal stress may result from the compaction and subsequent sagging of the fault contact upon initial loading of the poorly cemented molasse. The low friction-high friction faulting is believed to occur early in the development of the Muddy Mountain thrust fault. Late stage slip is restricted to a planar gouge zone 1-5 cm thick.

On a yet larger scale the effect of the salt pinchout on the Appalachian plateau may be considered in this discussion of low friction-high friction fault zones. Engelder and Geiser *(1979)* have mapped the orientation of strain and cleavage throughout the Devonian section of western New York (Fig. 3). This section was shortened during slip on a décollement within the Silurian salt beds of the Appalachian plateau. Strain extends beyond the salt pinchout, indicating that décollement slip also extends beyond the salt basin.

Two observations are pertinent to the low friction-high friction theme. First, over the salt, strain is at least 10% shortening parallel to the direction of tectonic transport and normal to the trend of the cleavage. At the salt pinchout strain drops abruptly to less than 2% shortening parallel to the direction of tectonic transport. The implication here is that at the salt pinchout décollement slip occurs along a zone of higher friction that inhibits slip and thus shortening parallel to tectonic transport. The second observation is that along the salt pinchout on the eastern side of the salt basin, the maximum shortening as indicated by cleavage rotates in such a manner to indicate a drag at the eastern edge of the salt basin. The drag is the result of a high friction contact east of the salt pinchout. Here the scale of high and low friction areas is too large to outline individual asperities, but rather the edges of large asperities are outlined.

In the last two examples of changes in friction along natural fault zones, the asperities were not actually bumps as the word asperity might imply but rather changes in lithology. Lithologic changes along a fault contact apparently control the frictional resistance to slip or reactivation of a fault zone. Here it seems intuitively obvious that changes in contact friction along a fault zone must influence the propagation of a rupture and stress drop during a major earthquake. Along a 250 km length of a fault zone such as the Guatemala fault zone, I might expect several lithologic changes and thus changes in the frictional properties of the fault zone.

VARIATION OF FRICTION BY ASPERITY INTERLOCKING

Like the previous examples asperity interlocking within fault zones may be seen on all scales from microscopic on up. Asperity interlocking is a mechanism for variable friction within a fault zone where the lithologies in contact do not change along the fault zone. This mechanism for variation of friction is also independent of fault type.

Some slickenside surfaces show wear grooves (Tjia, 1964, 1967; Engelder, 1974a). The wear grooves are caused by asperities ploughing into previously flat surfaces during slip (Fig. *4)*. That the grooves are spaced is evidence that friction varies with the wear grooves marking the site of high friction contacts. In general the wear grooves are seen on tectonically polished surfaces. The polished surfaces evolve from a shear fracture with gouge that is recemented, cut, and polished. The polished surface with wear grooves represents a late stage in the development of this type of fault zone. Slickensides observed within the Bonita fault zone, New Mexico and the Eureka Standard fault zone, Utah have wear grooves indicating asperities with a contact area on the order of 0.01 mm^2. These are comparable in size and shape to asperities seen in laboratory experiments. In these examples high frictional forces develop at the contact between the asperity tip and a flat surface, although ploughing is equivalent to surface interlocking. Field evidence and lab experiments indicate that some asperities are sheared after surface interlocking.

There are many examples of fault zones where shear displacement is accomplished not on one surface but rather on a complex of several subparallel shear fractures. On the outcrop scale the number of shear fractures increases with displacement along the fault zone until zones of deformation are nearly a meter thick (Engelder, 1974a; Aydin and Johnson, 1978). Figure 5 shows the development of a zone of deformation for normal faulting in a sandstone on the Colorado plateau. These zones of deformation have the same morphology as shear zones formed by fracturing of cylinders in the laboratory (Fig. *6)*. There are several ways to interpret these structures: 1) each subfracture (deformation band of Aydin and Johnson, 1978) strain hardens as it deforms, making it stronger than the parent rock. This shifts the deformation to the weaker host rock. Here the fault zone may be generated during one or several seismic events. 2) The deformation may be activated, forming a subfracture that then becomes indurated before the deformation band is reactivated. Here again the indurated subfracture becomes strong and shifts the deformation. Because induration takes time, this interpretation requires several seismic events. 3) If deformation is continuously creating irregular shear fractures, the irregularities may lock and shear off. In these three interpretations of the deformation bands documented by Engelder (1974a) and Aydin and Johnson (1978) friction changes in either time or space to effect a locking of the fault surface. Reactivation must be accomplished by shearing the undeformed rock, rather than reshearing the fault gouge of an existing fault surface.

In outcrops the interlocking and subsequent shearing of asperities is commonly seen along overthrust fault contacts. The Muddy Mountain thrust contact serves as a good example where blocks (asperities) of several m^3 are seen being sheared from the overthrust sheet during the last slip on the fault when the slip is restricted to a 1-5 cm thick gouge zone. Likewise, other blocks are incorporated in the fault zone below the fault contact and represent asperities sheared from the thrust sheet sometime prior to the last slip but during late stages of faulting. Here frictional resistance during the shearing of an asperity is greater than resistance to slip on the fault surface where it is not impeded by an asperity.

Another type of interlocking occurs where thrust faults cross stratigraphic sections on a ramp (Fig. 7). Serra (1977) documents a sequence of shear fractures where the very sharp changes in orientation of the fault's surface are gradually smoothed by the progressive shearing of the ramp and rocks in the overthrust sheet. An earthquake-like rupture propagating within this type of décollement thrust surface would certainly be slowed by the high friction at the ramp. Although Serra's description is limited to outcrop size ramps, the same change in frictional forces must be seen on a larger scale where ramp faults cut across hundreds of meters of section. Here the initial shear fracture forms a ramp at a higher angle to the slip vector than subsequent shears.

In many respects the geometry of the ramp fault is much like that seen for the great bends in both the Alpine and San Andreas faults (Scholz, 1977). The fault zone becomes oblique to the slip direction and thereby increases the normal stress across the section oblique to slip. The effect on all scales is the tendency for shearing off that part of the fault to smooth the band. Here again a fault zone oblique in the direction of slip is more likely to become locked.

Surface fault traces suggest the same low friction-high friction faulting. Rogers (1973) shows that the San Andreas and Calaveras fault zones from Hollister to San Jose are a complex pattern of subparallel to braided curvilinear fault traces, transected by a single, generally continuous, rectilinear fault trace (Fig. 8). His model for the evolution of the fault system includes locking at bends or asperities and subsequent shearing through the locked sections to restore a rectilinear path of least resistance. Rogers (1973) suggests that the system of locking at bends is more common where there is a contrast in bedrock strength on either side of the fault.

DISCUSSION

It is still not clear what parameters control the ultimate thickness of a fault zone. The Muddy Mountain thrust is an example of a fault that developed a 200 m wide fault zone early in its history only to have most of its late stage slip restricted to a 1-5 cm thick gouge zone with occasional asperities sheared off the overthrust sheet. The zones of deformation documented by Aydin and Johnson (1978) and Engelder (1974a) appear to thicken up to several meters before a through going fault surface develops along which most of the slip occurs. In spots the San Andreas fault zone is more than a km thick. With the complicated development of fault zones, it is not clear if any one stage of development is more likely to be seismic or the site of a multiple event earthquake.

CONCLUSIONS

In nature there are at least two mechanism that may cause an earthquake rupture to propagate in an irregular manner. These mechanisms are independent of the orientation of the fault zone. The first is a variation of friction along the fault contact caused by either asperities (change in normal stress) and changes in lithology. The second is the interlocking of asperities on irregular fault surfaces. In the former case the rupture does not actually break new material whereas the rupture fractures country rock in the latter case. To date it is hard to distinquish if either of these mechanisms is responsible for multiple event earthquakes.

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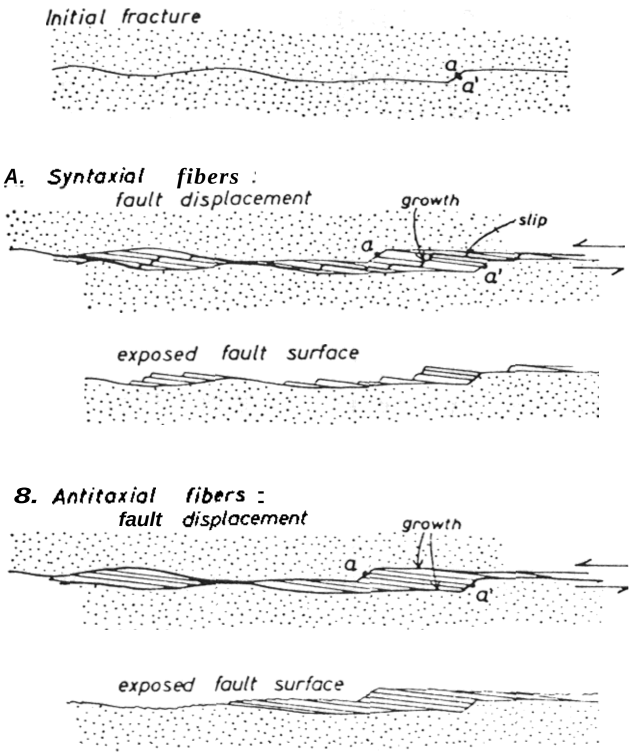


FIG. 1 Formation of crystal fibers along a fault surface by two different mechanisms. The fibers form at a low angle to the fault, and their total length is a measure of the total separation across the fault of two originally adjacent points *a* and *a'*. When the fault plane is exposed the fibers formed by both mechanisms have a characteristic step-like structure, a feature which gives rise to the "rough-smooth" effect when the surface is touched with the hand.

DURNEY & RAMSAY 1973

Figure 1

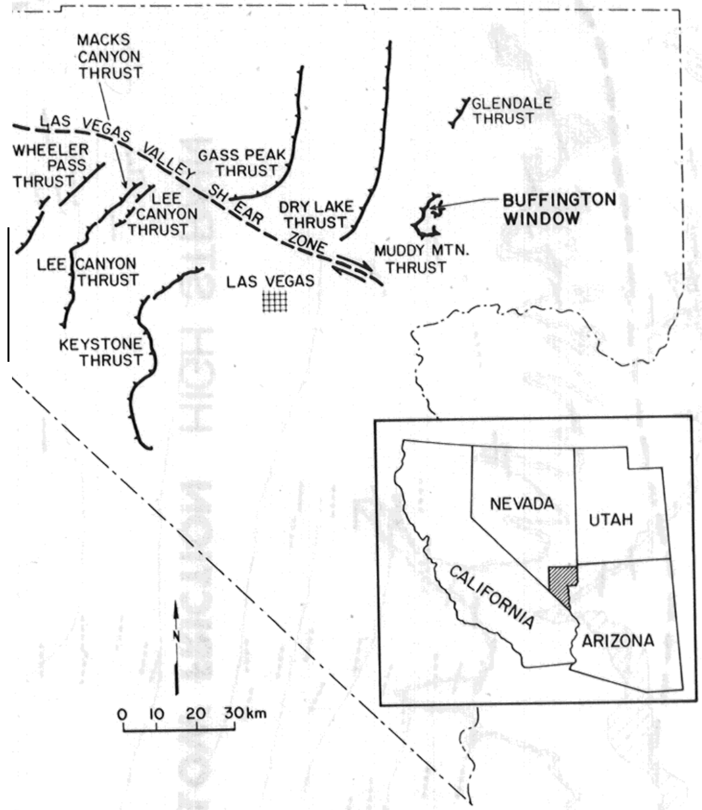


Figure 2. Major overthrust faults in Clark County, Nevada.

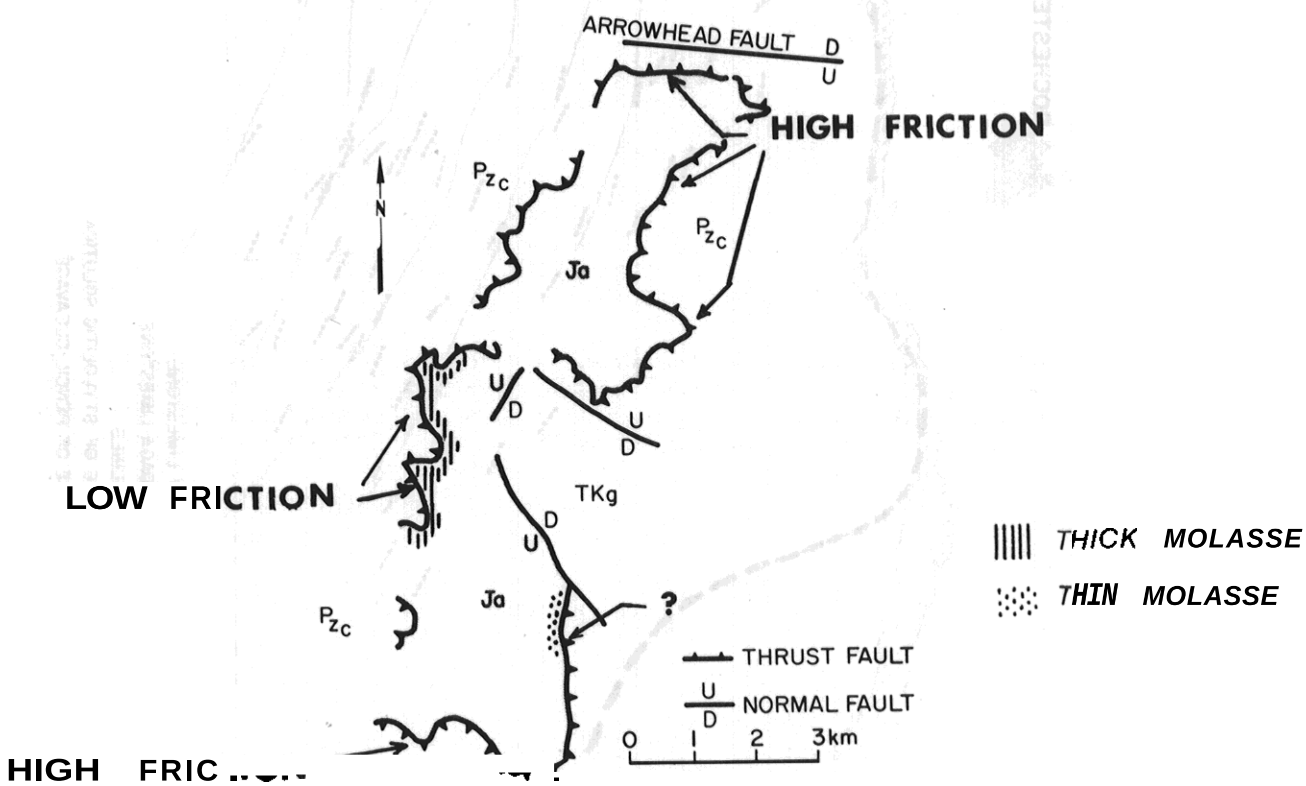
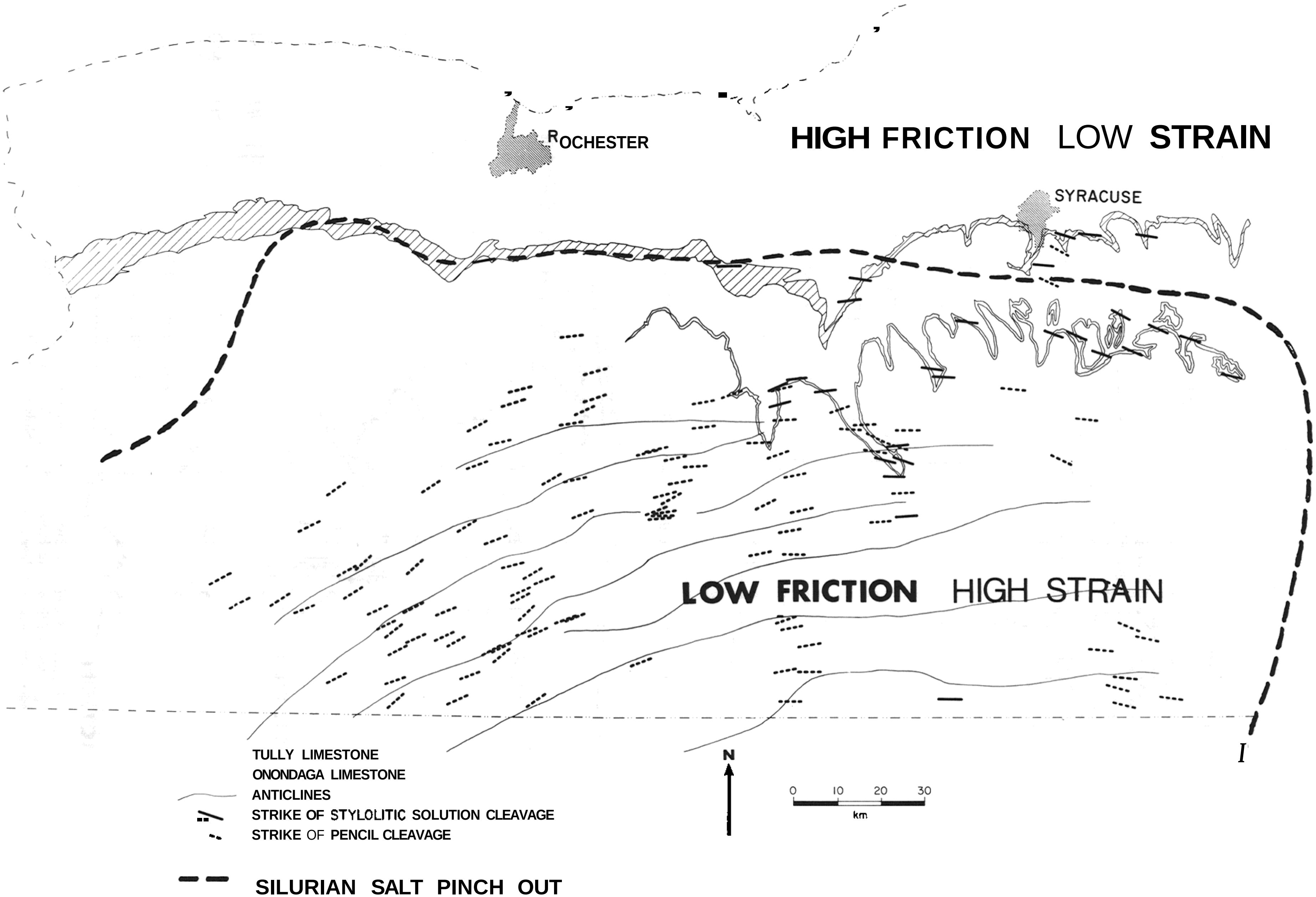


Figure 3. Geology of Buffington window, Muddy Mountains, Nevada. P_{z_c} = lower Paleozoic carbonate racks; J_a = Aztec Sandstone covered with minor amounts of molasse; TK_g = Gale Hills Formation.

Figure 2

BROCK AND ENGELDER, 1977



ENGELDER & GEISER, 1979

Figure 3

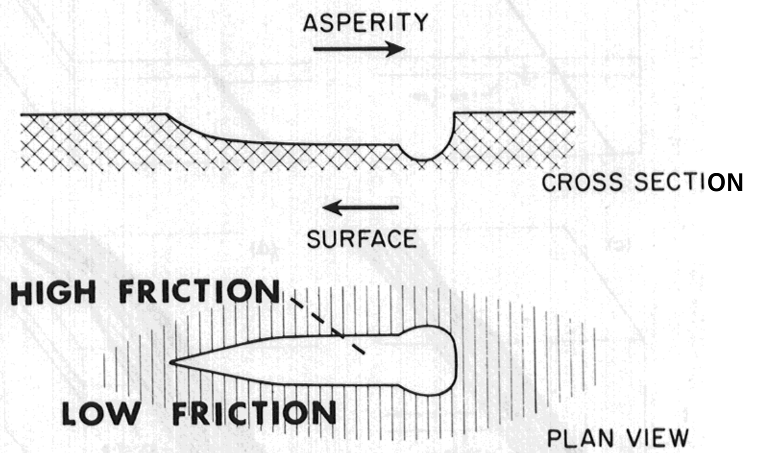


Figure 1. Schematic of an ideal wear groove in feldspar. Arrows indicate sense of slip.

ENGELDER, 1976

Figure 4

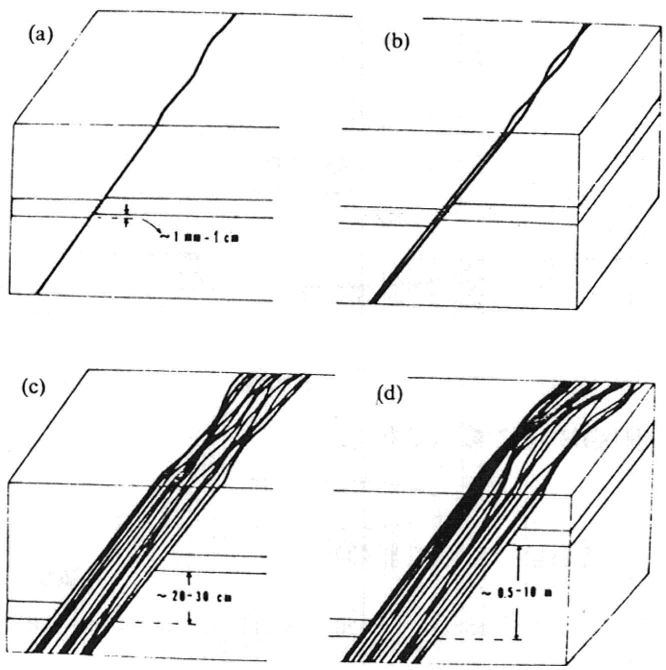


Figure 7
Series of block diagrams. showing sequential development from a single band to a slip surface.
(a) Single deformation band.
(b) Two inosculating bands.
(c) A zone of deformation bands.
(d) Slip surface has developed on left-hand edge of zone

AYDIN & JOHNSON, 1978

Figure 5

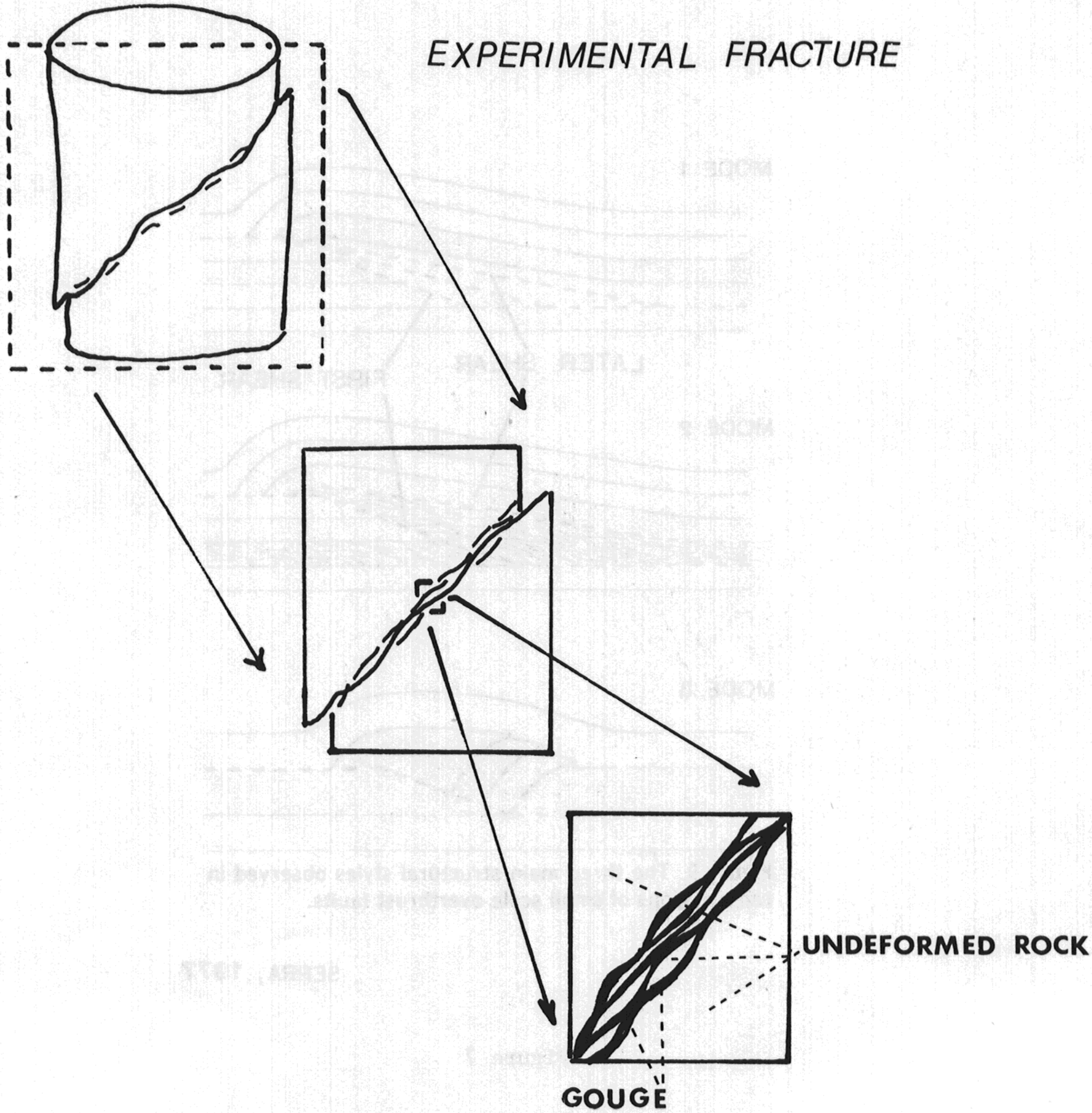


Figure 6

ENGELDER and others, 1975

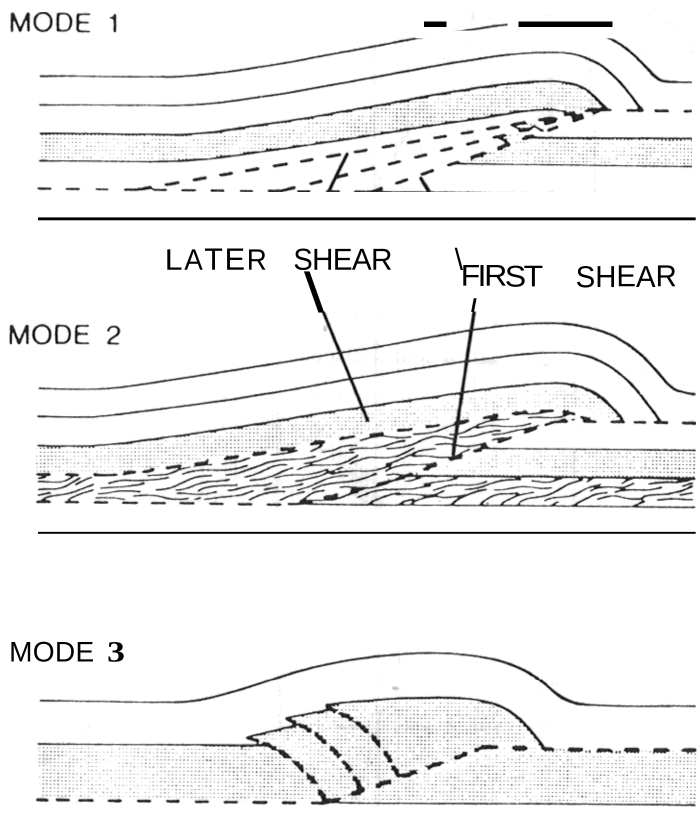


Figure 9. The three main structural styles observed in ramp regions of small scale overthrust faults.

SERRA, 1977

Figure 7

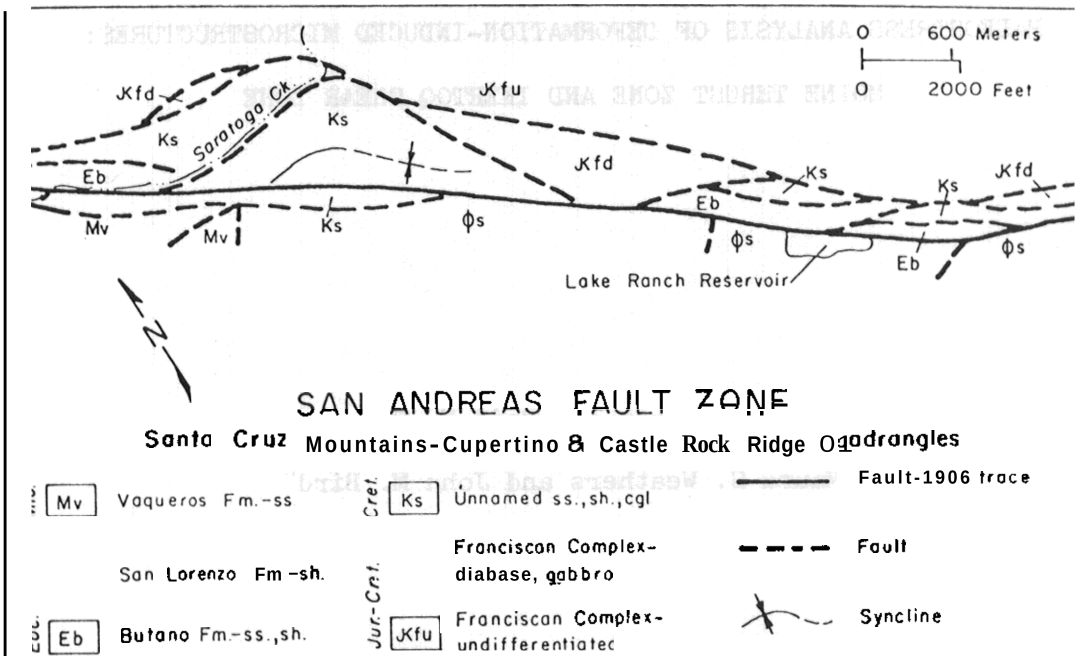


FIGURE 2

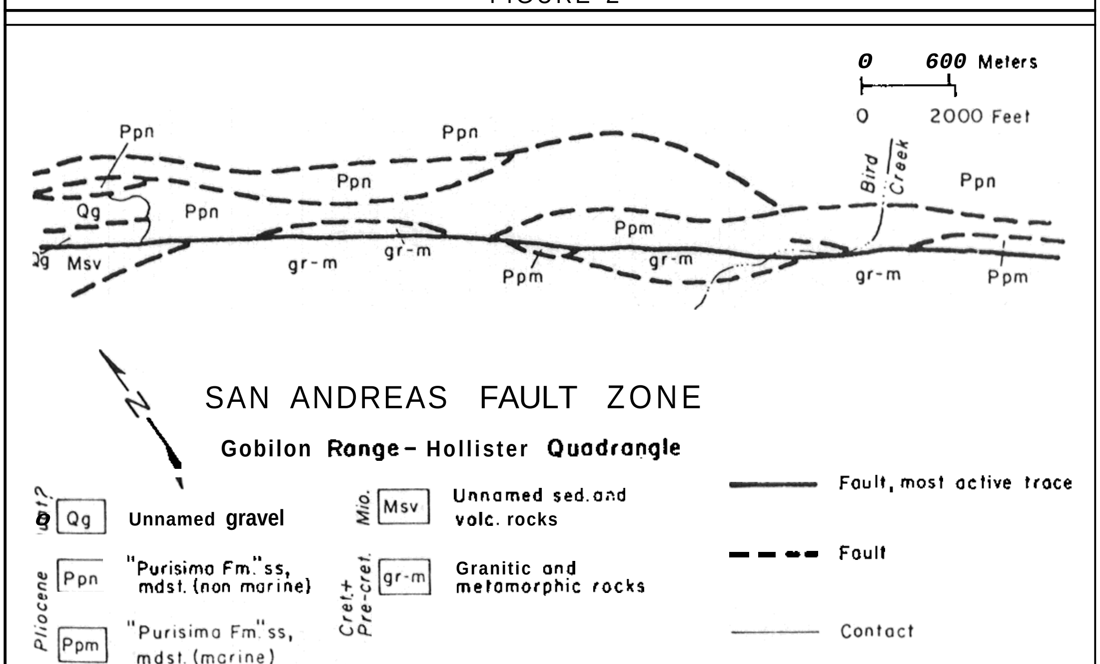


FIGURE 3

ROGERS, 1973

Figure 8