EVIDENCE FROM STRAIN RELAXATION TESTS

FOR THE EXCHANGE OF PRINCIPAL STRESS AXES

by

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ABSTRACT

Regional orthogonal joint sets suggest that stress axes must exchange during the history of the rock in which the joints form. Strain relaxation measurements from the Appalachian Plateau reflect the orientations of local structures and support the idea that stress axes exchange.

INTRODUCTION

In making interpretations of regional stress fields using _in situ_ stress data, earth scientists should remember that principal compressive stress axes may exchange during the mountain building process. Residual stresses reflecting this change in orientation of stress axes may influence _in situ_ stress measurements to give a pattern of data that suggests a complicated spatial distribution of stresses. The rotation of principal stress axes with depth is well established by hydraulic fracture tests (Haimson, 1978; Zoback, 1978) and complications in the horizontal stress field are emerging with increases in _in situ_ stress data (Sbar, Engelder, and Tullis, 1978).

The purpose of this paper is to examine some evidence suggesting that the exchange of principal stress axes is common and that the stress orientations following this exchange survive to affect _in situ_ stress measurements.

OUTCROP EVIDENCE FOR EXCHANGE OF PRINCIPAL STRESS AXES

The main outcrop evidence attesting to the exchange of principal stress axes is the presence of orthogonal joints. Conventional wisdom suggests that, for an isotropic rock, joints propagate in a deviatoric stress field such that the normal to the joint was parallel to the least

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principal compressive stress. For the isotropic case, an orthogonal joint can propagate only after the principal stress axes have exchanged position. However, other possibilities exist for the formation of joints intersecting at high angles. Some joints are inherited from underlying units and, thus, their orientation may be controlled by the local stress field associated with the underlying joints. Still other joints may be controlled by a tectonically imposed fabric and thus propagate in response to a stress imprinted on the rock but no longer the ambient stress field at the time of propagation.

Joints intersecting at angles greater than 70° are a paradox in light of currrent understanding of fracture mechanics. Shear fractures developed on the flanks of folds and within fault zones intersect at 60° or less (Stearns, 1968). Although extension fractures (joints) developed on or near the same structures intersect to form orthogonal sets, complicated histories involving the exchange of principal stress axes must be invoked. In fact, the exchange has been demonstrated for positions above the neutral fiber in experimentally produced folds (Handin et al., 1972). Excluding joints developed near faults or high amplitude folds, there is still a class of regionally pervasive joints that forms sets, intersecting at high angles. Examples are found in the cratonic edges of foreland fold and thrust belts such as the Appalachian Plateau (Parker, 1942; Nickelsen and Hough, 1967) and the eastern edge of the Canadian Rockies (Babcock, 1974).

Explanations of regional joint sets intersecting at high angles include both an origin by shear fracturing and an origin by extension fracturing. Stress systems generating extension fracturing include those associated with crustal flexure on erosion and uplift (Price, 1966, 1974) and thermal cooling (Voight and St. Pierre, 1974; Haxby and Turcotte, 1975). To generate failure by extension, models for uplift appeal to a Poisson contraction from removal of overburden or stretching about a surface of increasing arc lengths. Thermal models use contraction upon cooling. Stress systems for developing shear fractures are taken from Anderson (1952) where the bisector of the acute angle between the fractures is the orientation of σ_1 (Scheidegger, 1977).

The difference between Scheidegger's (1977) model for regional fracture sets intersecting at high angle and those models of Price (1974) and others is that stress axes must be exchanged during the history of rocks for the latter models. Shear fractures may form without appealing to a complicated stress history. In my experience on the Appalachian Plateau the regional joint sets have the characteristics of extension fractures. For this reason I will focus on the evidence for the exchange of stress axes during the tectonic history of a weakly deformed upper crust.

Evidence for the exchange of stress axes comes from major folds, as indicated by petrofabric data. Petrofabric studies measure the permanent strain of minerals such as calcite and assume coaxial principal axes for stress and strain to derive a stress history. Using petrofabric data, Burger and Hamill (1976) show that the Dry Creek Ridge Anticline was initially subject to layer-parallel shortening (compression) throughout the fold, followed by an exchange of principal axes above the neutral fiber upon bending. The critical piece of information missing from Burger and Hamill's study is the time at which the exchange occurred during the history of the fold.

The reason that timing of the exchange is critical is that orthogonal joint sets on the Appalachian Plateau indicate an exchange before the ratio of amplitude to wavelength increases to 10^{-2}. One interpretation is that rock in foreland fold and thrust belts is very sensitive to bending moments and bending moments in turn influence the in situ state of stress enough to cause exchange of principal stress axes.

Further evidence for the exchange of stress axes is found in the study of conjugate folds forming in response to a σ_1 parallel to the structural grain of mountain belts - for example, the Caledonides of Ireland (Anderson, 1968), the Sierra Nevada (Tobisch and Fiske, 1976) and the Precambrian rocks of India (Naha and Halyburton, 1974), Canada (Clifford, 1968), and South Africa (Ramsay, 1963). In these cases the shift of σ_1 came toward the end of multiphase deformation.

PREVIOUS TESTS INVOLVING RESIDUAL STRESS

Rocks subject to permanent strain also contain a component of recoverable strain that has principal axes correlating with the permanent strain (Engelder, 1979). The recoverable strain, sometimes called an elastic residual strain, is poorly understood and often confuses the interpretation of in situ stress measurements. The confusion arises when residual strain orientations are used to represent the orientation of modern tectonic stresses related to plate tectonic processes.

Data sets from folds in four different tectonic settings suggest: A) residual strains record the exchange in stress axes; B) the exchange of stress axes is complicated and does not affect residual strain data from all folds; and C) in situ stress data with other interpretations may also be interpreted in light of the data on exchange of stress axes.

Case 1. Foreland fold and thrust belts form in response to a compression normal to fold axes or outcrop traces of imbricated thrust faults. In the foothills of the Canadian Rockies that compression is oriented about N55°E (Fig. 1). Samples of Cretaceous sandstone from seven localities in the foothills were prepared for point-loading tests and uniaxial compression experiments (Reik and Currie, 1974). All samples were prepared so that loads could be applied normal to bedding. In general tensile fractures induced during these tests propagated parallel to the local fold axes and, thus, normal to the direction of compression causing the foreland fold and thrust belt. Elastic residual strain was measured with an orientation of maximum compressive strain parallel to the fold axes and the direction of fracture propagation. The fabric of the rock which is a residual strain controlled the orientation of fracture propagation. At some time during the history of the Canadian Rockies, the foothills stress axes were exchanged to impart a residual elastic strain parallel to the fold axes. The presence of a residual stress parallel to fold axes suggests that vertical joints parallel to fold axes formed later than vertical joints normal to fold axes.

Case 2. The formation of a fracture cleavage on unloading is seen in some minor folds in Carboniferous sandstones of Nova Scotia (Fig. 2). Here again point loading tests show a consistent propagation of the fractures parallel to the fold axes (Stringer, 1978). A pre-existing fabric in the rocks gives an anisotropy which controls the propagation of the fractures. Here the present fabric has the effect of controlling the propagation of fractures parallel to the fold axes and normal to the compression assumed to cause the fold.

Case 3. Residual strain has been measured on the flanks and nose of Rattlesnake Anticline, Wyoming, where three samples show maximum compression normal to the fold axis and the fourth sample shows no trend in residual strain (Fig. 3). Friedman (1972) shows that on the steep flank of Rattlesnake Mountain Anticline, the maximum compression (σ_1) is down-dip, whereas the least compressive stress (σ_2) is parallel to the strike of the beds and fold axes. In the nose of the anticline, the σ_1 is parallel to the strike of the bedding and σ_2 is parallel to the fold axis and dip direction. Here σ_1 does not change in orientation between the flank and nose of the fold and the orientation of the residual strain is opposite from that seen in Cases 1 and 2.

Case 4. Data from the Rangely Anticline, Colorado, show a combination of Cases 1, 2, and 3. Several near-surface *in situ* stress techniques were compared in an effort to measure a stress field associated with injection-induced earthquakes (de la Cruz and Raleigh, 1972). Making a very rough interpretation of the data, σ_1 is within 20° to the fold axes on the flanks of the fold, whereas σ_1 is 20° from the fold axes on the nose (Fig. 4). Here residual strain was measured only at the site on the nose of the fold and its orientation was the same as the stresses indicated by the several techniques for measuring *in situ* stress. If the data on the flanks of the fold are a measure of residual strain, the orientation of residual strain is the same as documented in Cases 1 and 2. The orientation of the residual strain on the nose of the fold is much like that in Case 3.

STRAIN RELAXATION DATA FROM CENTRAL PENNSYLVANIA

Strain relaxation data were gathered from rocks in the vicinity of Williamsport, Pennsylvania, where folds of the Appalachian Valley and Ridge Province to the south decrease abruptly in amplitude at the boundary with the Appalachian Plateau to the north. The Appalachian Plateau is characterized by folds with amplitudes of less than 100 m and wavelengths of 10-20 km. In general, these folds are cored with a series of small displacement (< 100 m) imbricate thrusts splaying off a master décollement (Gwinn, 1964). In contrast, the Valley and Ridge is a series of imbricated and folded thrust sheets where major parts of the Paleozoic section are repeated (Gwinn, 1970).

Strain relaxation was accomplished by overcoring strain-gauge rosettes bonded to bedding surfaces. To permit thermal compensation, one rosette per outcrop was not overcored, and thermal strains recorded with this rosette were subtracted from the total strains recorded with overcored rosettes. Further details of the experimental technique may be found in Engelder and Sbar (1976) and Engelder et al. (1977).

To gather data from both carbonates and sandstones, measurements were made on both the Valley and Ridge and Appalachian Plateau (Fig. 5). The exposures of bedding plane surfaces in sandstones are poor in the Valley and Ridge, whereas carbonates are not exposed on the Plateau. Rock types, formation names, and ages for each outcrop are listed in Table 1. To permit the measurement of strain relaxation parallel to bedding, outcrops with dips of less than 5° were selected. During the experiment described here, outcrops with steeply-dipping bedding were also sampled. However, the surface to which the rosette was bonded cut the bedding surface. Because the bedding plane anisotropy influences the measurement and because the relaxation parallel to bedding is of most interest here, the data from surfaces cutting bedding will not be considered.

Figure 5 shows the strain relaxation data for carbonates and sandstones in the vicinity of Williamsport. On initial inspection the data set appears to have no coherence; both expansions and contractions are seen for many of the outcrops and the orientations of maximum expansion are scattered. The lack of coherence, at least on the map in Figure 5, distinguishes this data set from relaxation data gathered elsewhere showing a consistent orientation and usually expansion upon relaxation (Engelder and Sbar, 1976, 1977; Engelder et al., 1977; and Engelder, 1979).

Making the assumption that there is a signal within the data and that this signal is related to the local structures, the following picture emerges. The strikes of all vertical joints seen in each outcrop are plotted in Figure 6. Fiester is a 3 m by 3 m outcrop in sandstone with no joints exposed. The other two sandstone outcrops, W.P.A. and Ganoga, contained two sets of joints intersecting at high angles. One set was subparallel to the local strike of bedding, whereas the other was subparallel to the local dip. In the limestones, Royer, an outcrop in the nose of an anticline, had one joint set subparallel to the local dip, whereas Slammer had two orthogonal joint sets with one subparallel to the strike of bedding. Jointing in the dolomite outcrops was more complicated because three to four joint sets could be distinguished. The orientations of joints clustered to a greater extent in some outcrops than in others. Joint sets in the sandstones correlate with those described in Nickelsen and Hough (1967) for the Appalachian Plateau. Joint sets in the carbonates vary in orientation depending on structural position within folds. The relationship between jointing and structure in the dolomites is not clear.

Expansion upon overcoring is most common during strain relaxation measurements from other areas in the northeastern United States. For this reason, attention is focused on the data from those relaxation tests resulting in expansion (Fig. 6). The sandstones show five tests where expansion occurred and the orientation of maximum expansion for all five cluster about the strike of the fold axes on the Appalachian Plateau. Here the fold axes strike about N70°E. The limestones in the Valley and Ridge show a maximum expansion parallel to the local strike of bedding. Although the outcrop, Royer, shows a N-S maximum compression, it is on the nose of a fold where bedding strikes N-S. The orientation of maximum expansion of the dolomites shows little tendency to cluster. For comparison the strain relaxation data are shown by aligning the strike and dip of bedding in Figure 7. Here it is shown that for those experiments where maximum expansion occurred, the maximum expansion in the limestones and sandstones was parallel to the strike of bedding.

DISCUSSION

I suggest that strain relaxation data from central Pennsylvania reflect a residual strain. The relationship between local structure and strain relaxation supports this. The strain relaxation data from Pennsylvania compare with the four cases for which the orientation of residual stress is known relative to various folds. Residual stress on the noses of folds was oriented with σ_1 across the fold axes. Strain relaxation at Royer fits this behavior. On the flanks of folds the residual stress may be parallel to the fold axes and this was the case indicated by strain relaxation data for the limestone at Slammer, and the three sandstone outcrops.

Assuming that the strain relaxation data from near Buffalo, New York (Engelder, 1979), can be compared with the data from near Williamsport, Pennsylvania two things are apparent. First, the strain relaxation of sandstone, limestone and dolomite was sampled in both areas and data from both areas show the orientation of maximum expansion for the limestone and sandstone clustered, whereas there was no tendency for the orientation of maximum expansion to cluster for the dolomites. The direction of maximum compression for both the outer edge of the Appalachian Plateau near Buffalo, New York, and inner edge near Williamsport, Pennsylvania, was NNW. Maximum expansion for residual strain near Buffalo is parallel to the direction of tectonic compression whereas near Williamsport, the direction of maximum expansion is normal to the direction of maximum tectonic compression. During formation of the Appalachian Mountains the rocks near Buffalo were subject to a weak deformation whereas those near Williamsport were subject to a more intense deformation. Those sandstones and limestones subject to the more intense deformation contain a residual strain indicating the exchange of stress axes sometime during the history of the rocks. This can be stated with certainty because zones of weak deformation migrate toward the craton during the development of foreland fold and thrust belts. Such a weak zone of deformation as now seen near Buffalo once passed through the Williamsport area.

The well-developed sets of joints that form at high angles on the Appalachian Plateau are seen in central Pennsylvania and extend into central New York but are not seen in the vicinity of the strain relaxation experiments near Buffalo, New York. The development of regional orthogonal joints evidently follows the imprinting of residual strain on the rocks. However, within the area where the two joint sets have formed strain relaxation texts suggest the stress axes exchange to permit the formation of orthogonal sets.

CONCLUSIONS

(1) Comparison of strain relaxation data from near Buffalo with that from Williamsport suggests that the principal axes of residual strain exchange from parallel to the direction of tectonic compression near Buffalo to normal to that direction near Williamsport.

(2) The orientation of strain relaxation near Williamsport compares with residual strain data from other folds where the maximum expansion upon relief is parallel to the fold axes.

ACKNOWLEDGEMENTS

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TABLE 1

Rock Types of Central Pennsylvania

Site	Rock Type	Age
Ganoga	Sandstone	Pennsylvanian
Feister	Sandstone	Pennsylvanian
W.P.A.	Sandstone	Devonian
Slammer	Carbonate	Silurian
Royer	Carbonate	Silurian
Oval	Carbonate	Ordovician
Pine	Carbonate	Ordovician

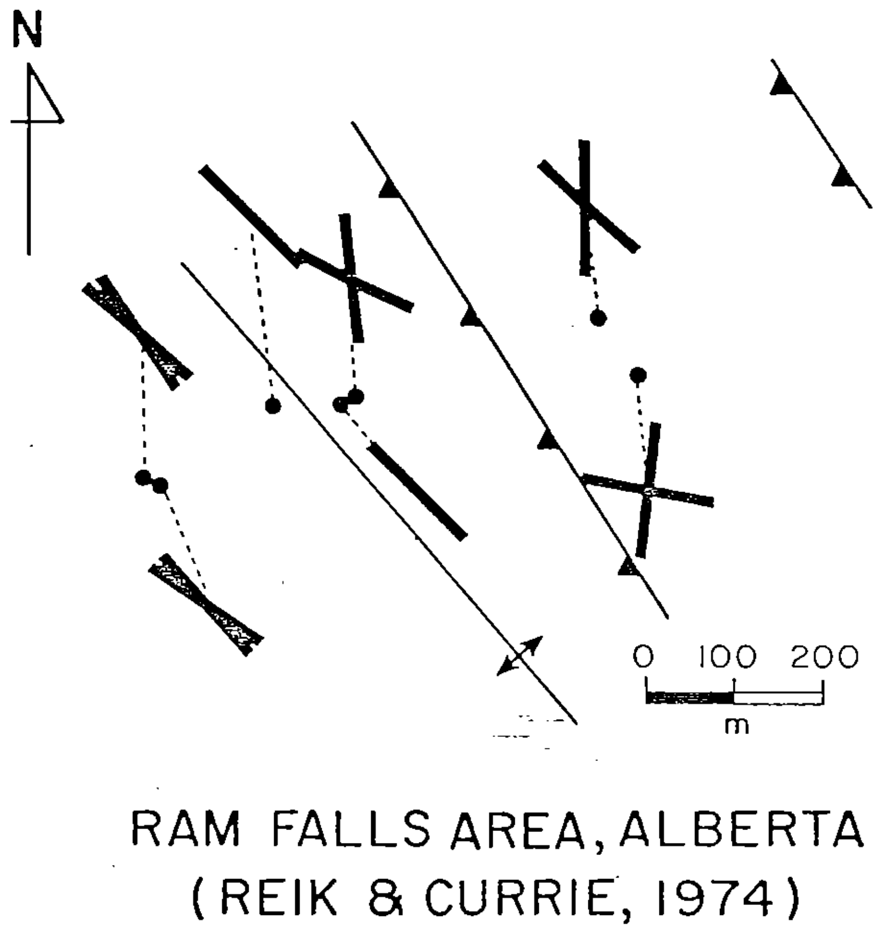


Figure 1. The orientation of fractures from point load tests taken from several localities in the foothills of the Canadian Rockies (after Reik and Currie, 1974). The traces of local fold axes and thrust faults are also shown.

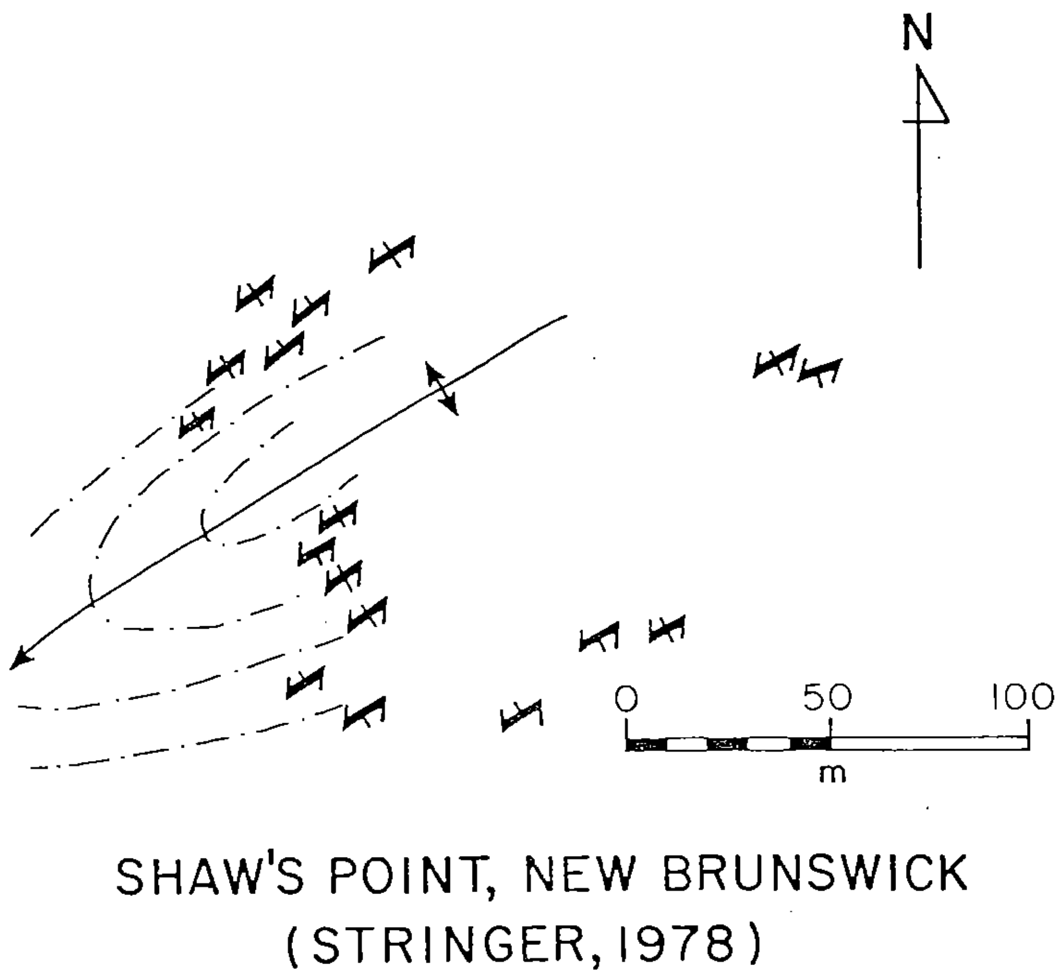
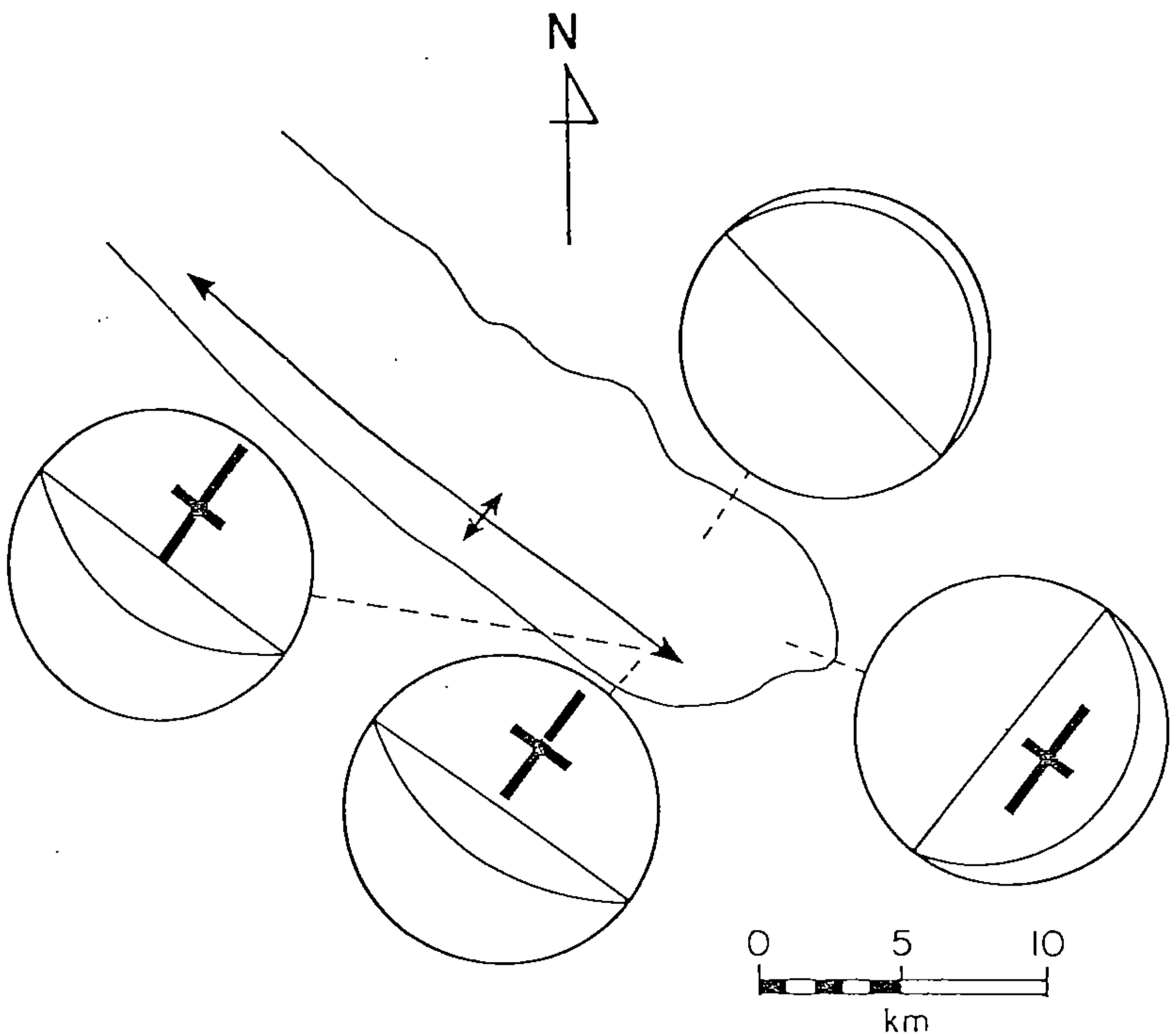
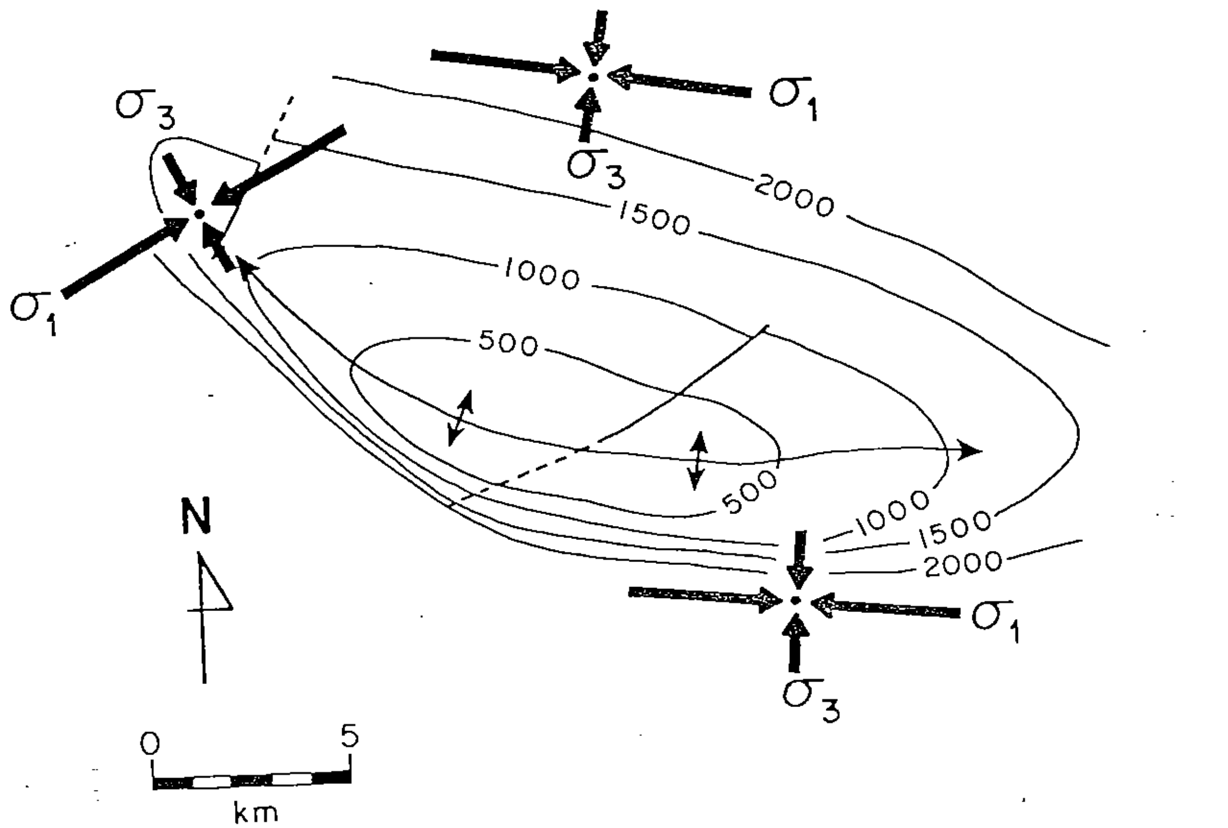


Figure 2. The orientation of fracture cleavage on an anticline at Shaw's Point, New Brunswick.



RATTLESNAKE MOUNTAIN, WYOMING
(FRIEDMAN, 1972)

Figure 3. The orientation of residual strain parallel to bedding on the flanks and nose of Rattlesnake Mountain Anticline (Friedman, 1972). The maximum compression is indicated by the long axes of the dark crosses. The lower hemisphere projections show the dip and strike of bedding at each of four sample localities.



RANGELY ANTICLINE, COLORADO
(RALEIGH, 1972; de la CRUZ & RALEIGH, 1972)

Figure 4. Orientation of in situ stress as determined by several near-surface stress techniques at the Rangely Anticline, Colorado. Structural contours are drawn from the top of the Weber sandstone.

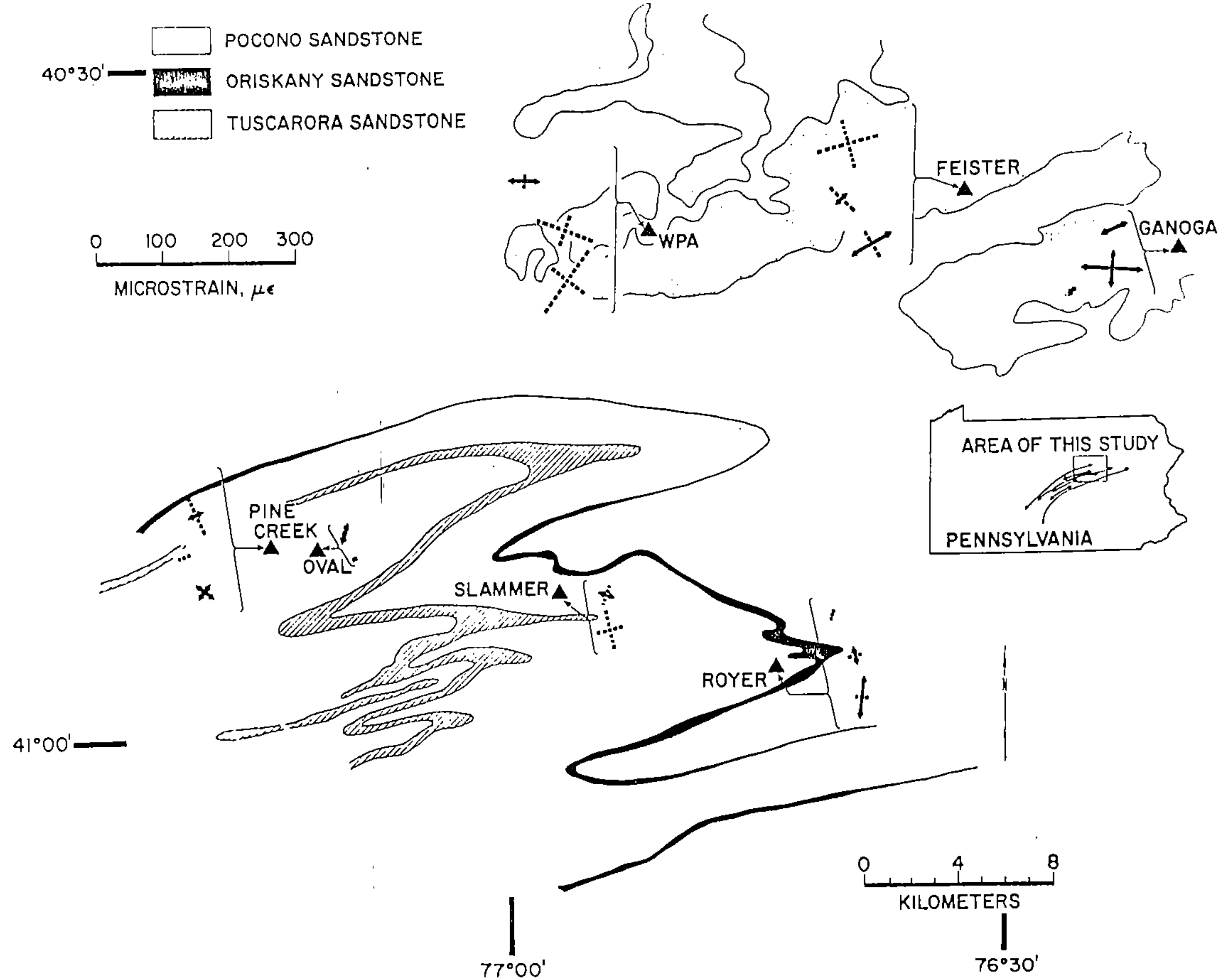


Figure 5. Geology and in situ strain in the vicinity of Williamsport, Pennsylvania. Anticlines of the Valley and Ridge are shown in the small map of Pennsylvania. Three sandstone units are shown to illustrate their areal distribution. The Pocono sandstone outcrops on the Appalachian Plateau, whereas the Oriskany and Tuscarora sandstones outcrop within the Appalachian Valley and Ridge. Seven sample sites are named. The magnitude and orientation of the strain relieved is indicated. Solid lines represent expansion and dashed lines represent contractions (negative expansions). A scale for the magnitude of the relieved strain is given in microstrain (με).

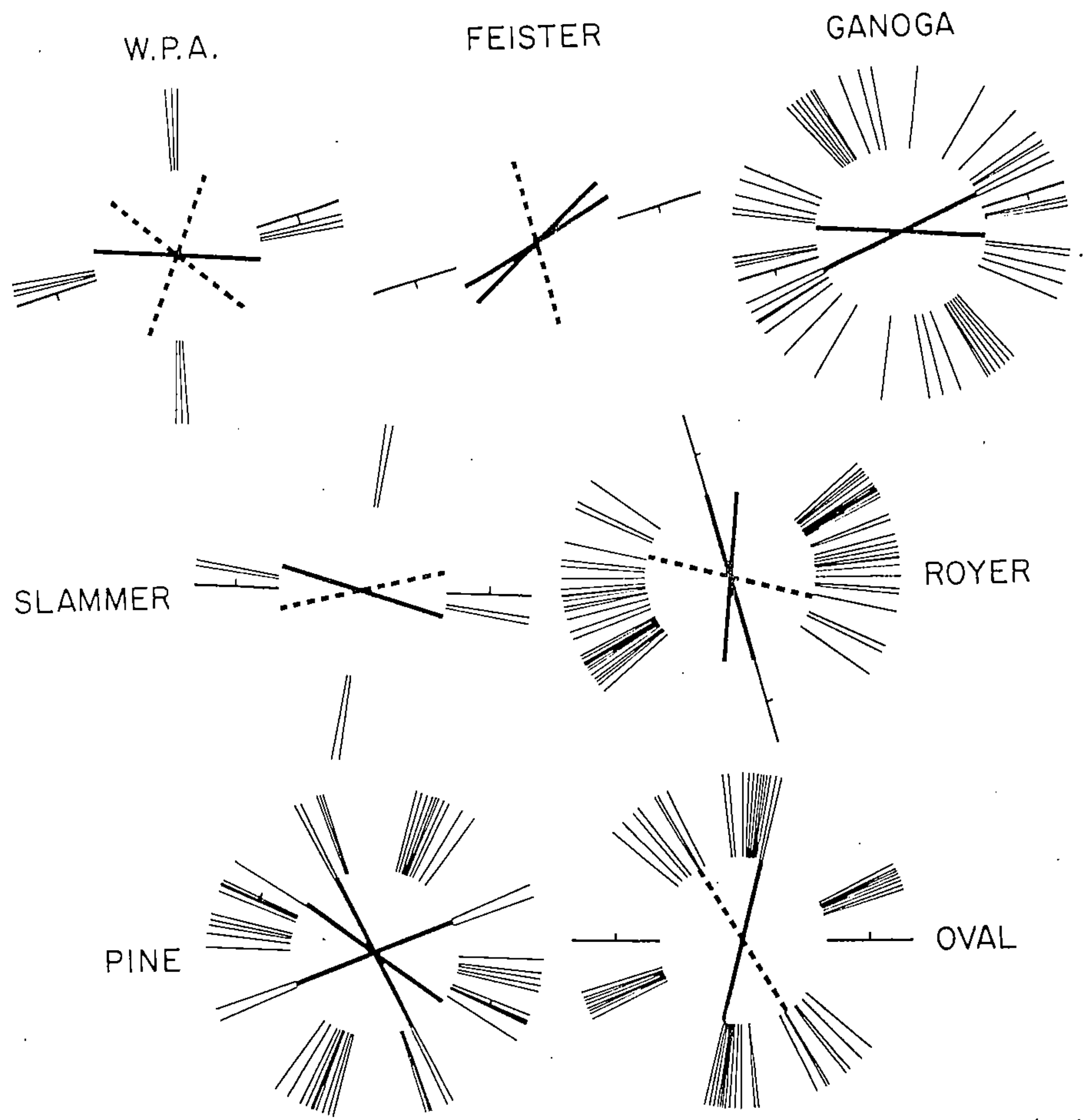


Figure 6. The orientation of maximum expansion, joints and attitude of bedding in seven outcrops in the vicinity of Williamsport, Pennsylvania. The top of the Figure is facing north. The maximum expansion is indicated with heavy lines, where the solid lines are positive and the dashed lines are negative. The

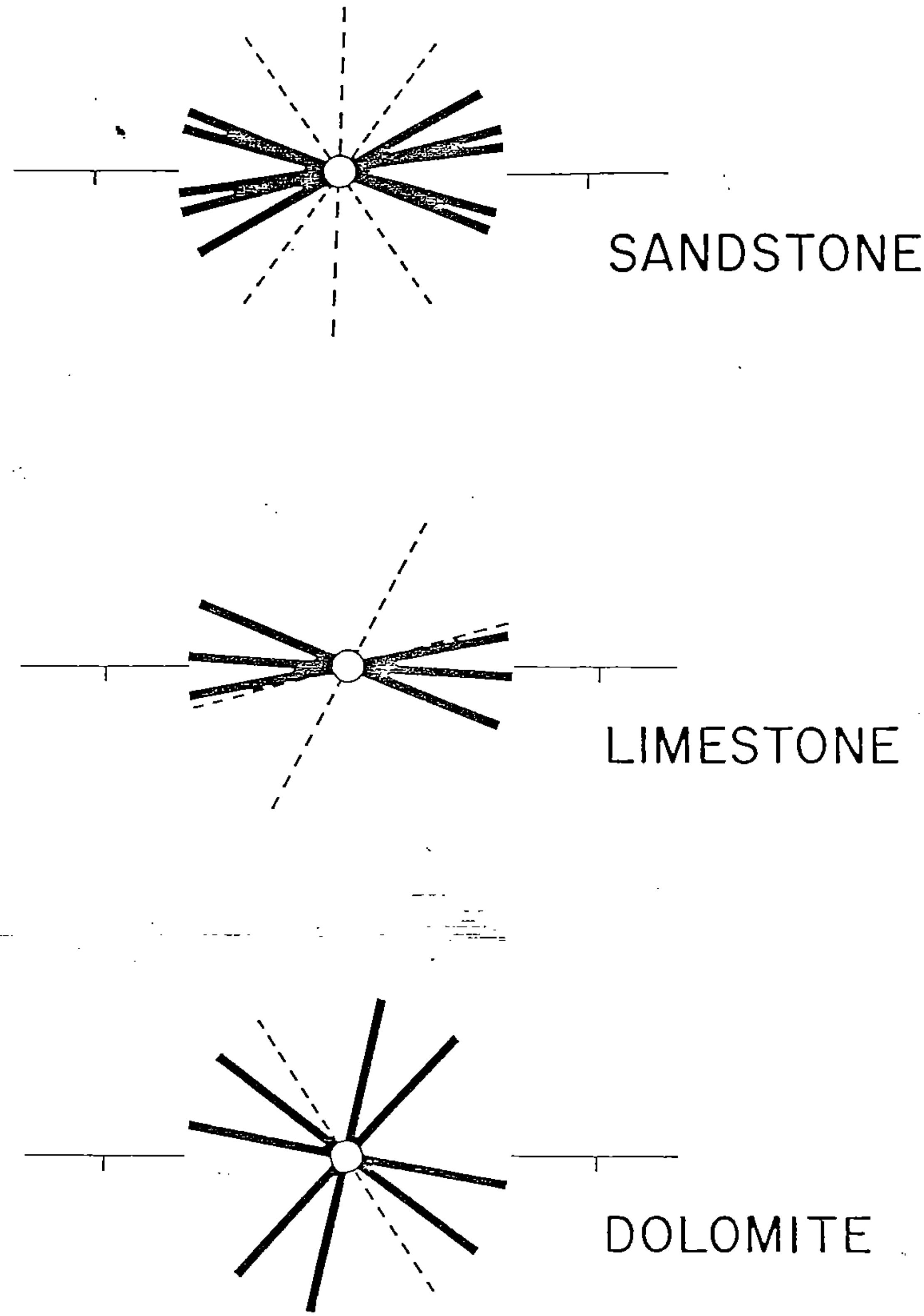


Figure 7. Maximum expansion for limestone, sandstone, and dolomite plotted with the strike of bedding rotated to a common orientation. Solid lines are the orientation of positive maximum expansion and dashed lines are the orientation of negative maximum expansion (contraction).