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Patterns and magnitudes of natural strain in rocks

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Strain values based upon the shape of the deformation ellipsoid as determined from measurements of deformed objects are described from five different tectonic environments. Patterns of strain are shown to vary significantly and systematically depending upon the geological structure. The studied regions are:
(*a*) The Cambrian slate belt of Wales: simple upright and symmetrical folded structures.
(*b*) The Taconic slate belt of New York and Vermont: overturned and asymmetrical folded structures.
(*c*) The Rhodesian Archaean shield: intensely deformed upright and isoclinally folded structures.
(*d*) The Helvetic nappes of the Alps: recumbently folded structures resulting from gravitational gliding.
(*e*) The Moine nappe of northwest Scotland: recumbently folded structures of an orogenic marginal thrust zone.

The finite permanent strains measured in these various environments cover a wide field of possible deformational modes, from almost pure flattening to almost pure constriction. Strain magnitudes are over 1000 % in extreme cases. Deformation by flattening, involving shortening of original dimensions by up to 75 % in one principal direction and concomitant extension of up to 250 % in another principal direction, is characteristic of high-level upright structures and the basal parts of superficial gravitational gliding structures. On the other hand, constrictional deformation involving extensions of between 250 % and 1500 % is characteristic of deep-level upright structures and marginal orogenic thrust fronts.

From the intensities of measured strain and the position occupied by the deformation ellipsoid in the field of possible strains, it should be possible to deduce the mode of origin and depth of formation of many major geological structures.

Introduction

For a complete understanding of the origin of common geological structures such as folds, it is necessary to have information regarding the stresses and resultant strains to which rocks have been subjected. Provided that the state of finite strain is known, some conclusion may be drawn concerning the orientation of the deforming stresses. In order to know something of the processes and mechanisms of natural crustal deformation which has produced an observed finite strain, the conditions under which strain took place must be known. Approximate temperatures may be known from a knowledge of the experimentally determined stability fields of various mineral phases. Similarly, confining pressures and differential pressures can be obtained only by experiment. In order for experimental deformation to be meaningful, it must be capable of producing artificial deformation to the levels of strain observable in nature and at rates of strain which are as realistic as possible.

Knowledge of natural strain rates may be provided by direct measurement of surface strains. These may be rebound rates dependent upon crustal unloading; displacement rates in regions of proven sea-floor spreading; or minimum estimates based upon rates of crustal shortening across orogenic belts. With regard to the latter, there is no general agreement concerning the extent of such orogenic shortening; nor indeed if there is any finite shortening (Shackleton 1969). Geodetic surface measurement of dimensional change is therefore perhaps the most worthwhile approach. It would seem that natural strain rates of 10^{-14} to 10^{-15} s^{-1} may be realistic.

Most large natural strains occurring at moderate to deep crustal levels will be achieved by high-temperature flow and probably take place at reasonably constant rates. Such steady-state flow has been experimentally achieved in marble at 400 °C, confining pressures of 0.5 to 1.0 GPa (5 to 10 kbar), and strain rates of 10^{-8} s^{-1} (Heard 1963; Heard & Carter 1972). It is known however that many strongly deformed materials in the lower metamorphic grades have never been at temperatures above 250 °C (Fyfe, Turner & Verhoogen 1958). This again suggests that natural strain rates are several orders of magnitude slower than 10^{-8} s^{-1}. The greatest need in experimental rock deformation is for long-term tests to simulate more closely the magnitudes of natural strains at strain rates as realistic as possible.

Thus, three distinct approaches are required: the direct measurement of surface strain; experimental rock deformation; and the measurement of natural finite strains. Account must be taken of each of these, which must be integrated in any full understanding of crustal structures. This communication provides new information concerning one of these aspects; namely the magnitudes of natural finite strain states associated with several types of common crustal structures.

Determination of natural finite strain

Materials

Many rocks contain objects of known initial shape, which have been distorted during deformational processes. Their usefulness as strain indicators has been appreciated from the time of Phillips (1843) and Sharpe (1847). The first realistic shape for an ellipsoid of distortion or deformation in naturally deformed rocks was given by Sorby (1855). Subsequently, the strain ellipsoid concept was introduced into geology by Harker (1886). Quantitative studies were undertaken by Heim (1878) and Wettstein (1886) using fossils, and by Cloos (1947) using distorted ooids. More recently, methods have been developed which enable objects of non-spherical and fluctuating initial shape to be used for strain determination (Ramsay 1967; Gay 1968; Dunnet 1969).

In addition to fossils and ooids, the materials available for measurements of finite strain include spherulites, accretionary lapillae, amygdules and vesicles in volcanic and pyroclastic rocks; concretions, nodules, pisolites and conglomerate pebbles in sedimentary rocks; reduction 'spots' in ferruginous rocks; and thermal spots in metamorphic rocks.

Provided that initial sphericity can be demonstrated and that volume changes during deformation have not been in excess of 10 to 15 %, the measured shape changes from equivalent volume spheres will record finite strains which are correct to within 5 %.

Method of plotting strains

Finite strain states are best represented by plotting the deformation ellipsoid for any deformed rock or tectonite on a deformation plot (Flinn 1956, 1962) capable of representing all possible ellipsoidal shapes. Most preferable is a logarithmic plot in which the ratio of the longer to intermediate ellipsoid axes (lg X/Y) is plotted as ordinate and the ratio of the shorter to intermediate axes (lg Z/Y) is plotted as abscissa (figures 2, 3, 5, 7, 9 and 10). Thus all prolate spheroids plot along the ordinate and oblate spheroids of increasing departure from spherical plot progressively along the abscissa. The logarithmic plot has the advantage that similar values of dimensional change in each of the three principal directions (X, Y, Z) compared to the equivalent volume sphere diameter (d) are straight lines.

Cambrian slate belt of Wales

This Caledonian structure is a zone of upright and open folds, some 24 km in length and up to 3 km in width, consisting of Lower Cambrian argillites and interbedded greywackes (figure 1 *a*). Deformation was achieved by a process of flowage during which slaty cleavage became uniformly developed parallel to the axial surfaces of folds. Ellipsoidal reduction bodies occur in the slates and have their principal symmetry plane (X–Y) precisely parallel to the planes of cleavage. It can be demonstrated geometrically that these objects were present before the deformation and that their original form was spherical (Wood 1971). Figure 2 shows the mean finite strain as determined from 30 complete ellipsoid measurements at each of 20 localities, and the orientation of the ellipsoid is indicated on figure 1*a*.

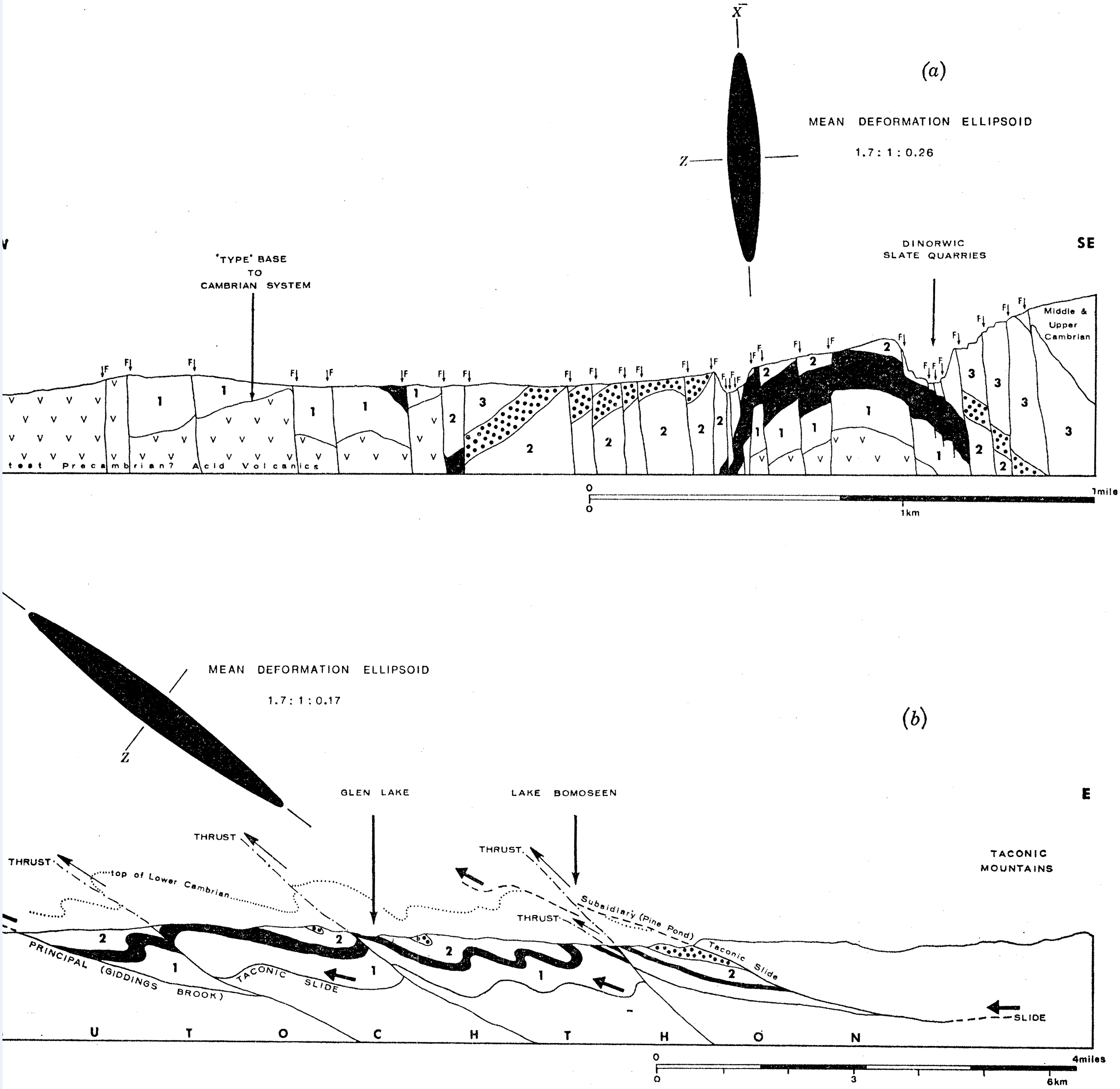


FIGURE 1. Comparative sections of (*a*) the Cambrian slate belt of Wales and (*b*) the Taconic slate belts of Vermont, showing mean states of finite strain. (*a*) Shows three principal Lower Cambrian slate groups (1, 150 m; 2, 400 m; 3, 450 m) separated by greywacke groups; (*b*) shows two principal groups separated by a greywacke group of variable thickness up to 90 m.

Although the mean deformation ellipsoid has the form 1.7:1:0.26, it is seen from figure 2 that there are considerable variations in strain throughout the region. The development of cleavage has entailed a dimensional shortening parallel to the short axis of the ellipsoid (Z) of between 55 % (Z = 45 % d) and 75 % (Z = 25 % d). Subvertical extension of between 60 % (X = 160 % d) and 170 % (X = 270 % d) has occurred. In every case there has also been extension in the direction of the intermediate axis (Y). From the shape and geometry of the ellipsoids it is concluded that deformation involved a process of flattening which resulted in an average dimensional shortening of 65 % across the slate belt.

Taconic slate belt of New York and Vermont

This northern Appalachian structure has its origin in two Lower Palaeozoic events. An exotic sedimentary sequence consisting mainly of Lower Cambrian argillites and turbidites was emplaced upon a time equivalent autochthonous sequence of carbonates and arenites. This emplacement probably took by gravitational gliding during the Middle Ordovician (Cady 1945; Rodgers

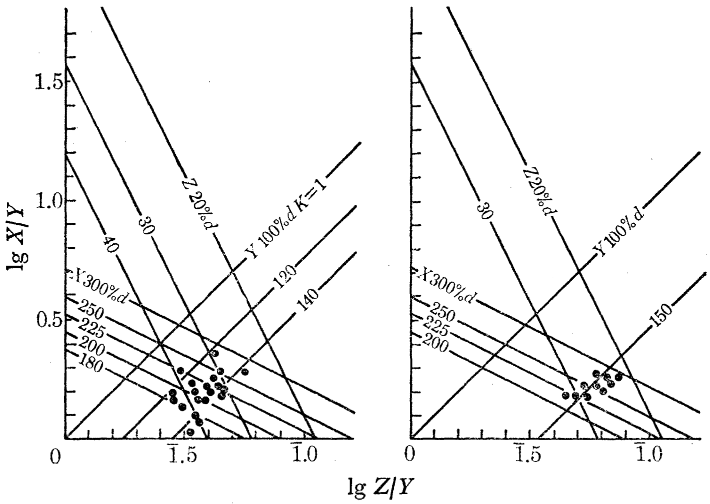


Figure 2. Logarithmic deformation plot for the Cambrian slate belt of Wales. Present ellipsoid shapes are compared to the equivalent sphere diameter (d). ●, Mean of 30 deformation ellipsoids at single localities.

Figure 3. Logarithmic deformation plot for the Taconic slate belt, Vermont, U.S.A. ●, Mean of 25 deformation ellipsoids at each of ten localities.

1951). The whole sequence was deformed together in late Ordovician and Silurian times (Harper 1968). The result is a region of overturned and asymmetrical folded structures with westerly vergence. Deformation involved the ubiquitous production of a slaty cleavage in argillaceous materials. The slates contain ellipsoidal reduction bodies identical to those of the Welsh Cambrian. The contrasting structure of the two slate belts, together with the shapes and orientations of the mean deformation ellipsoids are shown in figure 1.

Measurement of at least 25 strain indicators at each of ten localities reveal that the more extreme structures of the Taconic slate belt were developed in association with a stronger penetrative deformation (figure 3). The mean deformation ellipsoid has the form 1.7:1:0.17.

The average shortening across the cleavage and axial surfaces of folds is 75 % and the average lengthening in the direction of greatest finite extension (X) is 150 % (X = 250 % d) compared to an average extension of 110 % (X = 210 % d) for the Welsh slate belt.

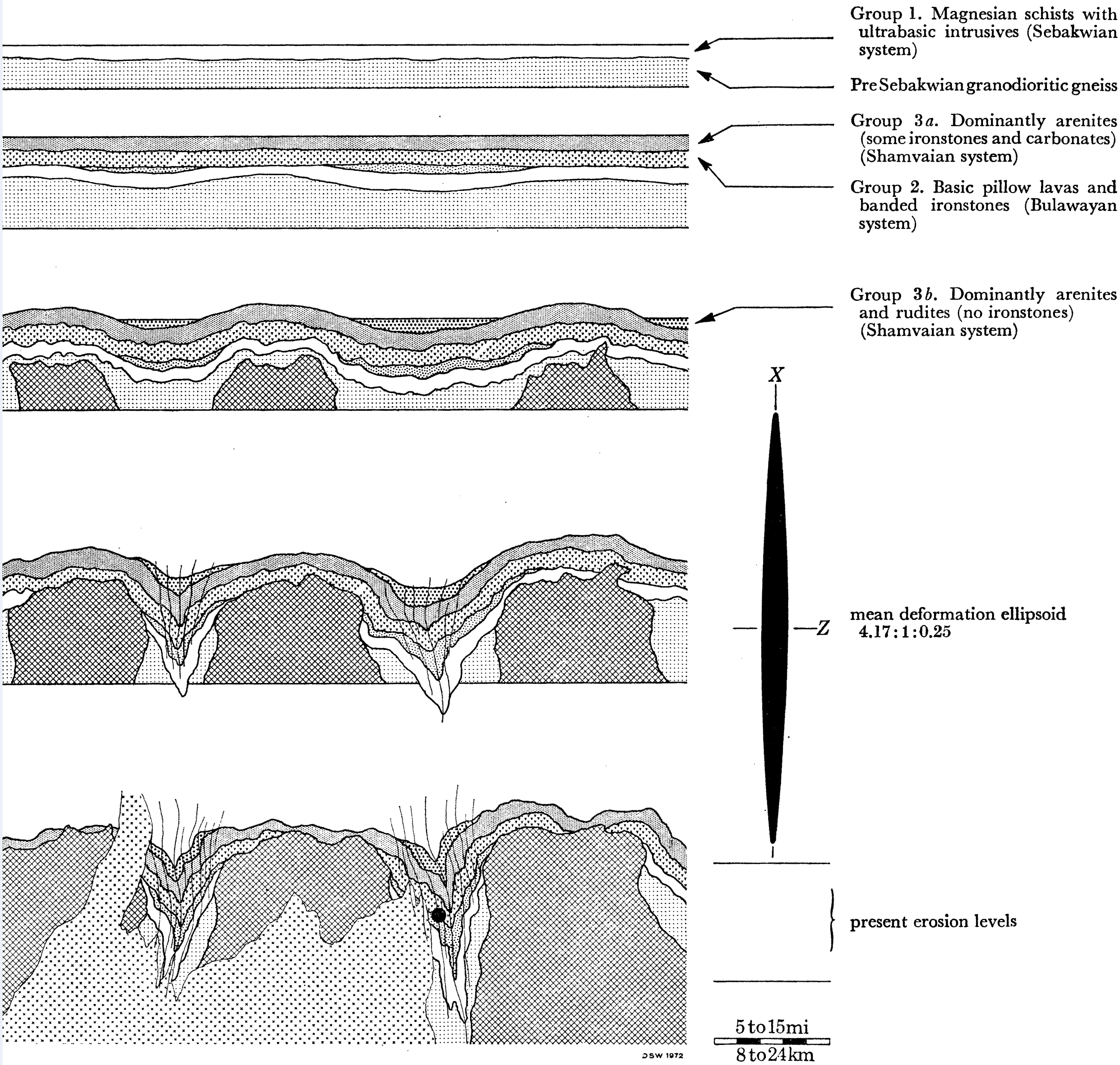
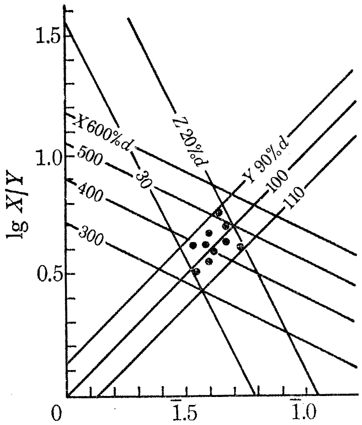
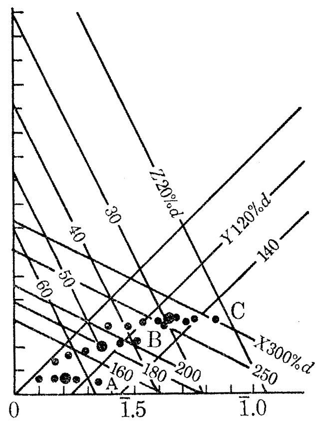


FIGURE 4. Stages in the evolution of a typical Rhodesian greenstone belt. Mean state of finite strain for the Umvuma greenstone belt. Stage 1: Mg-rich ultrabasics (group 1) emplaced at high level in thin granitic crust. Stage 2: warping of granitic basement: local u/c during deposition of group 2; deposition of group 3a. Stage 3: further warping of basement and supracrustal rocks; local u/c between 3a and 3b; phase 1 mobilization of basement; initial intrusion into group 1. Stage 4: rise of diapirs from basement; formation of rim-syncline schist belts; penetrative supracrustal deformation. Stage 5: phase 2 mobilization of basement; intrusion into phase 1 and cover; associated gold mineralization; deepening of rim synclines; continued penetrative deformation.

The Rhodesian Archaean shield

The Archaean structure of the Rhodesian shield is dominated by the presence of narrow and usually deep synclinal greenstone belts generated by the mobilization of underlying granitic basement (Macgregor 1951). This resulted in the formation of partially diapiric granite-gneiss domes with intervening rim synclines which are the greenstone belts. The metasedimentary and metavolcanic rocks involved consist, as elsewhere in Africa, of banded ironstones, basic pillowed lava and arenites. Basement mobilization occurred largely in the interval 3200 to 2700 Ma (Robertson 1968; Vail & Dodson 1970) and consisted, on the evidence of field relationships, of at least two major periods of mobilization. Primary gold mineralization was associated with the





lg Z/Y

FIGURE 5. Logarithmic deformation plot for the Umvuma greenstone belt, Rhodesia. ●, Mean of 35 pebbles at each of ten localities.

FIGURE 7. Logarithmic deformation plot for the base of the Morcles nappe, Switzerland. Paths of progressive deformation indicated for the localities shown on figure 6. ●, Mean deformation ellipsoid from 355 (A), 276 (B) and 288 (C) axial ratios of deformed ooids at three localities; •, points on proposed paths of progressive deformation.

later stages of this cycle (Wood 1966). Stages in the evolution of a typical Rhodesian greenstone belt are illustrated in figure 4. The structures are similar in many details to those of the Barberton Mountain Land of the Transvaal (Anhaeusser, Mason, Viljoen & Viljoen 1969). Together with other similar Archaean terrains, these areas represent an essentially nonlinear thermo-tectonic cycle which is unique in both nature and aerial extent, standing in sharp contrast to the linear thermo-tectonic belts of Proterozoic and Phanerozoic time. Moreover, it was almost certainly an event which affected a thin continental crust which was perhaps three times more extensive than at the present time.

Strain indicators used are pebbles in conglomerate intercalations within the greenstones of the Bulawayan system. The pebbles consist of felsite and quartzite in an amphibole-rich matrix. Some 35 pebble shapes were measured at each of ten localities in the Umvuma schist belt (Bliss 1962), situated at the southeastern margin of the Rhodesdale batholith (Wiles 1957), some 100 miles south-southwest of Salisbury, Rhodesia. All ten localities are situated within an area of less than 2.5 km² occupying a central position in the schist belt. The mean deformation ellipsoid has the form 4.17:1:0.25 and falls close to the median line of the deformation plot, $K = 1$ (Flinn 1962). This situation in which the intermediate axis of the ellipsoid is a direction of no finite length change, indicates, in the absence of significant volume change, deformation by a process of plane strain. The dimensional changes associated with these greenstone belt structures (figure 5) involve a shortening across the fold axial surfaces of 75% ($Z = 25\%\ d$) and subvertical extension in the axial surfaces of 300% ($X = 400\%\ d$). Thus, whereas shortening is comparable to that in the Taconic slate belt, the accompanying extension in the direction of greatest finite elongation is appreciably greater. Of economic significance is the fact that this strong elongation has resulted in separation by boudinage of formerly continuous gold-sulphide reefs which are subparallel or parallel to the fold axial surfaces.

The Helvetic nappes of the Alps

The Helvetic nappes of the high calcareous Alps consist largely of carbonate platform sediments deposited on Hercynian basement along the northern margin of the alpine geosyncline. The nappes themselves were produced by movements which first affected the zone in the Oligocene (Argand 1916).

The explanation of these structures by a process involving a combination of thrusting and gravitational cascade folding is in large part due to Schneegans (1938) and Lugeon & Gagnebin (1941). The mechanism was provided by Hubbert & Rubey (1959) who invoked drastic reduction in shear strength dependent upon the existence of abnormally high pore-water pressures.

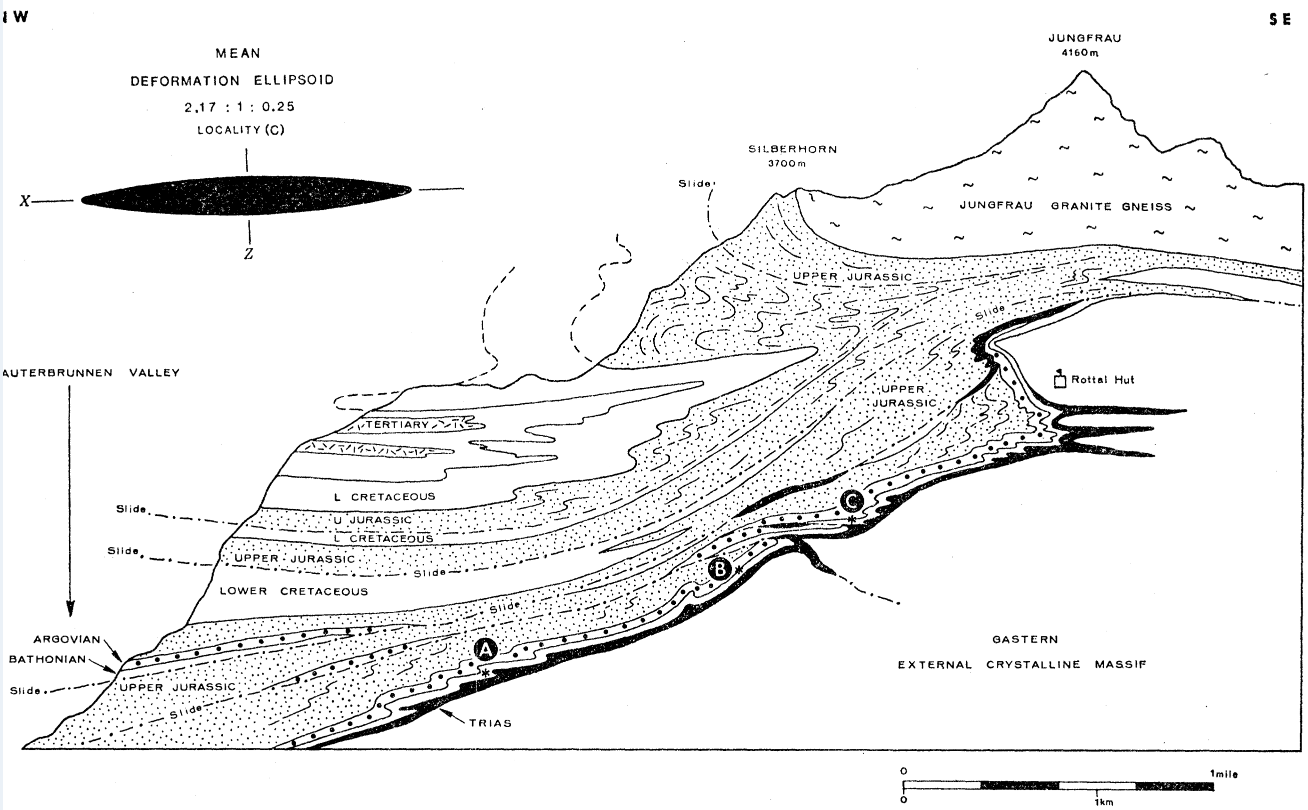


Figure 6. Section of the Morcles nappe showing mean state of finite strain in the base of the nappe.

The Morcles nappe is the lowest and largest such structure and its base comprises an upward passage from autochthonous sediments, through parautochthonous materials into a clearly allochthonous structure (figure 6). Immediately south of Lauterbrunnen, Switzerland, the Morcles nappe has a gneissic crystalline core. Between the latter and the underlying autochthon is situated a recumbent synclinal structure in the lower part of which strain indicators are provided in the form of an oolitic sequence of Bathonian to Callovian age (Collet 1927). The oolites consist of chamosite-hematite ooids set in a sideritic matrix.

The oolites have been sampled at the three localities indicated on figure 6. Some 355, 276 and 288 axial ratios have been determined by microscopic measurement at localities A, B and C respectively. The mean strain values indicated by the three prominent points on the deformation plot (figure 7) differ for each locality. The most significant variation is in the amount of flattening as indicated by the shortening of the Z axis from 30 % (70 % d) at locality A to 50 % at locality B and 70 % (30 % d) at locality C. This is interpreted as the result of differential weight of the overlying nappe. It is clear from figure 6 that the Jungfrau granite-gneiss core of the nappe never extended northward as far as localities A and B. Also indicated on figure 7 are points on proposed paths of progressive deformation for the three localities. These are based upon the variation of strain measured at any locality and the assumption that such variation reflects differing material rigidities in a system in which all components underwent deformation according to a single basic pattern. Thus the low point on curve C was obtained by pairing the 10 % least deformed values of $X:Y$ and $X:Z$ and the high point by pairing the 10 % most deformed values in both planes. Each curve tends towards the origin, and, as might be expected in such a tectonic environment the curves take virtually the shortest path across the lines of progressively increased shortening.

The Moine nappe of northwest Scotland

The Caledonian marginal thrust zone is a region which, from the gross viewpoint, has been subjected to large-scale simple shear, with movement concentrated, but by no means restricted to localized narrow zones. Above the thrust zone, the Moinian metasediments contain strongly deformed conglomerates which are best seen near to the summit and on the southeast slopes of Ben Thutaig, some 5 km to the east of the northern end of Loch Eireboll, Sutherland. The lowest conglomerate horizon, though approximately 600 m above the Moine thrust as placed by Peach & Horne (1907), is in fact only about 120 m above the highest recognizable dislocation surface. The position of the conglomerates, together with the shape of the mean deformation ellipsoid and its relationship to the general structure, is seen in figure 8. In an environment such as this, where deformation intensities are variable, it is the shape type of the ellipsoid rather than the precise strain values which is of significance.

The mean ellipsoid is extremely prolate in shape and has the form 25.0:1:0.9. The fabric of the rock on all scales is dominated by linear elements. Following the terminology of Flinn (1965), the rock is an $L \gg S$ tectonite, having undergone extension as a result of a constrictional deformation process. All folds plunge eastwards parallel to the long axis (X) of the deformation ellipsoid and parallel to the direction of nappe transport. It would appear that during constrictional deformation in which the intermediate axis of the deformation ellipsoid has been appreciably shortened, the most efficient method of achieving such shortening on the mesoscopic scale is by the development of folds with axes parallel to the direction of greatest finite extension.

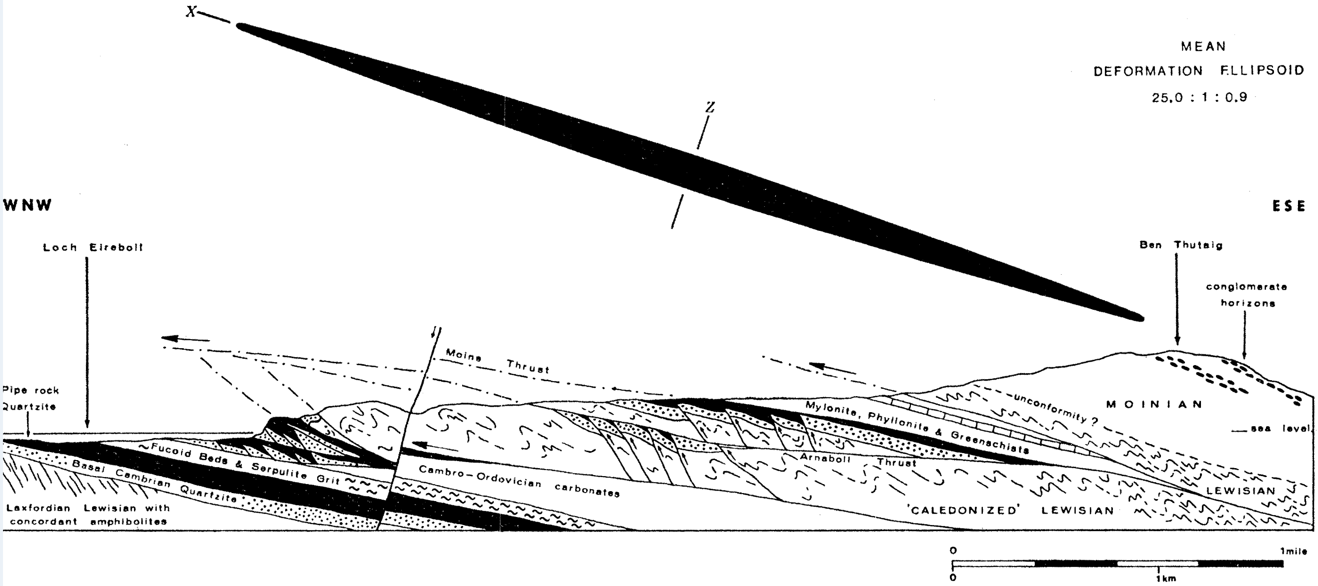


FIGURE 8. Section through the Moine thrust zone showing the position of the Ben Thutaig (Hutig) conglomerate and its mean state of finite strain.

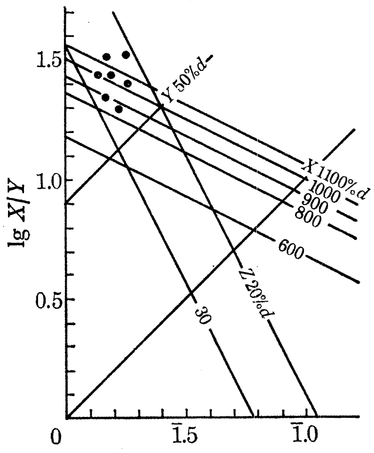


FIGURE 9. Logarithmic deformation plot for the Ben Thutaig conglomerate, Sutherland, Scotland, showing mean deformation from 34 measurements at each of seven localities in the Ben Thutaig conglomerate.

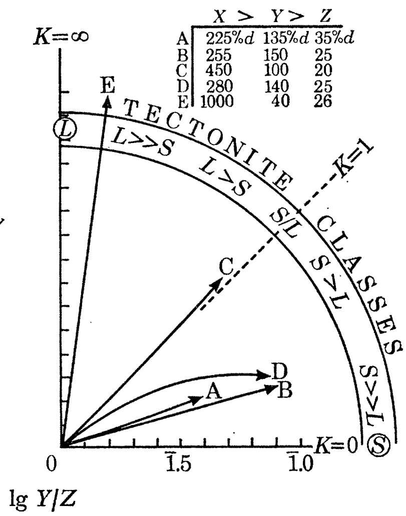


FIGURE 10. Deformation plot showing mean finite strains and tectonite classes for five contrasted environments.

In this case the elongation in the direction of nappe transport is frequently over 1000 %, and shortening of the Y- and Z-directions is of the order of 70 and 75 % respectively. This magnitude of strain and general deformation ellipsoid shape appears to be typical of such tectonic environments. Similar almost prolate ellipsoids occur in the late Precambrian Bygdin conglomerate of south central Norway where extensional strains approaching 3000 % have been measured.

Concluding remarks

The types of structures examined are from five distinct tectonic environments. Simple folded structures are associated with smaller penetrative finite strains than more complex structures. The Cambrian slate belt of Wales, with its simple open upright fold structures has undergone less deformation than has the Taconic slate belt of the United States where folds are tighter, asymmetric and overturned. Both deformations were however achieved by a flattening process, with the result that the planar fabric component is dominant and these, like all slaty rocks, are $S > L$ tectonites (figure 10).

The more extremely folded upright structures from the Rhodesian basement show both higher strain values and fall in a different field of the deformation plot. The process of deformation in this case was one of plane strain and the resulting tectonite (S/L) is one in which both linear and planar fabric components are well developed. On the other hand, the ellipsoid shapes described from the Moine thrust zone are probably typical of the constrictional strains developed close to orogenic margins where accommodation is necessary between a rigid foreland and a more mobile orogenic interior. The five mean finite strain states are summarized on figure 10.

From the position of any deformation ellipsoid on the deformation plot and from a comparison of strain magnitudes with those cited here, it should be possible to draw some conclusion regarding the nature of deformation in any tectonic environment. Given a particular structure, it is now possible to predict the strain and vice versa.

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