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Structural Analysis of Shear Zones in an alpinised Hercynian Granite

(Maggia Lappen, Pennine Zone, Central Alps)

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Abstract

Granite of Pre-Triassic age forms part of the basement core of the Maggia Decke, a lower Pennine recumbent fold nappe in the Swiss Alps. The deformation of this initially isotropic body is characterised by long, narrow, slightly curving shear zones in which a schistosity is strongly developed and high strains are recorded. These shear zones isolate lozengeshaped areas of low deformation. The variation, in magnitude and orientation, of the schistosity was elucidated by mapping at a scale of 1:40.

Strain analysis techniques based on both the variation in orientation of the schistosity and the change in shape of dioritic xenoliths were used to quantify the strains throughout the area. These data support the conclusion that the schistosity, developed in an initially isotropic rock, generally forms in the XY plane of the finite strain ellipsoid.

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1. Introduction

The special aim of this study was to investigate in detail the geometric features of deformed basement granites that resulted from Alpinisation of a Pre-Triassic crystalline basement. These granites were found to be characterized by the development of deformation in local zones, called ductile shear zones. These shear zones are long narrow zones of intense deformation produced by ductile flow of the rock within the zone while the rock outside the zone remained comparatively unstrained. They are developed by a differential displacement of one side of the shear zone relative to the other.

Shear zones, in the sense used here, are the analogues in a ductile material of faults in a brittle material. However, in contrast to faults, the displacement is continuous across the zone and there is no sharp surface of rupture. At a few localities shear zones do pass into faults, but often these faults show mineralogical and structural features which suggest that they may have developed later than the main ductile deformation. For example, some shear zones show a narrow chlorite and epidote filled crack along their centres.

In this study we extend the methods of analysis of shear zones, used by RAMSAY and GRAHAM (1970), by mapping in detail an area where shear zones are the dominant expression of the deformation. To elucidate the geometry of the apparently complex way in which the schistosity varies in intensity and orientation, as seen of the exposures, it was necessary to produce a map on a scale such that all details of this variation could be accurately represented.

We mapped an area of approximately 2300 sq.m. of smooth gently sloping glaciated slabs at a scale of 1:40 using a 1 m square grid marked off in 100 mm units. The mapped area, called the Laghetti area, is 500 m NNW of Pne. dei Laghetti in Canton Ticino, Switzerland (Landeskarte der Schweiz, sheet 265, 68851472), and is of comparatively easy access (motorable road Fusio to 68831480, then hillside walk to 68851473). The map (Plate 1) was prepared at a reduced scale from the field sheets, at this scale objects only a few centimetres across may be accurately represented. References to specific areas are given by grid references, in metres, from an origin at the lower left corner; first figure - left to right, second figure - bottom to top. The area was chosen for a detailed investigation because surface exposures are excellent, and these surface exposures provide sections which are close to true profile sections of the shear zone.

2. General Geology

The area has previously been described by GÜNTHERT (1954) and GÜNTHERT et al (1976), who produced a geological map at 1:25,000, briefly by RAMSAY and GRAHAM (1970) and by ALLISON (1974). HEIM (1922) discussed the regional setting and MILNES (1974) has synthesized new data on the structure of the Pennine zone. The pre-Triassic igneous rocks of the basement core of the nappe have been affected by at least three phases of deformation during the Alpine orogeny. The first two led to large scale ductile flow of basement and cover (RAMSAY and GRAHAM, 1970; HALL, 1972; AYRTON and RAMSAY, 1974). Petrographic and geochemical studies of the metamorphic rocks of the region let GÜNTHERT et al (1976) to put forward a detailed paragenesis, which was supported by U-Pb age determinations of detrital and newly formed zircons by Köppel. They suggested that the area had had a long and complex history with early periods of sedimentation and metamorphism in both Pre-Caledonian and Caledonian time, followed by the formation of the granitic rocks by recrystallisation and local anatexis during the Hercynian orogeny. Subsequent Alpine metamorphism led to a more or less total recrystallisation of the granite, but did not alter the previously formed zircons. They suggested that this implied rather lower Alpine P-T conditions (P, 4-5 kb; T, 550°-600° C) than had occurred in Hercynian times.

We are in general agreement with these results, but we would like to place more emphasis on features suggesting a primary magmatic emplacement of the granite during Hercynian times. We would also like to stress the rather complex nature of the later Alpine deformation of the granite both in the extent of variation of total deformation (and variation in the rock fabric) as well as the polyphase nature of the Alpine events (in three major phases - see AYRTON and RAMSAY, 1974).

The main features suggesting primary magmatic emplacement are as follows:

1. The granitic mass is in general chemically homogeneous (eg. see analysis in GÜNTHERT et al, 1976).
2. The contacts with the surrounding country rocks, both diorites and meta-

Fig. 1: Flow banding of igneous origin parallel to the contact of the granite cut by schistosity of Alpine age, locality (51,09).

morphic paragneisses, are generally sharp. There is always a definable contact and sudden rock type change between the country rock and homogeneous granite over a distance of less than 5 cm.

3. The granite mass contains many xenoliths of dark, mica-rich meta-diorite. All stages can be seen between: (i) massive diorite country rock; (ii) diorite with thin granite dykes; (iii) angular blocks of diorite surrounded by dykes of granite up to 20 cm in width, but still more or less fitting together like a jig-saw puzzle; (iv) completely separated, "free floating", angular, mafic xenoliths from 1 cm to up to 1 metre in granite; (v) isolated, sub-ellipsiodal blocks of mafic xenoliths in homogeneous granite, sometimes showing mineral reaction zones along the granite-xenolith contact. These features are highly suggestive of an intrusive emplacement by wall-rock stoping.
4. Along contacts of the granite which are un-alpinised, flow banding and flow layering are present parallel to these contacts over a distance of up to 1 metre (Fig. 1). This structure is cut by the later Alpine schistosity.
5. The granite is cut by rather regularly oriented (when un-Alpinised) aplite, pegmatite and lamprophyre dyke swarms typical of a cogenetic, magmatic assemblage intruded after the emplacement of the main pluton. An intrusion sequence can be established from intersection relationships: diorite, granite (strictly adamellite), two periods of aplite dykes, two periods of lamprophyre dykes.

Much of the region of granitic rocks described here has been fairly strongly affected by Alpine deformation, and the granites and associated dykes show a schistosity and alignment of minerals and mineral aggregates. However, there are a few localities where almost undeformed granites and dykes can be found and these enable the pre-Alpine geometric features of the mass to be established.

About 1 km south of the area shown in Plate 1, the schistosity becomes more regular in development and orientation and passes into fairly uniform schistose gneisses. Similarly, towards the Mesozoic envelope along the northern contact of the Maggia Decke, the primary igneous features of the rocks are soon lost and the rocks become schistose as the basement-cover unconformity is approached.

3. Evidence for an alpine deformation

All contacts between the meta-igneous rocks in the Laghetti area are sharp. Blocks of diorite can be seen in all stages of detachment from the parent mass (51,09) implying that the granite has intruded by the mechanism of stoping. Where undeformed these dioritic xenoliths retain an angular blocky appearance. That the granite behaved in a brittle manner during the intrusion of the aplite and lamprophyre dykes is suggested by comparing opposite sides of the lamprophyre dykes (36,33 and 54,28) where the same irregularities in the margins are preserved on both sides of the dyke. The en-echelon arrangement of the aplite veins (48,31) is also characteristic of brittle fracture.

RAMSAY and GRAHAM (1970) cited two reasons supporting the conclusion that the deformation was tectonic and did not result from primary magmatic flow, and that it was of Alpine age:

1. The lamprophyre dykes cut the xenoliths and are folded and cut by a schistosity common to the dyke and surrounding rock (compare [06,17] and RAMSAY and GRAHAM, 1970, Fig. 18) "The schistosity therefore must have developed after the emplacement of the plutonic igneous country rocks".
2. The schistosity of these basement rocks can be traced out into the cover of Mesozoic rocks and must therefore be of Alpine age and unconnected with the emplacement of the igneous bodies localised in the pre-Mesozoic basement.

4. Strain analysis techniques

In the Laghetti area it is possible to quantify the state of strain by several different methods and, in some cases the displacements calculated by these methods may be compared with the actual displacement as recorded by deformed markers, such as dykes and igneous contacts.

The strain measurement techniques are mostly based on an analysis of the schistosity variation in simple shear zones (RAMSAY and GRAHAM, 1970), and the shapes of deformed ellipsoidal particles (the $R_f/\varnothing$ method, RAMSAY [1967], DUNNET [1969] and DUNNET and SIDDANS [1971]).

These two techniques are not directly comparable since the first is an interpretation of variations of rock fabric arising from inhomogeneous strains, while the second is a technique applicable to areas in which the strain is homogeneous or nearly so. Although this area is characterised by the inhomogeneous nature of the deformation there are domains of homogeneous deformation of sufficient size and with enough xenoliths to use the second method. The uniformity of intensity and orientation of the fabric is the criterion for selecting areas of homogeneous deformation.

4.1 SCHISTOSITY VARIATION METHOD

The shear zone illustrated in Fig. 2a is exposed on a subhorizontal surface. It is planar, of fairly uniform width and shows a sigmoidally curving alignment of

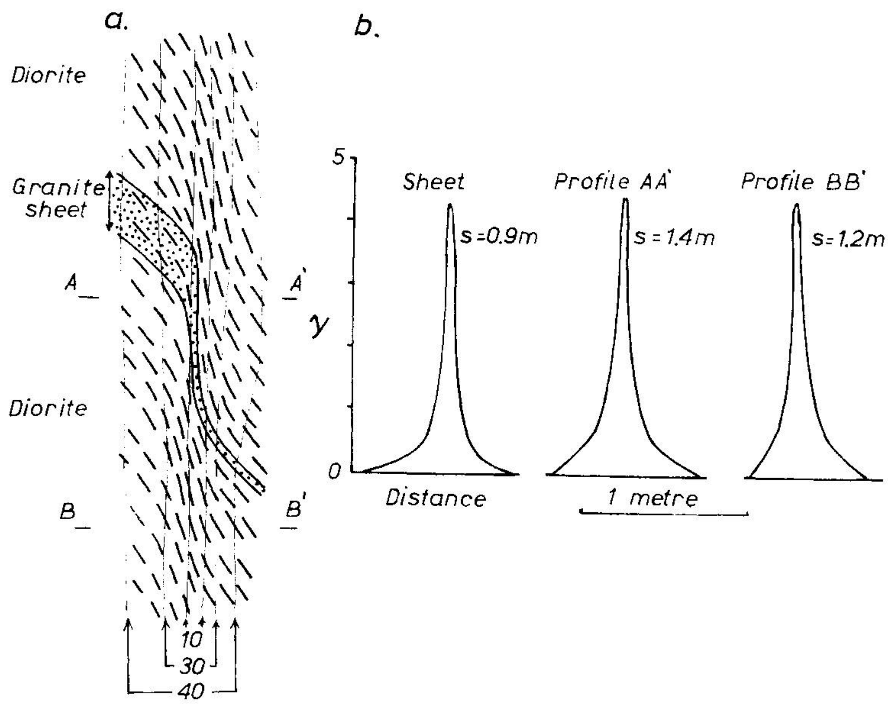


Fig. 2:
a. Shear zone cutting an intrusive granite sheet at locality (49,10); dashed lines indicate trace of schistosity and fine continuous lines represent contours of equal orientation of schistosity.
b. Shear strain (γ) distance plots derived from a geometric analysis of the distorted granite sheet, and from two schistosity profiles. The total shear (s) across the zone has been calculated by integrating variations in γ.

mineral aggregates. Both the deformation fabric and the shear zone are vertical and the surface is a profile section. The shear direction deduced from the orientation of the curving fabric and from the relations of the fabric to the walls of the zone is horizontal. Throughout most of the length, the zone is bounded by rock showing no fabric or a very weak fabric, and because no compositional changes occur that might be suggestive of volume change during deformation, the geometric features of the zone satisfy the boundary conditions of deformation by simple shear (RAMSAY and GRAHAM, 1970). Where the fabric first becomes visible it makes an angle of 45°–40° to the shear direction. Shear strain/distance curves from two sections across the zone are shown in Fig. 2b and the total displacement, computed from the area under the curves, represents a dextral movement of between 0.9 m and 1.4 m across a zone 1.0 m wide.

An independent value of the total displacement has been calculated from the shift of flow banding parallel to the granite-diorite contact. This contact makes an initial angle of $\alpha = 63°$ with the shear zone and using the relation $\cot \alpha' = \cot\alpha + \gamma$ (RAMSAY, 1967, p. 88) where α' is the angle after deformation and γ is the shear strain. The total displacement across the zone calculated from the shear strain/distance curves is 1.3 m and compares reasonably well with the observed displacement of the contact. The slight differences in calculated values of displacement using the two methods probably arise because of difficulties of determining the orientation of the fabric with sufficient accuracy; at high strains slight variations in orientation of the schistosity trace represent large changes in the value of the shear strain.

The applications of these techniques are sometimes limited because the boundary conditions do not always comply with the simple shear constraints. Sometimes the wall rocks of the shear zone are deformed in a non plane strain manner, and sometimes the shear zones are not of constant width and show variations in strain profile along their length.

4.2 $R_f / \varnothing$ METHOD

The dioritic xenoliths are generally sub-angular and approximately ellipsoidal where they are undeformed (47,18) but even at comparatively low strains they can be seen to become preferentially elongated in certain directions (02,18). They are very useful for calculating finite strain values since a shape analysis can be used to give the two dimensional strains on a particular surface no matter what orientation that surface has in relation to the three dimensional

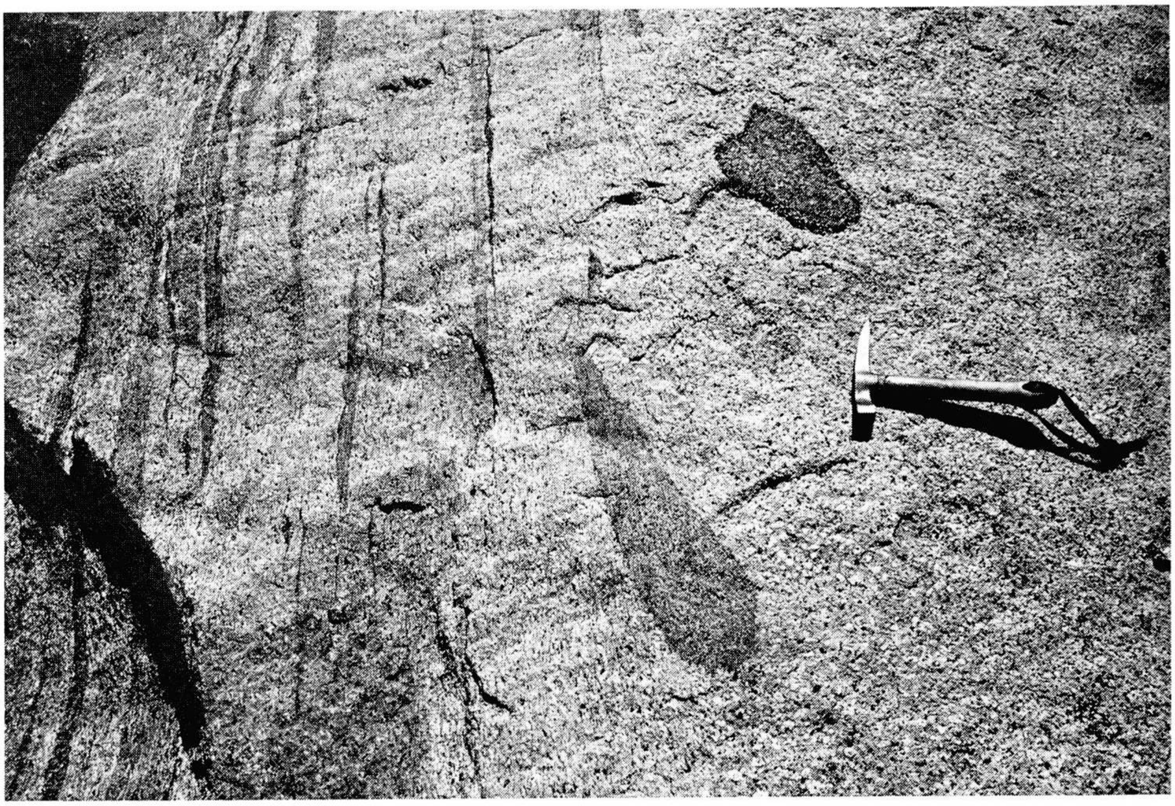


Fig. 3: Strain variation at the edge of a shear zone, locality (58,23) apparent from the shapes of deformed xenoliths and variations in intensity of schistosity.

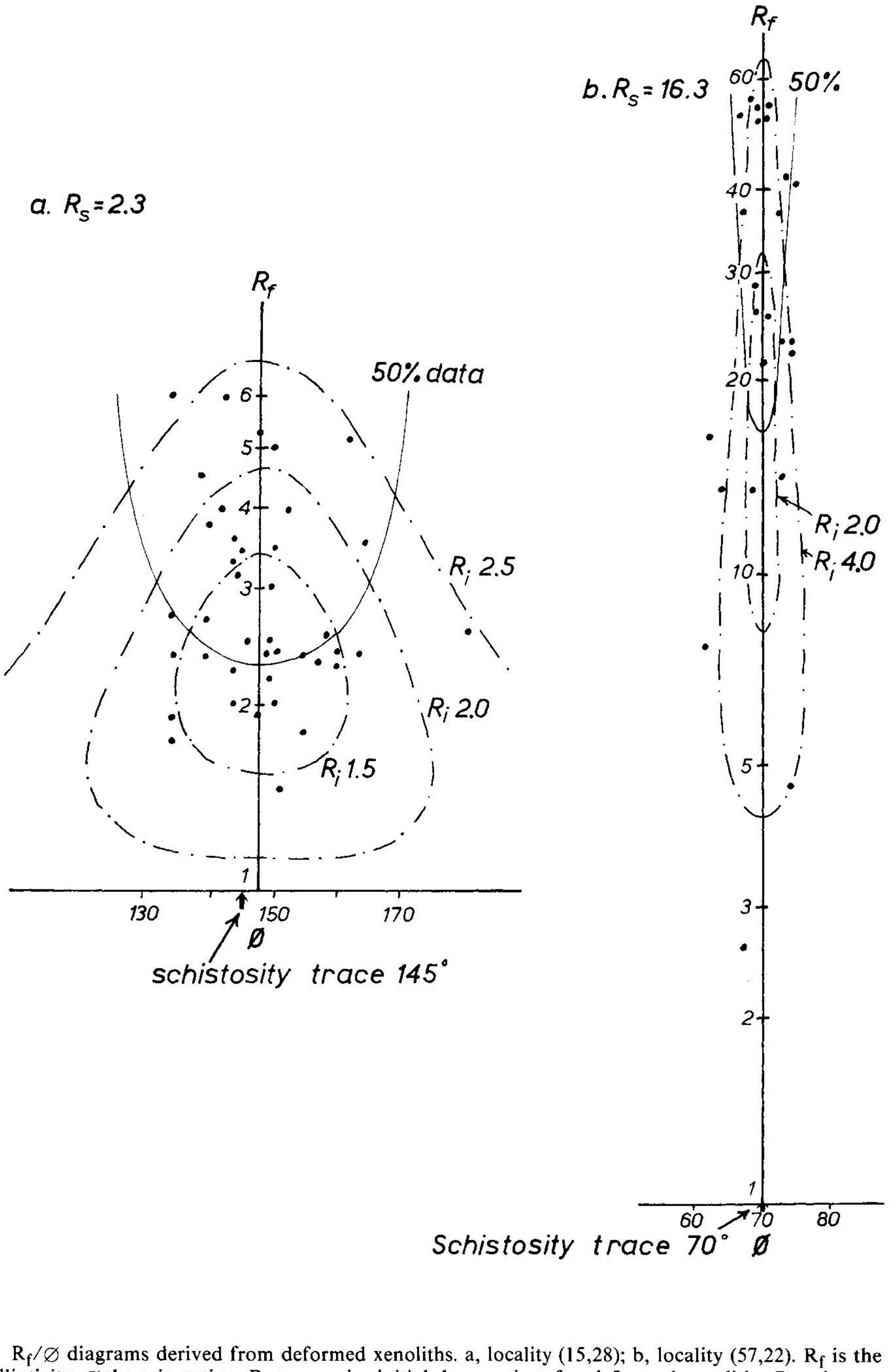


Fig. 4: $R_f/\varnothing$ diagrams derived from deformed xenoliths. a, locality (15,28); b, locality (57,22). R_f is the measured ellipticity, $\varnothing$ the orientation. R_i curves give initial shape ratios of undeformed xenoliths, R_s values are best calculations of the tectonic strain.

strain ellipsoid. At some localities, this technique cannot be used because the values of the principal strains vary too rapidly to obtain sufficient measurements of xenolith shapes over a homogeneous strain field (Fig. 3). The shapes of a minimum of about thirty xenoliths must be recorded before an accurate calculation of the finite strain can be made.

To record the finite strain suffered by the rock it is a necessary condition that there must be no competence contrast between the xenolith and the surrounding granite. In the Laghetti region this condition appears to be met because the deformation fabric is never seen to be deflected where it passes across a granite-xenolith contact. The deformed xenoliths approximate to ellipses on the exposure surfaces. The lengths of the major and minor ellipse axes, and the orientation of the major axis of each xenolith were measured in the field. The ratio, R_f, of the major axis to the minor axis of the elliptical xenolith was plotted against the orientation of the major axis $\varnothing$. The strain on the surface was calculated using the techniques described by DUNNET and SIDDANS (1971). For example, the

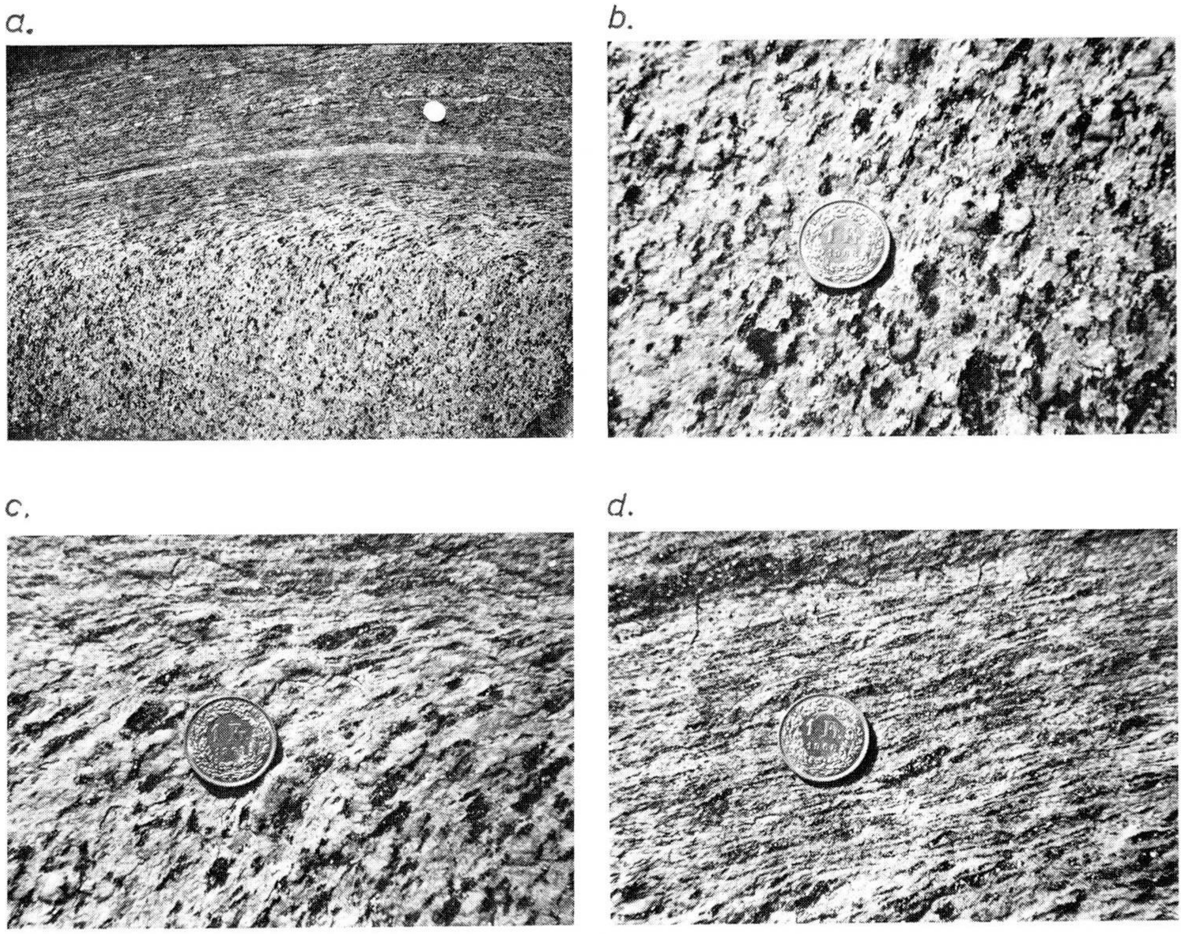


Fig. 5: Variations in fabric across the wall of a shear zone in granite:
a. shows the complete transition, the narrow zone of pale rock is deformed aplite dyke;
b. shows the fabric in the least deformed granite, and
c./d. show the transitional zone and centre of the shear zone respectively.

two dimensional surface strain calculated in this way from the xenolith train (15,28) has a value of $1 + e_1/1 + e_2 = 2.3$ (Fig. 4a). The orientation of the maximum extension direction is parallel to the trace of the deformation fabric. The $R_f/\varnothing$ plot of the deformed xenoliths is symmetrical about the maximum extension direction indicating an initially random distribution of long axes of elliptical xenoliths. This method applied to the highly deformed xenolith train on the edge of the main shear zone (57,22) gives a principal strain ratio of 16.3 (Fig. 4b).

4.3. QUARTZ AGGREGATES AS STRAIN MARKERS

In the undeformed granite quartz grains occur as approximately equidimensional aggregates on any two dimensional surface (Fig. 5b). These aggregates become elliptical within the shear zones (Figs. 5c + 5d). The change in shape and orientation of the aggregates has been recorded by plotting their ellipticity against the orientation of the major axis. The shapes of individual grains and of the aggregates are dependent on several factors including the state of finite strain and mechanism of deformation. It would seem unlikely that the shapes of quartz aggregates are identical to those of the finite strain ellipsoid. However, different aggregate shapes obtained by this method can be compared one with another to indicate qualitatively the degree of deformation.

HALL (1972) studied an area of Alpinised pre-Triassic granitic gneisses in Val di Bosco, 16 km south of the Laghetti area. He described the results of an analysis he made by plotting the ellipticity against orientation of the long axes of deformed quartz grains. He described this method as the $R_a/\varnothing$ - technique. Fig. 6 shows the $R_a/\varnothing$ plots from the three orthogonal faces of a rock chosen to illustrate this technique. The distribution of points is symmetric about the trace of the weak schistosity. The plots indicate that the quartz aggregates were initially elliptical with an axial ratio that rarely exceeded 2:1 and that the long axes of these aggregates initially were distributed randomly in space. The three dimensional R_a ellipsoid plots near the line of prolate ellipsoids on a Flinn diagram and it accords with the dominant linear fabric of the rock.

The variations in two dimensional strains on the sub-horizontal outcrop surfaces which were established using the techniques described above are very complex, and a contoured map is presented in Fig. 7. At a few localities it was possible to investigate the strains in steeply inclined surfaces and thus to establish the three dimensional strain state. At some localities this was of the plane strain type, at others the strains were either apparent constrictions ($k<1$) or apparent flattenings ($k>1$) using the terminology of RAMSAY and WOOD (1973). The reasons for these variations were not always clear because the general two dimensional nature of the outcrops made it impossible to investigate in sufficient detail all the three dimensional variations whereby one strain state passed into another. However, some of the variations appear to be related to the complications of displacement at the terminations of shear zones and a discussion of these will be presented below. Others may relate to complexities of finite strain produced by the superposition of several different phases of deformation or, perhaps more likely, by regional bulk strains of non-planar type going on during the formation of the shear zone.

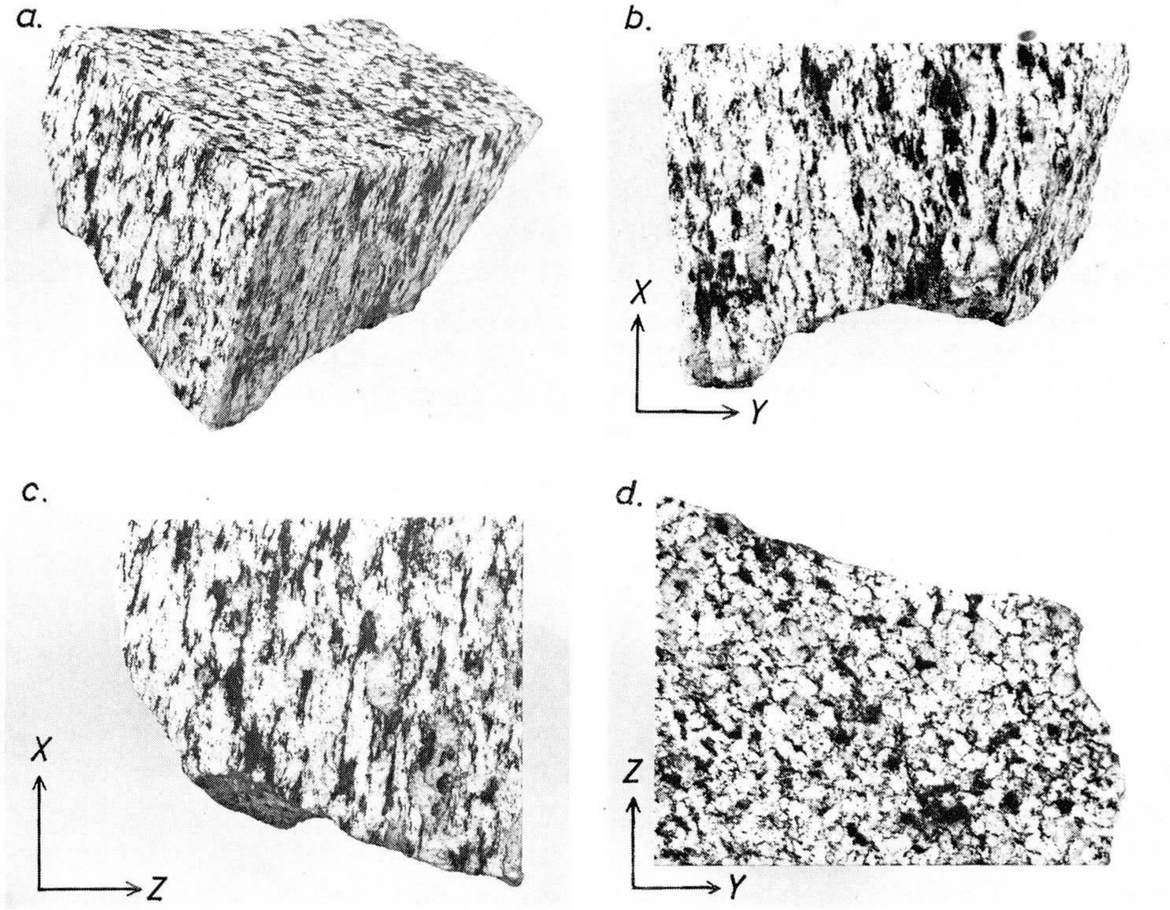


Fig. 6: Three dimensional fabrics in a deformed granite;
a. shows the block with polished faces parallel to the principal strain planes;
b./c. & d. illustrate the appearance of the XY, XZ and YZ faces, and
e./f. & g. show the analysis of the shapes of the mineral aggregates (R_a) on these faces which enables the characteristic finite strain values for each face (R_s) to be determined;
h. shows a Flinn diagram plot of the three principal strains, and the likely field of three dimensional strain.

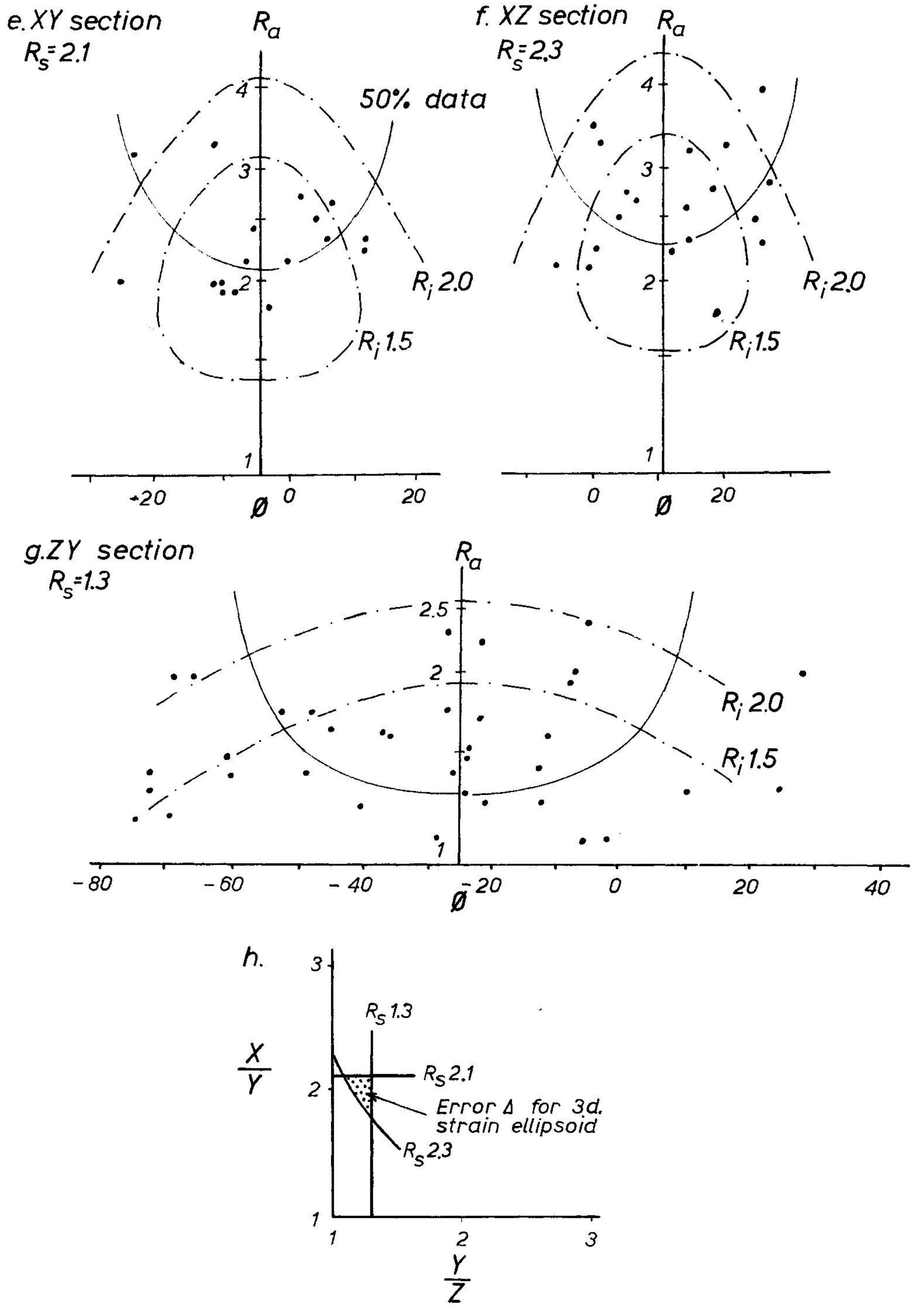

e. XY section
R_s=2.1
R_a
50% data
R_i 2.0
R_i 1.5
-20
0
20
Ø
f. XZ section
R_s=2.3
R_a
R_i 2.0
R_i 1.5
0
20
Ø
g. ZY section
R_s=1.3
R_a
2.5
2
R_i 2.0
R_i 1.5
1
-80
-60
-40
-20
0
20
40
Ø
h.
X/Y
Y/Z
R_s 1.3
R_s 2.1
R_s 2.3
Error Δ for 3d. strain ellipsoid

Fig. 7: Map showing the variations in intensity of two dimensional strains in the outcrop surface. The principal strain ratios in the surface are $x/z = R$ and are not always coincident with the principal X/Z 3-dimensional strain ratios.

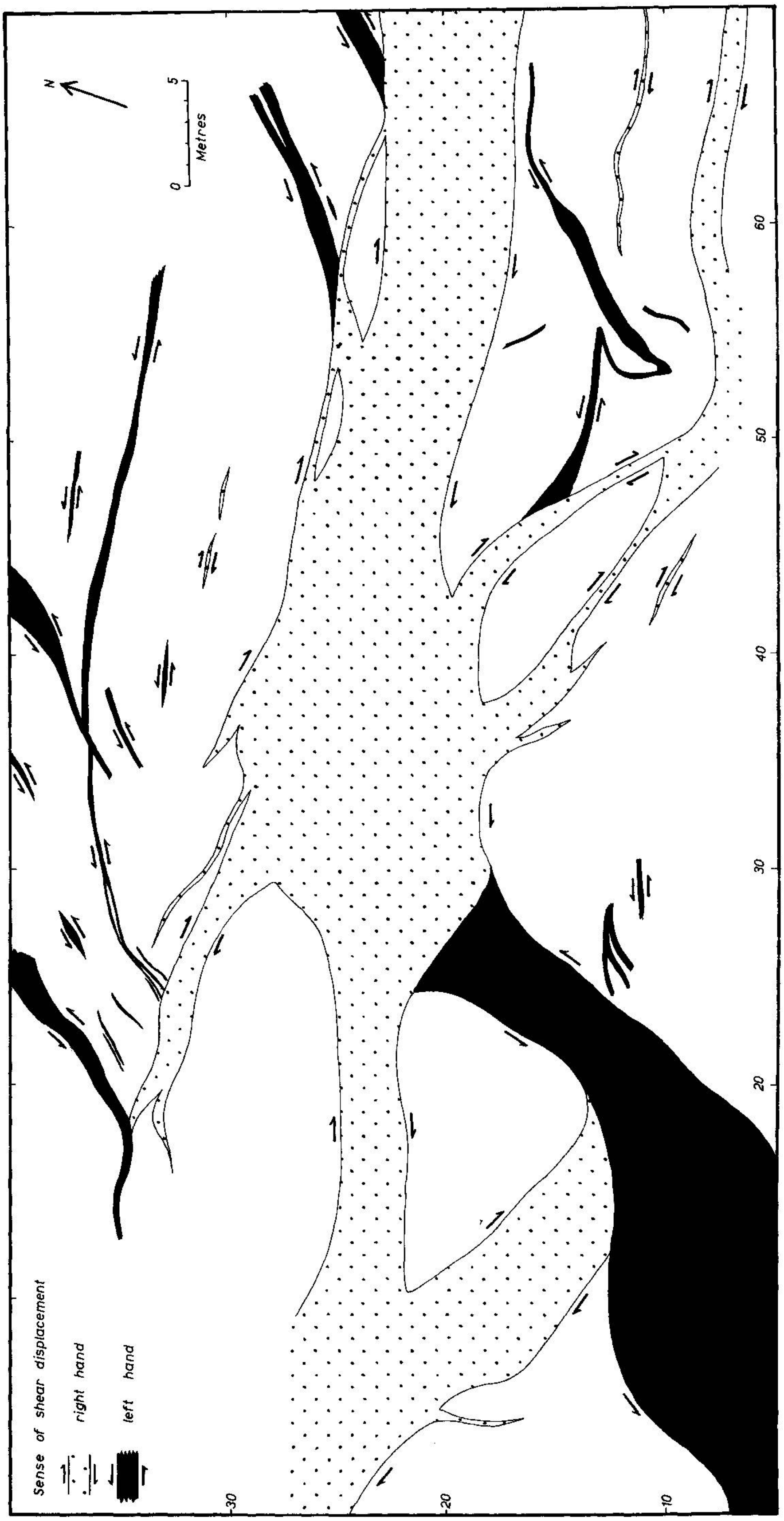


Fig. 8: Displacement sense of the shear zones.

5. Shear zone pattern

The Laghetti area is characterised by shear zones of different orientations and with differing relative displacements between the zone walls, some left-handed others right-handed (Fig. 8). The right-handed shear zones are generally oriented N 80° E, the left-handed N 50 ° E. The two sets of zones intersect each other and the intersection relationships (sometimes 1.h. zone later than r.h. zone, other times vice-versa), and generally continuous nature of the rock fabrics, schistosities and finite strains at the intersections seem consistent with an overall synchronous development of the two sets. The obtuse angle between the two sets faces the maximum shortening produced by the zones, the acute angle is bisected by the maximum elongation produced by the zones.

At some localities differently oriented shear zones with the same displacement sense cross one another. Although these zones generally show a difference in development of timing (eg. 36, 37), the schistosity variation from one zone to the other is not separable into phases, and it seems likely that the zones represent sub-phases within a single main deformation event, perhaps resulting from a change in the direction of the principal strains of the incremental shortenings.

At two localities crossing shear zones show complex relationships at their intersections suggesting a greater separation of deformation sequence compared

Fig. 9: Characteristic "lozenge" shaped masses of relatively undeformed granite surrounded by more platy and schistose rocks of the shear zones seen in the frost shattered cliffs of the North east face of Pizzo del Lago Scuro.

with that normally seen. One (54,12, continuing to 60,14 and 65,16) shows a completely folded zone which is cross-cut by the main schistosity and the regional fabric of the area (Fig. 16). The other (66,24) shows schistosity fabrics related to a left-hand zone intersecting and cross-cutting a schistosity fabric related to a right-hand zone.

The intersecting shear zones give rise to a highly characteristic overall regional structural pattern whereby lozenge-shaped masses of relatively undeformed granite are surrounded by shear zones and separated from each other (eg. the mass with centre 18,17). When one observes the more steeply inclined cliffs above the mapped area, differential frost erosion can be seen to have revealed this structure very clearly (Fig. 9). The schistose zones have been preferentially weathered away to leave a residual pod-like mass of more massif granite.

6. Schistosity variation

Schistosity is produced by the parallel alignment of groups of biotite flakes and aggregates of other mineral grains and a statistically preferred orientation of biotites (Fig. 5). A comparison of Plate 1 and Fig. 7 will show that there is generally a clear relationship between the orientation and intensity of the schistosity trace and the orientation and ellipticity of the two dimensional strain ellipse on the outcrop surface. The areas of very strong schistosity always coincide with regions of high finite strain and areas of weak schistosity coincide with regions of low or no finite strain. The trace of the schistosity almost everywhere coincides with the long axis of the two dimensional strain ellipse (RAMSAY and GRAHAM, 1970) on the outcrop surface.

Plate 1 and Fig. 7 show that the two dimensional deformation pattern is characterised by elliptical or lozenge-shaped areas of low deformation bounded by

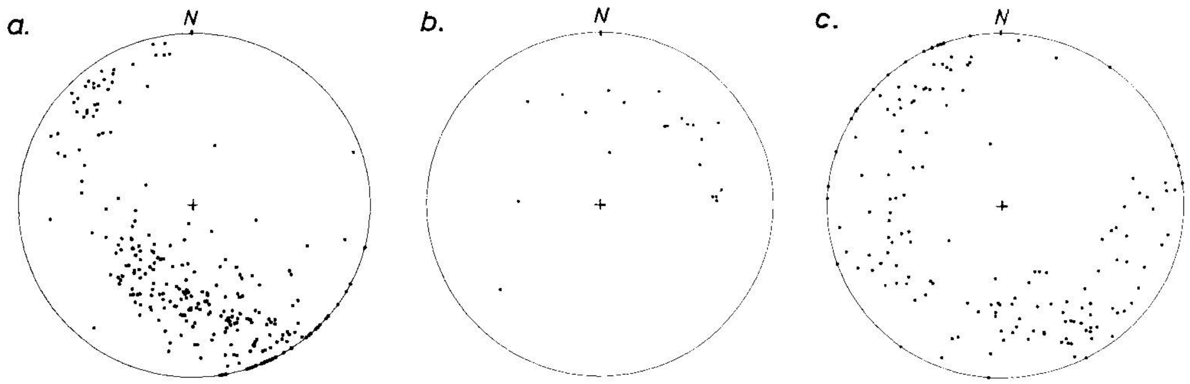


Fig. 10: Equal area projections of structural data;
a. poles of schistosity (263 data points);
b. mineral lineation directions (21 data points);
c. poles to lithological boundaries (165 data points).

curving shear zones in which the rocks are highly deformed. The highest strains and most intense schistosity are always localised in the central portions of the shear zone, and great variations of strain magnitudes occur across the margins of the shear zones (Fig. 5). In a distance of a half metre or less, rocks with strains with principal strain ratios less than 2:1 pass into others with ratios of greater than 10:1 (Figs. 3, 5). Such rapid finite strain gradients often occur over the space of a single xenolith, and the xenolith acquires a characteristic tear-drop shape (Fig. 3).

The three dimensional orientation of the schistosity planes is more systematically regular than the two dimensional surface traces. In the regions of low deformation the schistosity dips at moderate angles towards the northeast. As the shear zones are approached the schistosity steepens and curves into the zone to lie generally vertically in a NE-SW or ENE-WSW (Fig. 10a). The lithological boundaries (Fig. 10b) show variations which are clearly related to the shearing deflections in the shear zones.

6.1 LOW DEFORMATION AREAS

The areas of low deformation are crossed by narrow shear zones. These shear zones are especially interesting in that the variation of schistosity and the way the shear zones die out can be determined.

In the low deformation areas (eg. 52,33) there are small areas of completely unsheared granite with little trace of a planar fabric and differential strain on the surface exposure, but which often show a strongly developed linear fabric steeply inclined to the surface (Figs. 5 and 6). This linear fabric is produced by the prolate form of quartz and biotite aggregates. It seems likely that the development of this linear fabric was contemporaneous with the displacements which went on when the shear zones were formed, because the directions of maximum elongation plot in a field approximately sub-normal to poles of schistosity planes (Fig. 10 a, b). The linear fabric is therefore probably not the resultant of a initial linear fabric that has later been superposed by geometrically unrelated zones of simple shear. Where the wall rocks of the shear zones are deformed, the displacement plan in the zones cannot be ascribed to deformation by simple shear alone and the total strains set up are therefore not plane strains.

6.2 HIGH DEFORMATION AREAS

High deformation areas are characterised by the very intense development of schistosity aligned almost parallel to the shear zone walls. In these zones dykes which are normally highly inclined to the shear zones become sub-parallel to the zone walls (57,22). The dykes often show a great reduction in thickness and an accompanying development of strong schistosity (eg. the lamprophyre passing from 05,17 to 10,18). Aplite dykes often show boudinage indicative of sub-horizontal stretching (eg. aplites passing from 15,19 to 12,18). All these geometric features are consistent with strong sub-horizontal differential movement between the shear zone walls.

6.3 MICROSTRUCTURE

A common feature of shear zones developed in isotropic rocks is a very marked grain size reduction within the shear zone (TEALL, 1885, 1918; WATTERSON, 1975). This effect is not very marked in the Laghetti shear zones although the grain sizes within the quartz aggregates do decrease slightly as the aggregates become elliptical. However, the microstructure does change. Initially it consists of separate aggregates of felsic and mafic components (Fig. 5b). Within the shear zones this separation becomes less pronounced and the rock takes on a more even distribution of phases and sometimes a banding (Fig. 5c, 5d). The fact that there is no marked grain size reduction is probably to be attributed to the main Alpine metamorphism which was partly coeval with and partly post-dated the shear zone formation and which led to the growth of garnet, staurolite and kyanite in the adjacent Mesozoic cover rocks. The present mineralogy of the granite is much like that which would be expected to occur in the original igneous rocks.

Where strains are high the mineral components are often separated into bands, implying that chemical redistribution of the components has taken place. The development of a banded gneissic fabric from an originally unbanded parent granite seems to us to be a geochemical problem of great interest worthy of further investigation.

7. Relationship of structures developed in dykes to strain state and competence

The dykes of aplite and lamprophyre which cut the plutonic igneous rocks are caught up in the deformations which have taken place in the shear zones; they are deflected sideways by the shear displacements and because of competence contrasts between the dykes and their matrix, various characteristic geometric features are set up. The aplite dykes appear to be more competent than the enclosing granite and diorite. When subjected to a finite elongation they become boudinaged (68,17) Fig. 11, dyke A, whereas when they are subjected to a

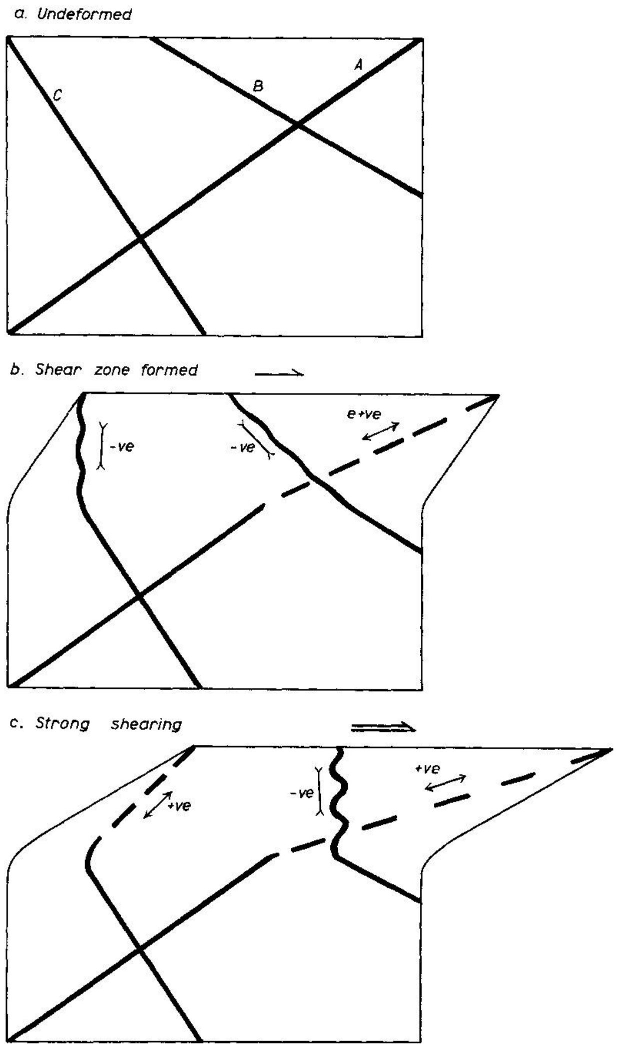


Fig. 11: Instabilities set up in competent aplite dykes as a result of displacements in a shear zone. Dyke A is continuously elongated (extension e is +ve) and becomes progressively boudinaged. Dyke B is continuously shortened (e is -ve) and develops folds whose wavelength and amplitude increases with shearing. Dyke C first undergoes shortening (and is folded) and is then elongated so that the folds are unfolded and eventually boudinaged.

shortening they are buckled into ptygmatic structures (67,16) Fig. 11, dyke B. The development of boudinage or folding in these dykes depends upon the initial orientation of the dyke in relationship to that of the shear zone and upon the sense of movement in the shear zone. In certain orientations the dykes become progressively boudinaged by the shear, whereas with other initial orientations the dykes become folded by the same shear zone (Fig. 11). There are certain critical orientations where the initial shear leads to shortening and fold formation and subsequent extension of the folds to give rise to boudinage or boudinaged folds (68,16) Fig. 11, dyke C.

The behaviour of the lamprophyre dykes is more varied than that of the aplites; the ductility contrasts can be either greater or less than that of the en-

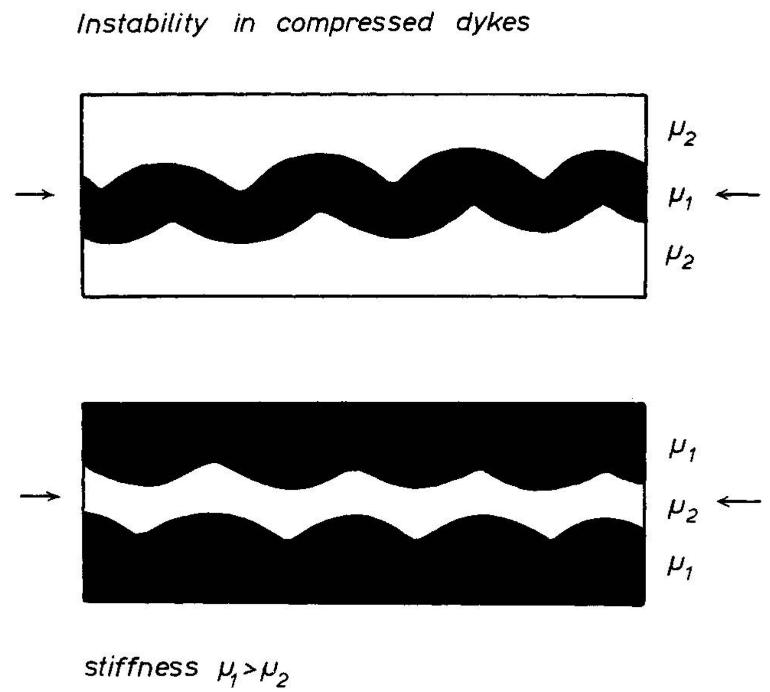


Fig. 12: Characteristic cusp structures developed at the contacts of compressed dykes where the stiffness (apparent viscosity) μ_1 exceeds that of μ_2.

closing plutonic rocks. The reasons for this variation seem to depend upon the mineralogical composition of the lamprophyres. Those dykes that are rich in biotite are generally less competent than their matrix. Whereas those with a high proportion of pyroxene are generally more competent than their matrix. The fold-like instabilities which are set up by these competence contrasts when the dykes are shortened parallel to their length are summarised in diagrammatic form in Fig. 12. Cuspate forms develop at the dyke matrix interface (RAMSAY, 1967, p. 383); these have sharp pointed forms which are directed inwards towards the dyke where the dyke is more competent than the matrix, whereas the reverse is seen where the dyke is less competent than its matrix.

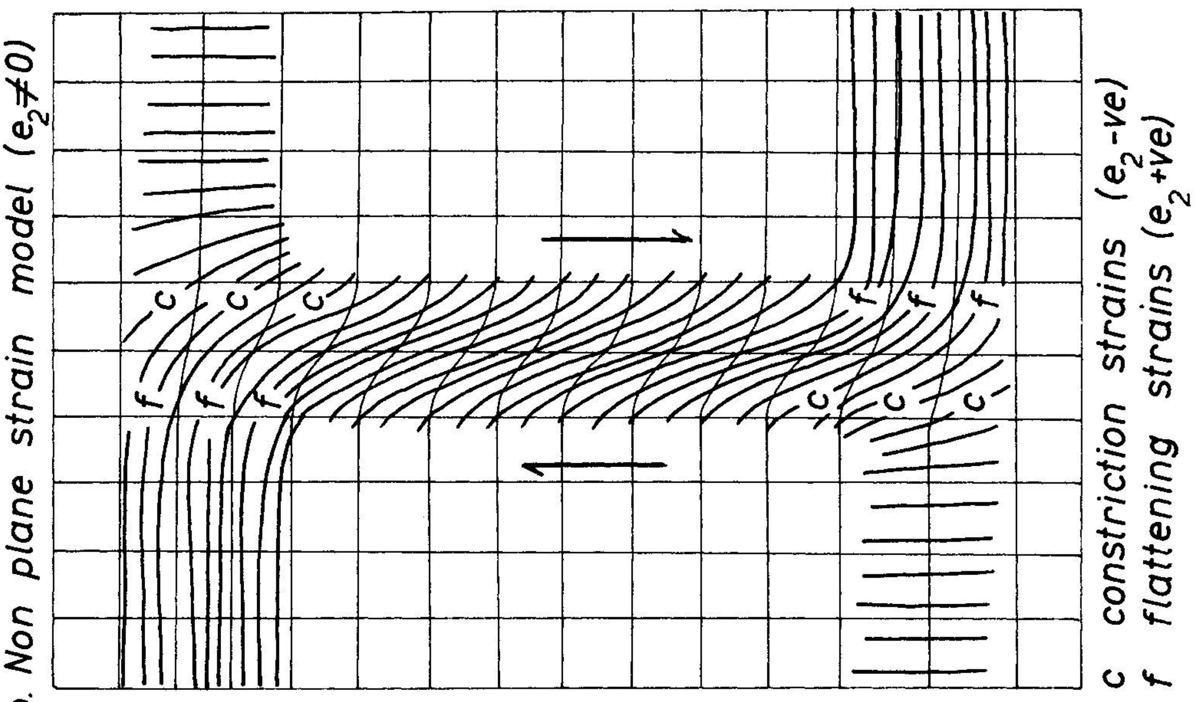
The strain patterns and schistosity variations that are seen around the dykes depend upon how the local strains related to the buckling displacements are combined with strains related to the shear zone deflections. At (67,16) for example, the variations in intensity and orientation of the schistosity around a buckled competent aplite dyke correspond well with the variation expected in the zone of contact strain (RAMSAY 1967, p. 417; 1976; DIETERICH, 1970) with characteristic high strain areas in the inner arcs of the folds and low strains on the outer arcs of the folds (Fig. 13). Much more complex strain variations are apparent where the buckling and shearing are combined and locally the two strains may cancel each other to give rise to a finite neutral point (locality 38,18 where the schistosity radiates in all directions from this neutral point). Perhaps the most striking feature of the schistosity pattern shown in Plate 1 is the great variation caused by different systems of displacement being superposed in a complex manner.

8. Terminations of shear zones

The problem of how shear zones die out along their length and how the large strains of the central parts are dissipated is a complex one. FREUND (1974) presented an analysis of the kinematics of transcurrent faults, with particular regard to their termination. He noted how displacement was spread over an increasingly wide area with the production of splay faults and a bending of individual faults towards the receeding side.

For the main shear zone deformation to be a plane strain the terminal displacements must be spread over an increasingly wider area until the strains are so low that no visible structure forms. Accommodation of the terminal displacements under plane strain conditions can produce complex sideways displacements and consequent complex strain patterns (Fig. 14a, and the shear zone between 24,39 and 20,36). If the plane strain condition is relaxed then

Fig. 13: Folded aplite dyke at locality (67,16). The schistosity pattern around the buckled competent aplite is in accord with the finite strain trajectory pattern.



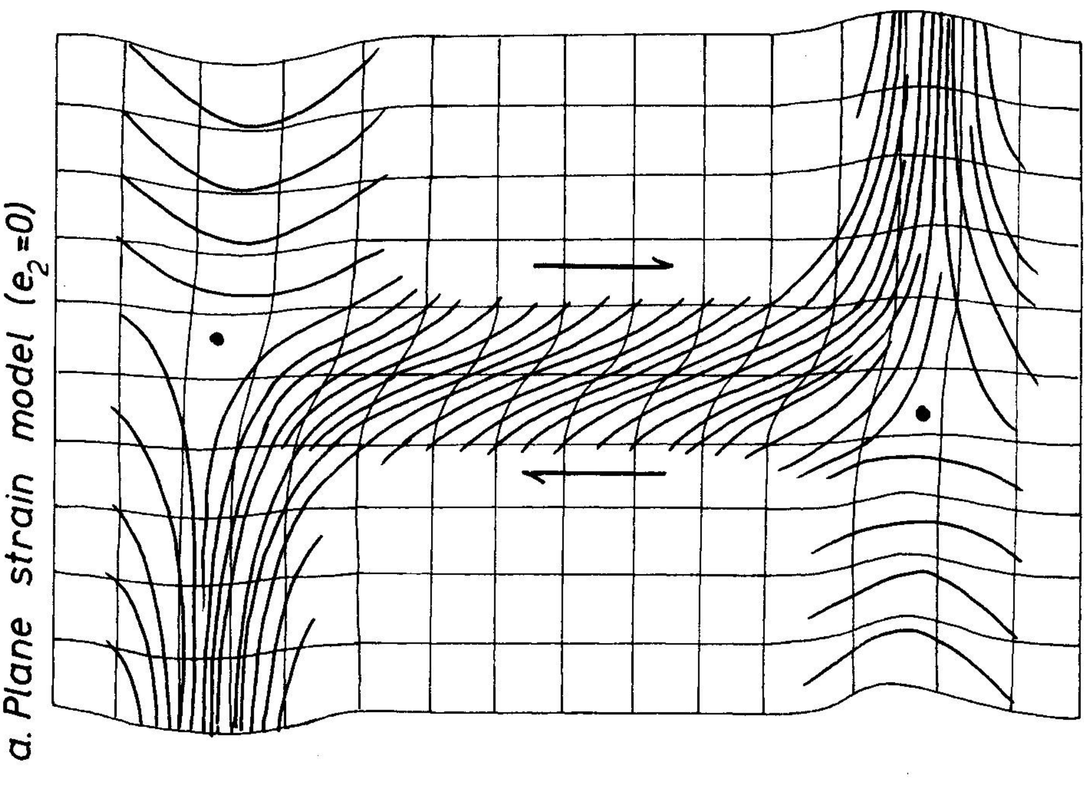


Fig. 14: Types of strain distribution that can arise near the terminations of shear zones;
a. shows a plane strain model where all displacements occur in the plane of the diagram;
b. is a non-plane strain model. The heavy lines represent principal strain directions, points represent isotropic zones,
f. and c. represent flattening (e_2 + ve) and constrictive (e_2 – ve) strains respectively.

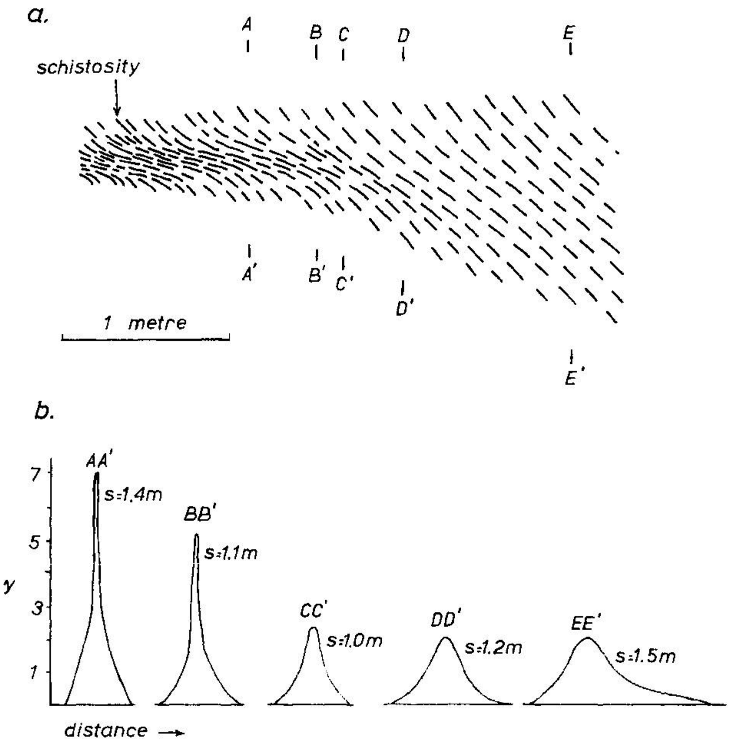


Fig. 15:
a. Variation in orientation of schistosity at the end of a shear zone (locality 41,37).
b. Shear strain (γ) – distance profiles across the shear zone, s is the total displacement across each of the profiles.

shear zones may terminate abruptly producing zones of constriction and flattening strains on either side of the termination (Fig. 14b).

Isogons drawn on the schistosity trace at the end of one shear zone (Fig. 15a, 42,37 and 40,37) clearly show the increase in width of the zone and the associated decrease in the value of the strain. The shear strain/distance graphs (Fig. 15b) show that although the maximum shear strain decreases, the computed displacement remains approximately constant. In this case there can only be a small departure from overall plane strain conditions.

Not only do shear zones widen as they die out but they may also curve (51,14). Right-handed shear zones curve towards the left at their termination is approached, left-handed zones towards the right. The schistosity may be unequally developed at the end of the zone being strong on the advancing side and weak on the receding side (50,14).

9. Intersecting shear zones of differing ages

In the eastern part of the map area a markedly curving shear zone with a well developed fabric can be traced over a length of some 10 metres (54,13 to 55,12). A similar structure appears in a small area north of the main shear zone (65,24). The shear zone is the result of a tectonic deformation and not magmatic flow

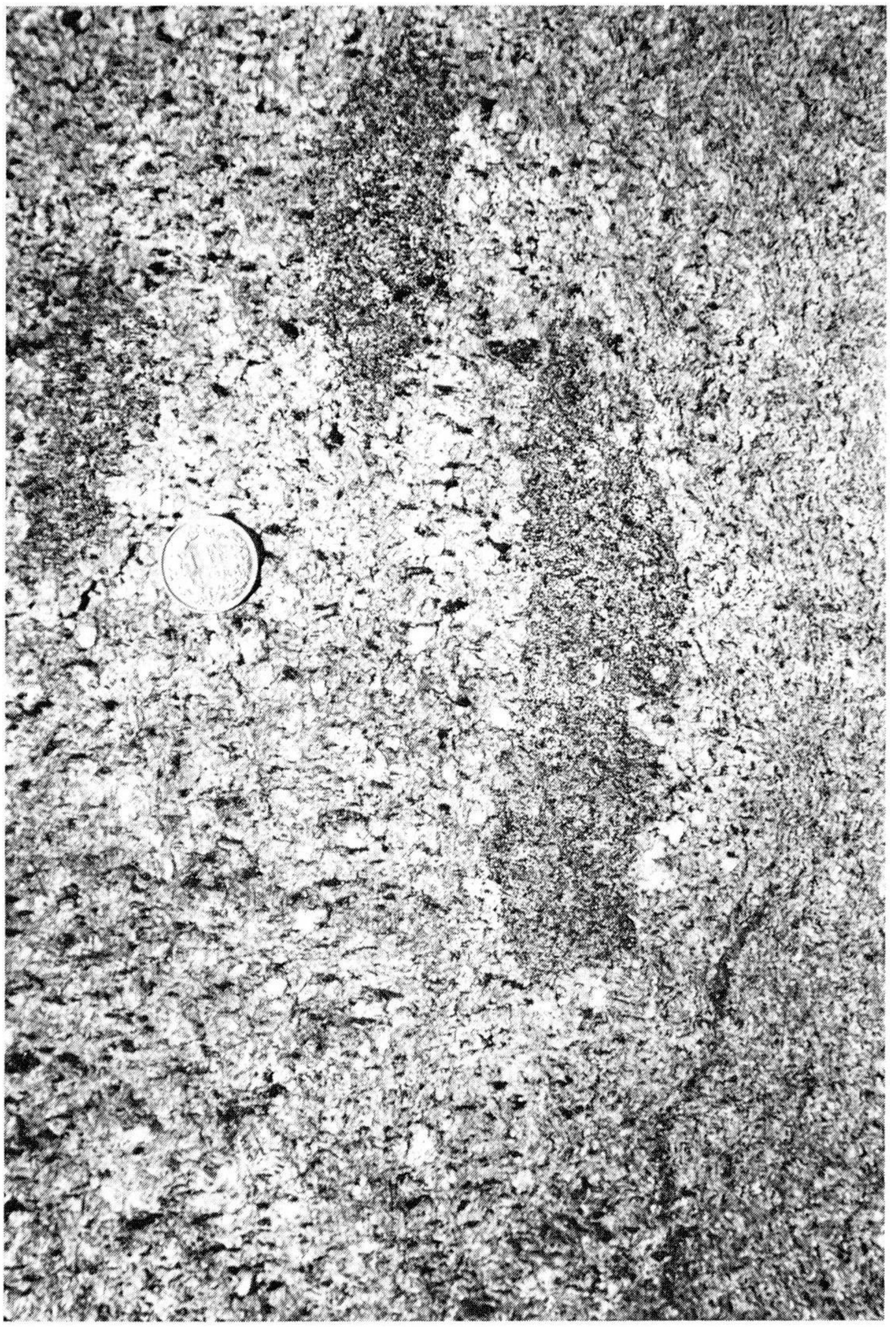


Fig. 16: Xenoliths strongly elongated during an early deformation which have been subjected to a later shortening along their length. The schistosity cross-cuts the finite XY plane (recorded by the long axes of the xenoliths) and is probably parallel to the XY plane of the superposed deformation.

because the granite-diorite contact and the biotitic schlieren within the granite are deformed by it. The dominant schistosity throughout the area cross-cuts the curving shear zone and it appears that the zone is earlier than the main schistosity. Analysis of the xenolith shapes in the early shear zone indicates strain ratios in excess of 4:1. The later schistosity cuts obliquely across the fabric of the earlier zone with no deflection (Fig. 16). At this locality the schistosity does not coincide with the XY plane of the finite strain ellipsoid. Traces of an initial mica fabric occur and the mineral aggregates are crenulated with the development of micro-folds in the elongate lenses of felsic and mafic components. The cross-cutting schistosity arises from the parallel orientation of individual mica flakes. Thus there is only a single true schistosity. In a xenolith from this earlier shear zone the microstructure is related entirely to the later cross-cutting schistosity although the long axis of the xenolith parallels the shear zone. The strains attained during the later deformation were sufficient to produce a new microstructure but insufficient to cancel the extension of the xenoliths parallel to the old shear zone and produce an extension in the direction of the new schistosity trace.

Further east (52,14) along this early shear zone a train of xenoliths becomes parallel to the later schistosity. Qualitatively the xenoliths appear more deformed than the intensity of the schistosity would suggest. The schistosity appears to be related more to the tectonic strain of the later deformation than to the finite strain compounded of both deformations.

10. Shear zones and regional structure

Detailed structural studies in the Pennine Zone during the past decade have revealed large scale, post-nappe emplacement, recumbent folds and even later large backfolds. The existence of these folds necessitates that the traditional view of the structure of the Pennine Zone (for example HEIM 1922) be modified and HALL (1972), AYRTON and RAMSAY (1974), and MILNES (1974) have new models for the structure and metamorphism. The tectonic events of the Pennine Zone may be separated into three major phases viz: -

a. nappe emplacement
b. post-nappe recumbent folding
c. late Alpine folding with the formation of backfolds.

The first two phases led to large scale ductile flow of the basement and cover rocks. Although the nappe emplacement was associated with complex deformation variations, the dominant schistosity appears to be related to a phase of postnappe recumbent folding.

The formation of the schistosity probably occurred during the second Alpine deformation phase, although it is also possible that it may be a composite structure combining both first and second phases. The only structure in the mapped area that predates this main shear zone formation is that of the "earlier shear zone" (section above) but elsewhere in the Maggia nappe first phase folds can be seen to be cross-cut by the main schistosity. Particularly fine outcrops illustrating these relationships can be seen along the North-East ridge of Cristallina at 68511470. It seems possible that some of the rocks in the nappe core did not undergo much strain while being displaced many kilometres northwards during first phase nappe formation.

The formation of the third phase late Alpine folding, led to the backfolding of previously formed structures and to the development of a crenulation of the previously formed schistosity. At Laghetti this crenulation cleavage is not uniformly developed. It is best developed within the earlier strongly foliated rock of the shear zone and poorly developed or absent in the unsheared granite.

11. Conclusions

As a result of detailed mapping of the geometric features of Hercynian granite and diorite bodies, their associated dykes and enclosed xenoliths it was possible to establish a picture of the deformation characteristics of the Alpine events associated with the formation of, and later history of, part of the Pennine nappe terrain of Central Switzerland.

1. Alpine deformations have led to a predominant pattern of intersecting ductile shear zones. The strains in these zones are generally much higher than in the zone walls, and the overall geometry is best explained by a differential simple shear displacement between the zone walls, superimposed on, or developed at the same time as, other non-plane strains outside the zones.
2. The shear zones characteristically occur in conjugate sets with right- and left-handed displacement sense. They intersect and merge to give rise to an overall structural pattern of relatively undeformed lozengeshaped masses surrounded by highly deformed shear zones. In contrast to brittle faults it is the obtuse angle between the conjugate zones which faces the greatest regional shortening inducing the shear displacements.
3. Shear zones with a similar displacement sense sometimes cut each other and suggest a change in the orientation of the shear system with progressive deformation.
4. A few shear zones showing markedly differing geometric characteristics appear to indicate succesive development with large changes in the strain history between the formation of the different zones.
5. The schistosity and tectonic fabric developed in the granite and dyke rocks is practically everywhere related to the finite strain state as deduced from an analysis of xenolith shapes and dyke displacements. Schistosity is generally formed normal to the direction of maximum total shortening (parallel to XY in finite strain ellipsoids with $X \geq Y \geq Z$). Schistosity increases in intensity with deformation and it shows very characteristic curved sigmoidal patterns

which accord with the variations of finite strain trajectories. The strain variations which arise from the movements along conjugate intersecting shear zones show very great variations of schistosity orientation over a comparitively small area. In highly deformed granites a banded gneissic structure is developed parallel to the schistosity.

Although these general conclusions on the significance of schistosity accord with those resulting from previous work in the area (RAMSAY and GRAHAM, 1970); a study of two localities shows that exceptions do occur to this general rule. At one locality two crossing schistosities are found and it is suggested that each schistosity relates to a differently orientated deformation event. At another locality where an early shear zone is strongly folded by displacements along another (see 4 above), the main schistosity crosses the XY plane of the finite strain ellipsoid at high angle. This schistosity appears to have been generated during the strains associated with the formation of the later shear zone.

6. A study of the geometric forms of dykes and dyke-country rock contacts enables differences of competence of the rock types to be established. A competence contrast sequence has been established: -

Most competent	Pegmatite
	Aplite
	Pyroxene bearing lamprophyre
	Granite ≈ Diorite
Least competent	Biotite bearing lamprophyre

Rocks with high proportions of feldspar and coarse grain size appear more competent than rocks with high proportions of biotite and with fine grain size.

7. The geometric features of the shear zones are in general accord with a concept of strain softening during deformation, that is, as finite deformation becomes stronger the rock becomes progressively easier to deform. Although the minerals in the shear zones appear to have been quite strongly recrystallised after deformation, some indications are present that suggest that strain softening may have been accomplished by grain size reduction during deformation. There are no mineralogical features indicative of softening by the addition of chemical inputs from far outside the shear zone systems, although there are indications of local chemical migrations (eg. production of banded gneissic granites from non-banded parent rocks).

8. Variations of strain at the terminations of shear zones are analysed and models based on plane strain and non-plane strain mathematical models are proposed (Fig. 14) which accord best with our field observations.

12. Acknowledgement

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13. References

ALLISON, I., 1974. A petrofabric investigation of shear zones from the Swiss Alps and North West Scotland. Unpub. Ph.D. Univ. London.

AYRTON, S.N. and RAMSAY, J.G., 1974. Tectonic and metamorphic events in the Alps. Schweiz. mineral. petrogr. Mitt. *54*, 609-639.

DIETERICH, J.H., 1970. Computer experiments on mechanics of finite amplitude folds. Canadian Jour. Earth Sci. *7*, 467-476.

DUNNET, D., 1969. A technique of finite strain analysis using elliptical particles. Tectonophysics, *7*, 117-136.

DUNNET, D., and SIDDANS, A.W.B., 1971. Non-random sedimentary fabrics and their modification by strain. Tectonophysics, *12*, 307-325.

FREUND, R., 1974. Kinematics of transform and transcurrent faults. Tectonophysics, *21*, 93-134.

GÜNTHERT, A., 1954. Beiträge zur Petrographie und Geologie des Maggia-Lappens (N.W. Tessin). Schweiz. mineral. petrogr. Mitt. *34*, 1-159.

GÜNTHERT, A., W.B. STERN UND H. Schwander, 1976. Isochemische Granitgneissbildung im Maggia-Lappen (Lepontia der Zentralalpen). Schweiz. mineral. petrogr. Mitt. *56*, 105-143.

HALL, W.D.M., 1972. The structural geology and metamorphic history of the Lower Pennine Nappes, Valle di Bosco, Ticino, Switzerland. Ph.D. Thesis, University of London.

HEIM, A., 1922. Geologie der Schweiz. vol.2. Chr. Herm. Tauchnitz. Leipzig.

MILNES, A.G., 1974. Structure of the Pennine Zone (Central Alps): A new working hypotheses. Bull. Geol. Soc. Am. *85*, 1727-1732.

RAMSAY, J.G., 1967. Folding and Fracturing of Rocks. McGraw-Hill, New York.

RAMSAY, J.G. and WOOD, D., 1973. The Geometric effects of volume change during deformation processes. Tectonophysics *16*, 263-277.

RAMSAY, J.G., 1976. Displacement and strain. Phil. Trans. R. Soc. London A 283, 3-25.

RAMSAY, J.G. and GRAHAM, R.H., 1970. Strain variation in shear belts. Can. J. Earth Sci. *7*, 786-813.

TEALL, J.J.H., 1885. The metamorphosis of dolerite into hornblende schist. Q. Jl. Geol. Soc. London *41*, 133-145.

TEALL, J.J.H., 1918. Dynamic metamorphism. Proc. Geol. Ass. *29*, 1-15.

WATTERSON, J., 1975. Mechanism for the persistence of tectonic lineaments. Nature *253*, 520-521.

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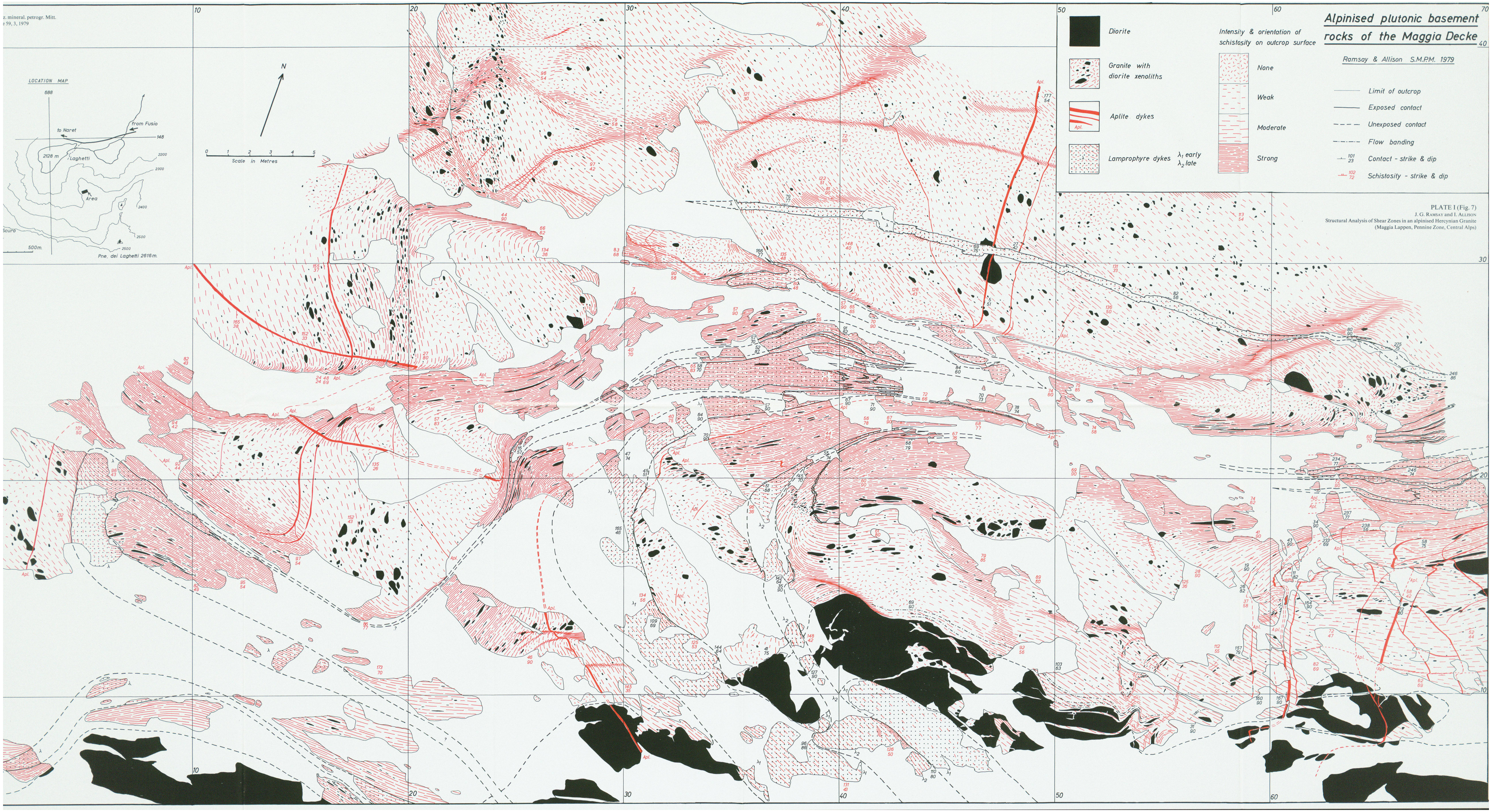

PLATE I (Fig. 7)
J. G. RAMSAY and I. ALLISON
Structural Analysis of Shear Zones in an alpinised Hercynian Granite
(Maggia Lappen, Pennine Zone, Central Alps)