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# Brittle-ductile shear zones northwest of the Grenville Front Mylonite Zone, Killarney, Ontario.

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By

Suzanne Nacha

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## Abstract

Small shear zones located northwest of the Grenville Front Mylonite Zone exhibit both a brittle and ductile deformation history. Textures reveal that an earlier mylonitic rock has been overprinted by one which demonstrates textures typical of brittle cataclasis.

ductile deformation has occurred under greenschist facies conditions, while a later, brittle event has occurred below lower greenschist temperatures. These produce high shear strain values which lie between 14.44 and 10.79.

Upper and lower age limits for the initiation of shear zones have been determined as being prior to the emplacement of pegmatite dykes, and up until the formation of brittle-ductile shear zones found locally. Thus, they have developed between 1400 ± 50 Ma and approximately 1100 + ? Ma.

## Acknowledgments

I would like to express my thanks to the following people to whom I am indebted:

Thanks to Dr. Clifford for supervising this thesis. Thanks also to Jack Whorwood for his photographic work and Len Zwicker for preparing thin sections. Lastly, thanks to my sister Deanna for typing, Joe Boyce for drafting and Lisa Mech for doing my school work for me.

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# CHAPTER 1. INTRODUCTION

## 1.1 Location/Access

The study area, a series of islands located east of Killarney, on Georgian Bay, is situated just south east of the entrance to Collins Inlet (at South Point) at approximately 45°58'20" N latitude and 81°23'10" W longitude (figure 1). The area is easily accessed by boat down Collins Inlet to South Point and then travelling southeast approximately 2.5 km, along the coast of Georgian Bay.

Many small islands in the immediate area were examined; however two particular islands became the focus of the study. Their excellent exposure and concentration of various geologic structures made them ideal for the research topic.

## 1.2 General Geology

The study area, including the immediately adjacent parts of Killarney Provincial Park has been considered part of the Grenville Province (Frarey, 1985). This is the youngest geological province of the Canadian Shield (≈ 1Ga) and is exposed in Labrador, Quebec, Ontario, and adjacent New York State, U.S.A..

In the immediate vicinity of the study area rocks of the Killarney Complex (Davidson, 1986) are exposed. This complex comprises two distinct units. The study area, which is located in the southeastern portion of the complex, is composed of a foliated porphyry which may be pyroclastic in origin. This unit runs approximately southwest to northeast. To the northwest, it comes into contact with the Killarney granite, the division between the two units being based on the change in grain size and the appearance of foliation in the porphyritic felsic unit. Further to the northwest, the Killarney Complex has a steep and very sharp intrusive contact with rocks of the Huronian Supergroup.

Less than a quarter of a kilometer southeast of the study area, the porphyritic felsic unit is terminated by the Grenville Front Mylonite Zone. Running approximately northeast to southwest, this zone is flanked on the southeast by a protomylonitic granitoid gneiss grading into layered gneisses.

The Grenville Front Mylonite Zone divides the area into two structurally distinct parts. A major change in structural age, style, and orientation is seen from one area to the other. On the southeast side of the Front, structural features can be correlated to the Grenville event. To the northwest, however, structural data record evidence of an earlier deformation event. The study area is located to the northwest of the Front, and shows geologic features which are of a different structural style than the Grenville Front

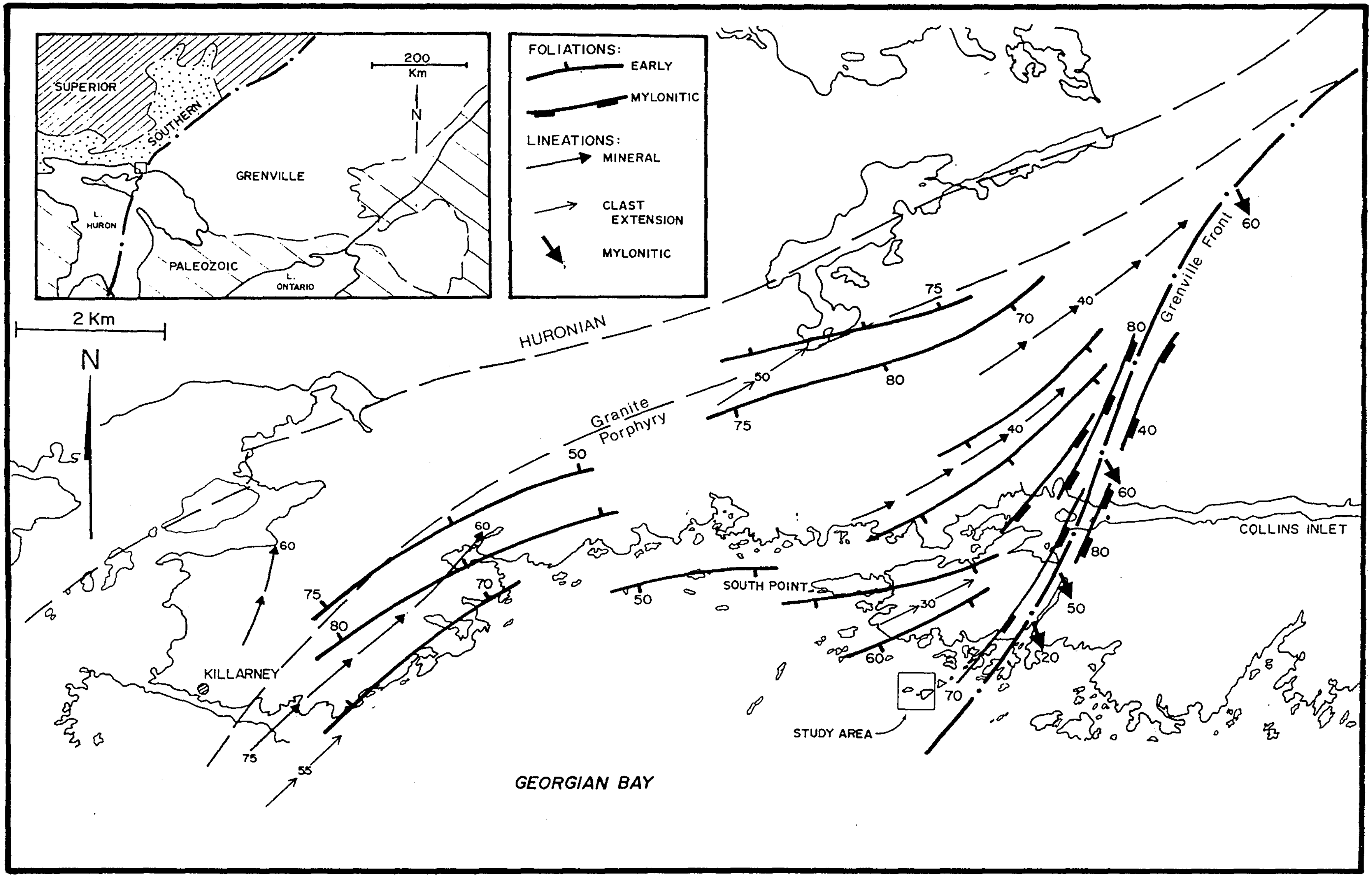
Mylonites. Some of these features are: an earlier formed foliation and lineation, en-echelon fractures post dating the foliation, and an array of small to medium scale brittle-ductile shear zones, also post dating the foliation (Clifford, in press; MacKinnon and Clifford, 1988).

## 1.3 Nature of Study

The main purpose of this study is to investigate small scale structures located in the area northwest of the mylonites. Specifically a set of small brittle-ductile shear zones, located very close to the mylonite zone, have been examined both in outcrop and microscopically.

The ideal would be to determine the exact relationship, if any, between the small shear zones and mylonites. If this cannot be determined, however, a relative age or possible upper and lower age limits for the shear zones are desired. Either would further clarify the structural relationships in the vicinity of the Grenville Front Mylonite Zone.

**Figure 1.** Geological and structural map of the Killarney Area. Regional foliation and lineation data after Clifford (in press); lithological boundaries are from Davidson (1986) and Clifford (in press).


SUPERIOR
SOUTHERN
GRENVILLE
L. HURON
PALEOZOIC
L. ONTARIO
200 Km
N
FOLIATIONS:
EARLY
MYLONITIC
LINEATIONS:
MINERAL
CLAST EXTENSION
MYLONITIC
2 Km
N
HURONIAN
Granite Porphyry
Grenville Front
COLLINS INLET
SOUTH POINT
KILLARNEY
STUDY AREA
GEORGIAN BAY
75
70
80
50
40
60
55
30
20

# CHAPTER 2. HOST ROCK DESCRIPTION

## 2.1 Field Observations

The host rock is a porphyry, possibly of volcanic origin, containing rock fragments up to 15 cm in length. These fragments, however, are sparsely distributed throughout the host rock and do not define any visible compositional layering (plate 1). Evidence of a possible volcaniclastic origin is best observed at the microscopic level where glassy rounded fragments are common.

A foliation, defined by the preferred orientation and alignment of muscovite gives the host rock an obvious fabric. The oriented muscovite separates elongate zones of feldspar and quartz crudely resembling 'ribbons'. The long axes of elongate clasts are aligned roughly parallel to the foliation, and commonly define a lineation. Mineral lineations defined by quartz aggregates are also common. The foliation is not continuous across the fine grained clasts, as these are lacking in muscovite. The local foliation, trending 045°, dips steeply at approximately 60-70° (figure 1). This orientation varies as the foliation is diverted by small shears crosscutting the island. This foliation/shear zone relationship will be discussed further on in the text.

The overall appearance of the host rock is that of a well foliated granite or granodiorite. This, however, holds true only in places where fragments are absent.

## 2.2 Mineralogy/Microstructures

The host rock porphyry is composed predominantly of four major mineral species. It consists of approximately 30% quartz, 30% potassium feldspar, 15% oligoclase and 10% muscovite. Minor amounts of garnet, biotite, chlorite, sphene and epidote are also present.

A typical porphyritic texture is observed in thin section. (plates 2 and 3) Quartz grains, approximately 0.05mm in diameter, show hypidiomorphic texture and exhibit undulose extinction, subgrains and recrystallized grains. Larger, anhedral feldspars (≈ 0.60mm average diameter) and grain aggregates (up to 4.0mm diameter) appear to be rounded, exhibiting undulose extinction, but lacking in subgrain formation. The textural evidence displayed by quartz and feldspar grains along with muscovite alignment suggests that the host rock has undergone deformation.

The microstructures exhibited in the ductile deformation of quartz are indicative of dislocation substructures which accommodate imposed stress. Lattice dislocations play an important role in this process. Undulatory extinction, and subgrain formation are the first observable effects of strain on quartz grains. A further increase in stress results in recrystallized grains (White, 1976) and then the development of quartz 'ribbon' aggregates (Simpson, 1983; Bouchez, 1977) which occur at the onset of greenschist facies temperatures. Quartz textures earlier described, indicate that subgrain formation followed by the recrystallization of older grain cores has occurred. Ribbon structures, however, if present at all, are not well defined.

At similar temperatures, feldspars deform differently than quartz. They behave in a brittle manner, fracturing at the onset of stress. At higher temperatures, feldspars show ductile deformation features such as undulatory extinction. Eventually, subgrains will begin to form and deformation will proceed in a similar manner to quartz (White, 1975). This deformation of feldspar by truly ductile processes occurs into amphibolite facies (White, 1975; Hamner, 1982; White and Mawer, 1985). At this point recrystallization of subgrains accompanied by chemical changes will be the main deformation mechanisms.

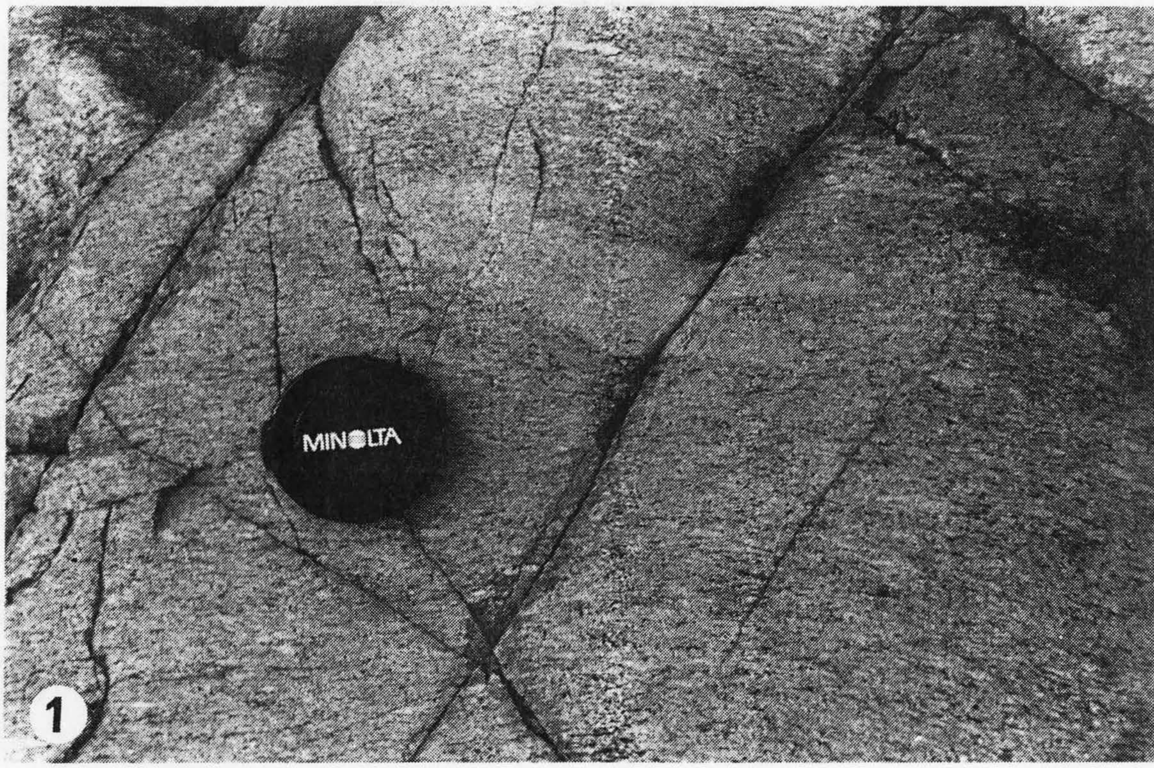
The feldspar textures exhibited in the host rock do not indicate truly ductile deformation. Slight undulatory extinction and rounding of grains are common, however, sub grains and recrystallized grains are absent.

The proposed metamorphic grade for the prior deformation is in agreement with Davidson (1986). Greenschist facies conditions have not been exceeded. Evidence is provided both by textural features and metamorphic minerals present.

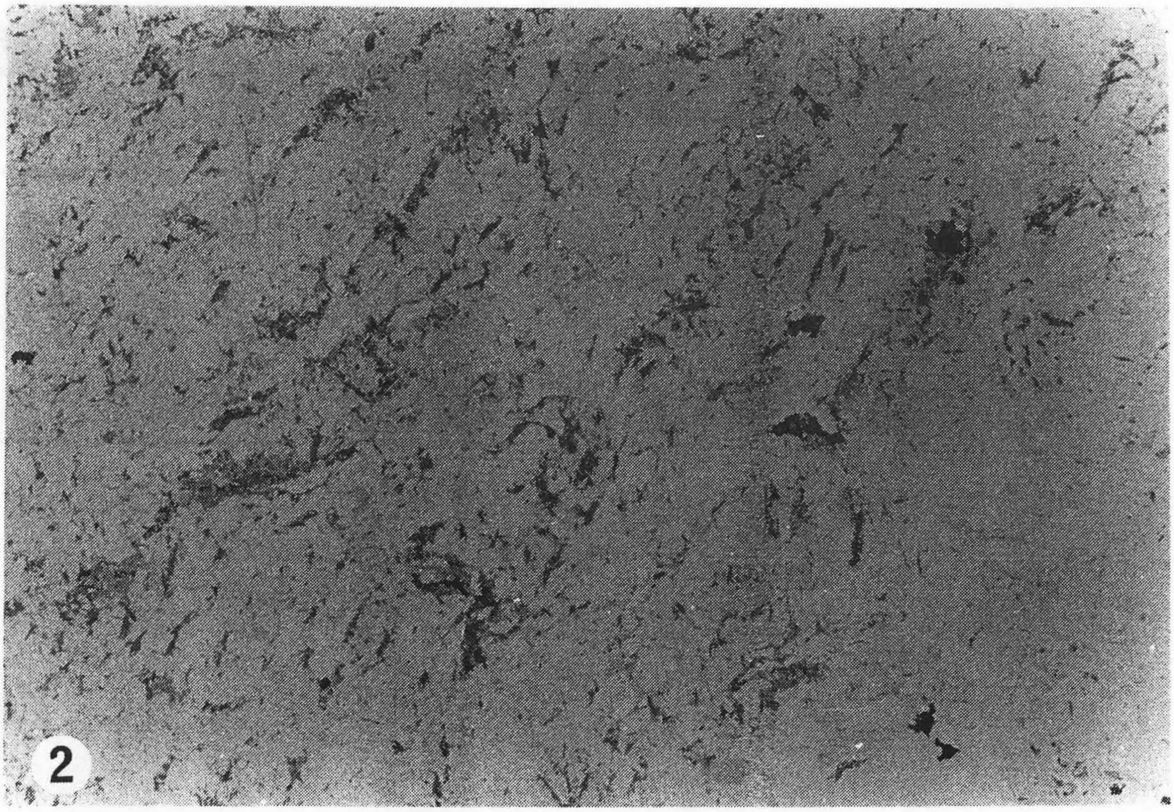
Textural evidence suggests that feldspars have undergone brittle fracturing and abrading with the development of slight undulatory extinction. Quartz grains demonstrate undulatory extinction, subgrain and recrystallized grain formation. This puts the limit of metamorphism below amphibolite grade (based on feldspar textures), and above lower greenscist facies (based on quartz evidence).

Metamorphic index mineral assemblages show similar results. Index minerals such as muscovite, biotite, garnet, chlorite, and epidote are present. These assemblages give further evidence that greenschist facies conditions have not been exceeded during the deformation that produced the host rock fabric.

Plate 1. Host rock porphyry showing elongate fragments aligned parallel to regional foliation.


MINOLTA
1

**Plates 2 & 3.** Host rock showing typical porphyritic texture. Quartz grains exhibit undulose extinction, subgrain formation and recrystallized grains. Feldspars are rounded and show slight undulatory extinction. Thin sections were photographed in both plane polars (plate 2) and crossed polars (plate 3). The width of the photographs is approximately 2cm.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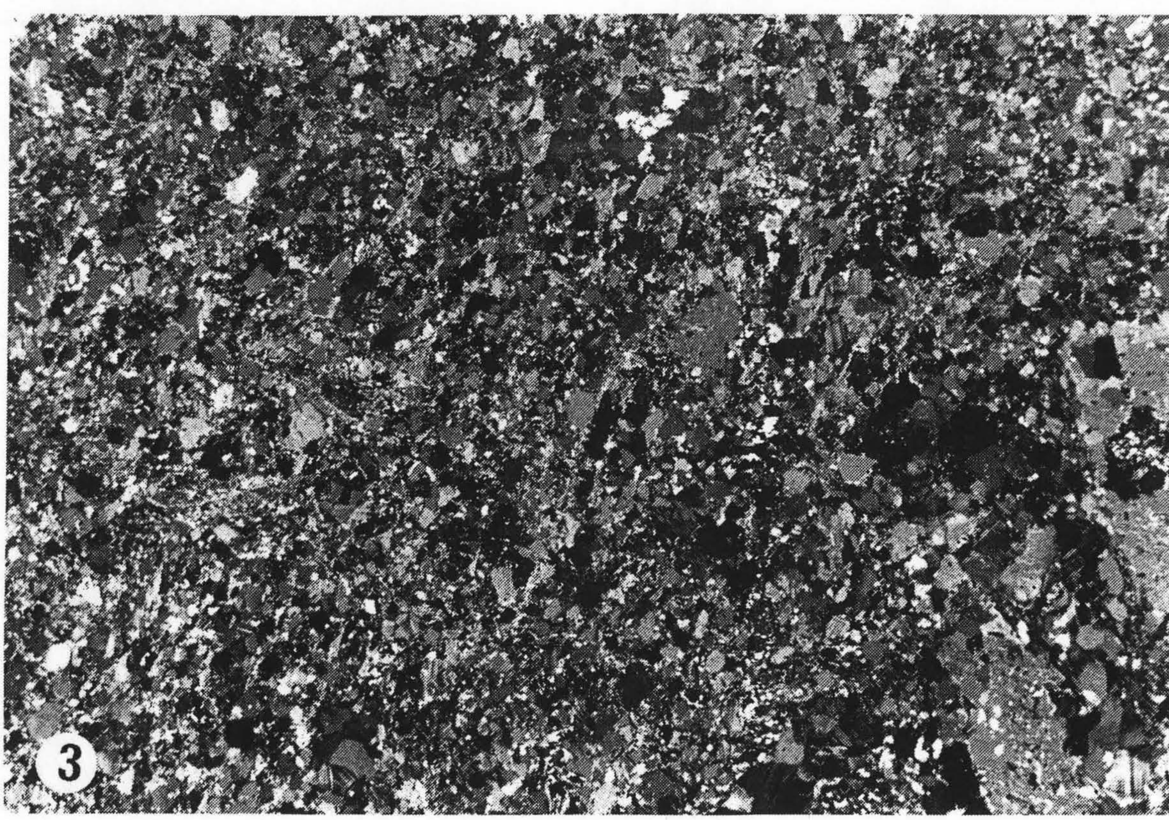

3

# CHAPTER 3. SHEAR ZONE DESCRIPTION

## 3.1 Field Observations

### 3.1.1 General Description

The shear zones found within the study area are parallel, planar sided features which generally trend at 225° (figure 2). They range in size from small incipient (≈ 1 m) shears, a few centimeters wide and approximately 1.0m in length, to the largest shear zone which is approximately 0.5m wide and crosses the entire width of the smaller island.

The terminations of these shear zones exhibit both of Simpson's (1983) brittle to brittle-ductile termination types. Type one shows a symmetrical decrease in the area of finite strain and in strain intensity as the shear zone terminates. Type two, in contrast, shows a symmetrical increase in the strained area as stress intensity decreases.

The only visible relationship between the two termination types and the range of shear zones represented is possibly one based on shear zone size. Field relations suggest that smaller shears, those on the order of 1.0m in length are of type 1, while larger shears, including the largest observed, are predominantly of type two. Unfortunately, the termination of the largest shear zone is submerged and cannot be inspected. Splays of the type 2 kind, however, are observed along its entire length as small shears splay off the main zone. This perhaps implies that the termination of this particularly large shear zone occupies a very large area (plate 4). Any relationship which may be predicted between termination type and size of the shear may be due to the existence of two sets of shear zones separated in time. No direct evidence for this is seen.

The shear zones typically cut the host rock foliation obliquely (plate 5). They generally trend at ≈ 225° and dip steeply at 70-80° to the northwest (figure 3). Measured foliations on the islands are variable due to the crosscutting and distortion by shear zones. The general trend is 095° with variable dips between 30° and 85° south (figure 4).

### 3.1.2 Nature of Shearing

The nature of the shearing is consistent with what Ramsay (1979) refers to as brittle-ductile. Varying degrees of both end members are found. In some cases (plate 6) a small pegmatite dyke may be deflected towards the shear zone boundary indicating ductile deformation. It is then offset by a brittle microfault along a clear discontinuity. In another case (plate 7), a very large pegmatite dyke is dragged into and across a shear zone boundary largely reflecting ductile deformation. This case, however, is only accompanied by minor microfaulting.

It is unlikely that brittle and ductile deformation will occur synchronously. Ductile processes will occur at lower crustal levels, while brittle deformation will continue to accommodate strain at a higher level (Ramsay, 1979).

The largest shear zone exhibits a banding parallel to shear zone boundaries (commonly referred to as fluxion structure in mylonitic rocks). These bands vary in colour and range from 1 mm to several cm in width. They commonly contain varying degrees of a fine grained matrix supporting rounded feldspar porphyroclasts.

Lighter coloured zones exhibit a protomylonitic texture, while darker zones are more mylonitic to ultramylonitic. The feldspar porphyroclasts which are typically rounded and lack asymmetric tails, appear to diminish in size from lighter to darker bands. The matrix is generally fine grained, becoming more abundant and finer grained from lighter to darker bands. Generally, ultramylonite bands are found closer to the centre of the shear zone where strain is higher. Protomylonites are commonly seen near shear zone boundaries where the lowest strain is present. Gradations from light coloured coarse bands to darker glassy bands are found across the entire width of the shear zone. In shear zone splays terminating in type two fashion (Simpson, 1983), the divisions between bands becomes more diffuse until eventually all strain is accommodated in the adjacent host rock. Type one terminations are found only in smaller shears which do not display banding. The type one termination style of banding therefore, cannot be assessed.

**Figure 2.** Map of study area as outlined in figure 1. Refer to appendix at back of text for orientation data.

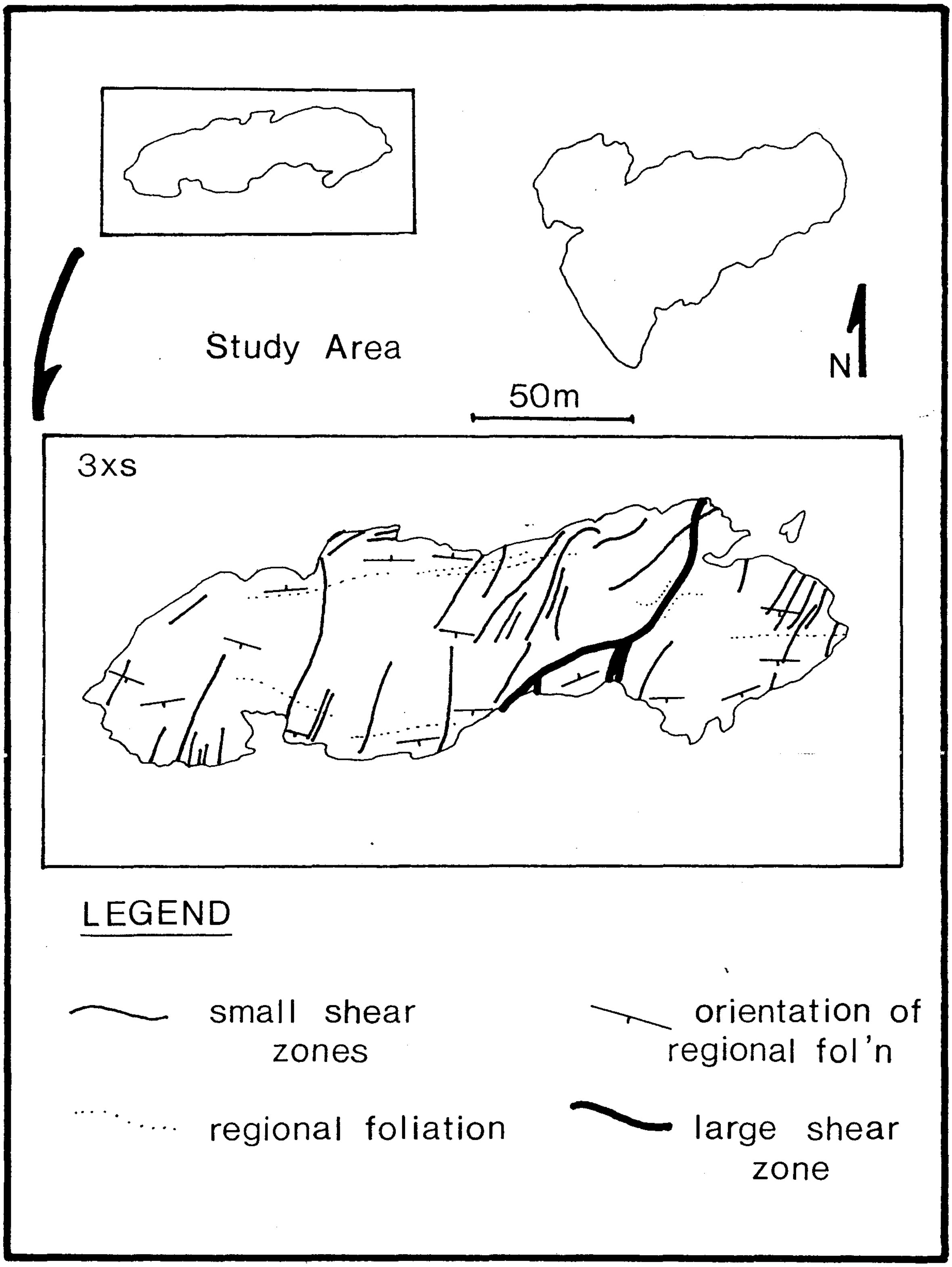
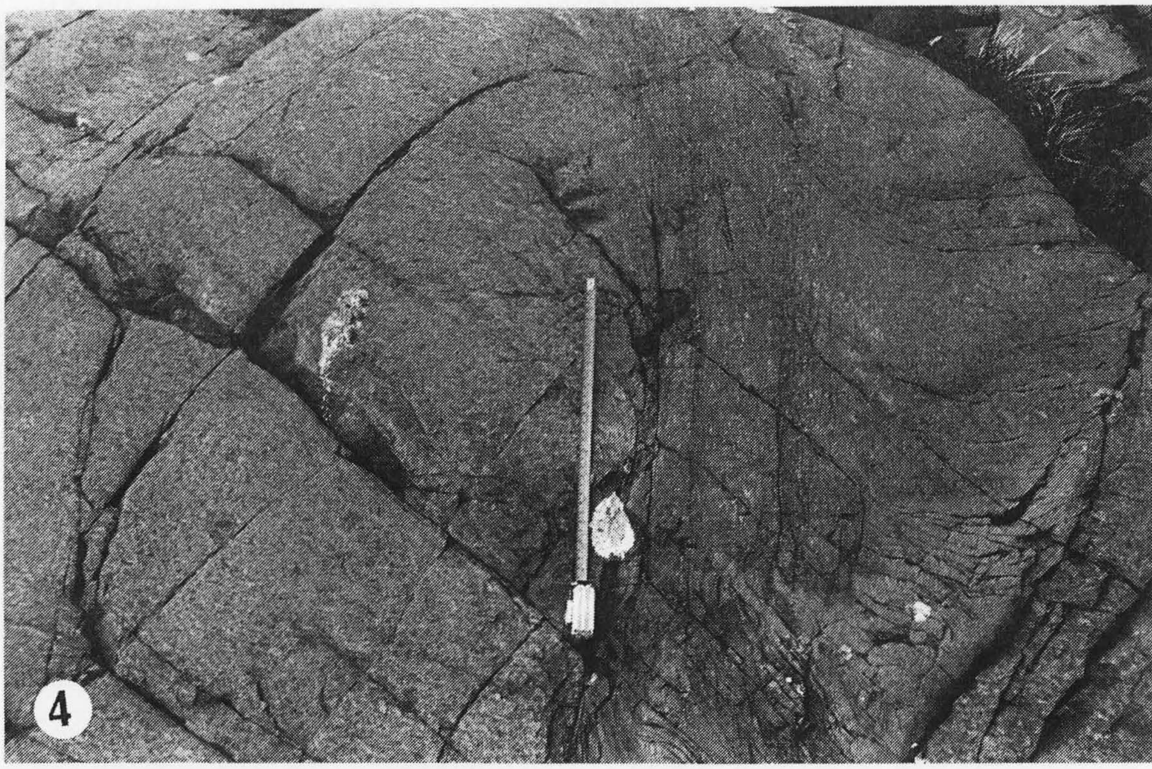
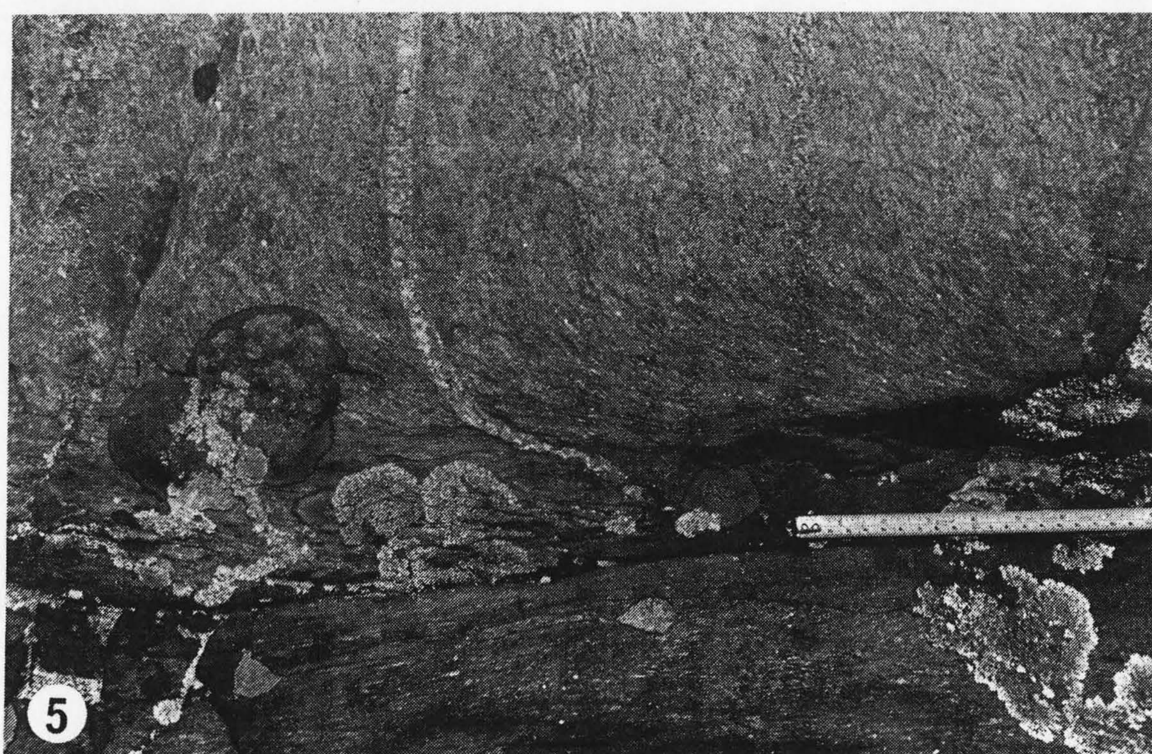

Study Area
N
50m
3xs
LEGEND
small shear zones
orientation of regional fol'n
regional foliation
large shear zone

**Plate 4.** Large shear zone showing minor shears splaying off the central zone.

**Plate 5.** Foliation obliquely cut by a small incipient shear zone.





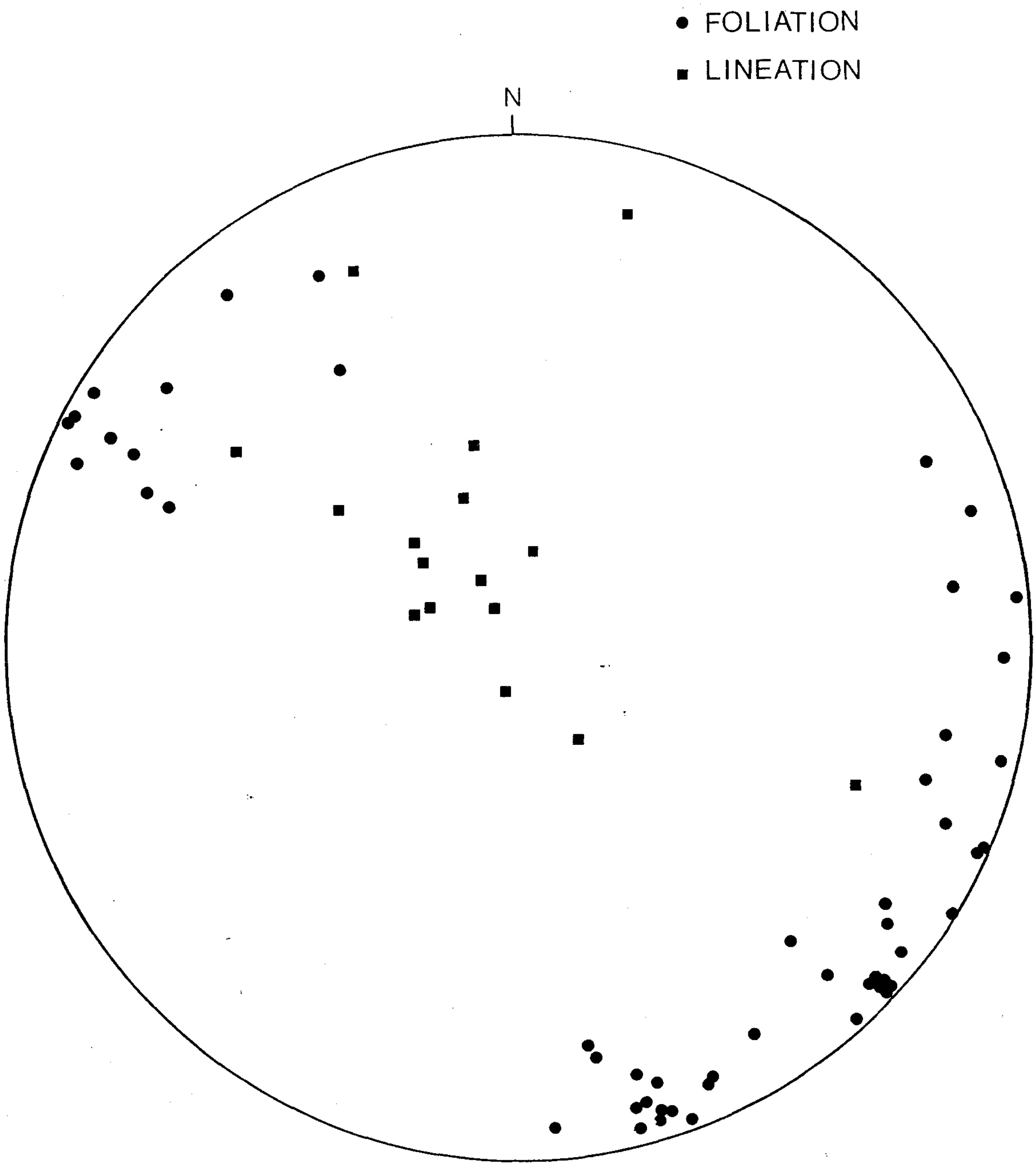



Figure 3. Stereoplot of shear zone foliation and lineation orientations.

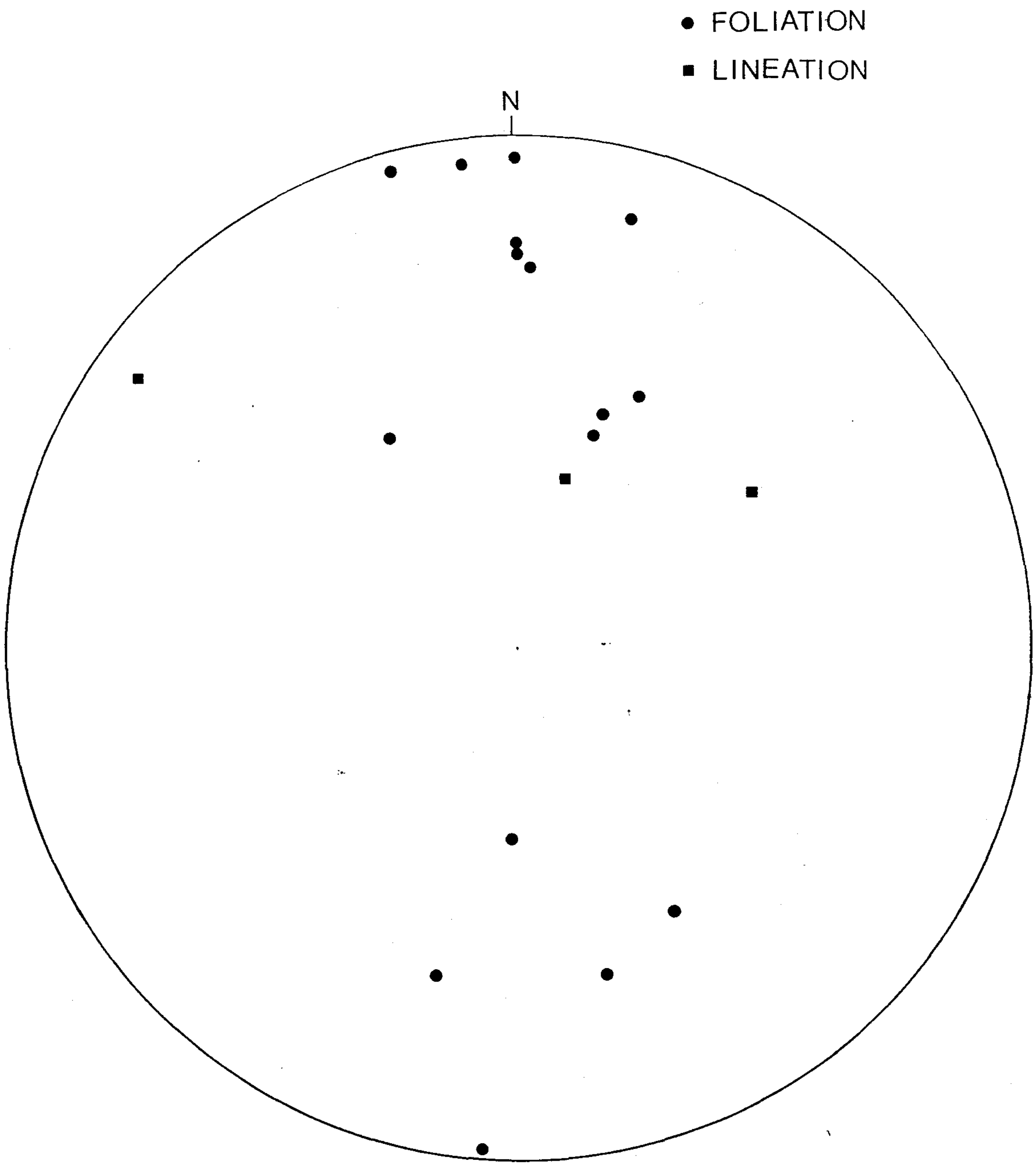
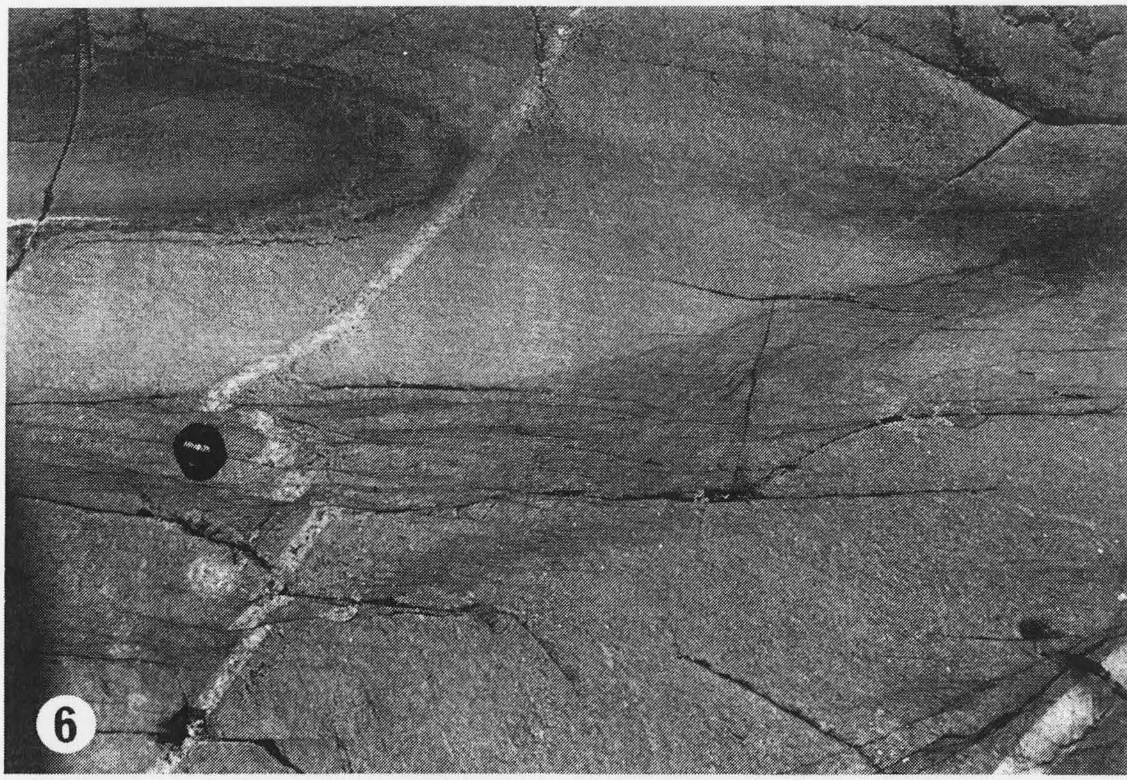



**Figure 4.** Stereoplot of local host rock foliation and lineation orientations.

Plate 6. Pegmatite dyke ductilely deflected towards the shear zone boundary and offset by minor microfaults within the shear zone.

Plate 7. Pegmatite dyke exhibiting ductile behaviour as it is deflected by a small shear zone.



## 3.2 Microscopic Aspects

In examining the microscopic features of a strained rock information about the mode of deformation, relative temperatures, strain ratios, and kinematics are sought. Particularly, one is interested in observing systematic changes in grain shapes, sizes and average long axis orientation. This study attempts to accomplish this by systematically observing microstructures over a large shear zone from near the shear zone boundary to the centre of the zone. Thus, a complete sample representing an increase in strain across the shear zone is examined.

### 3.2.1 Method of Study

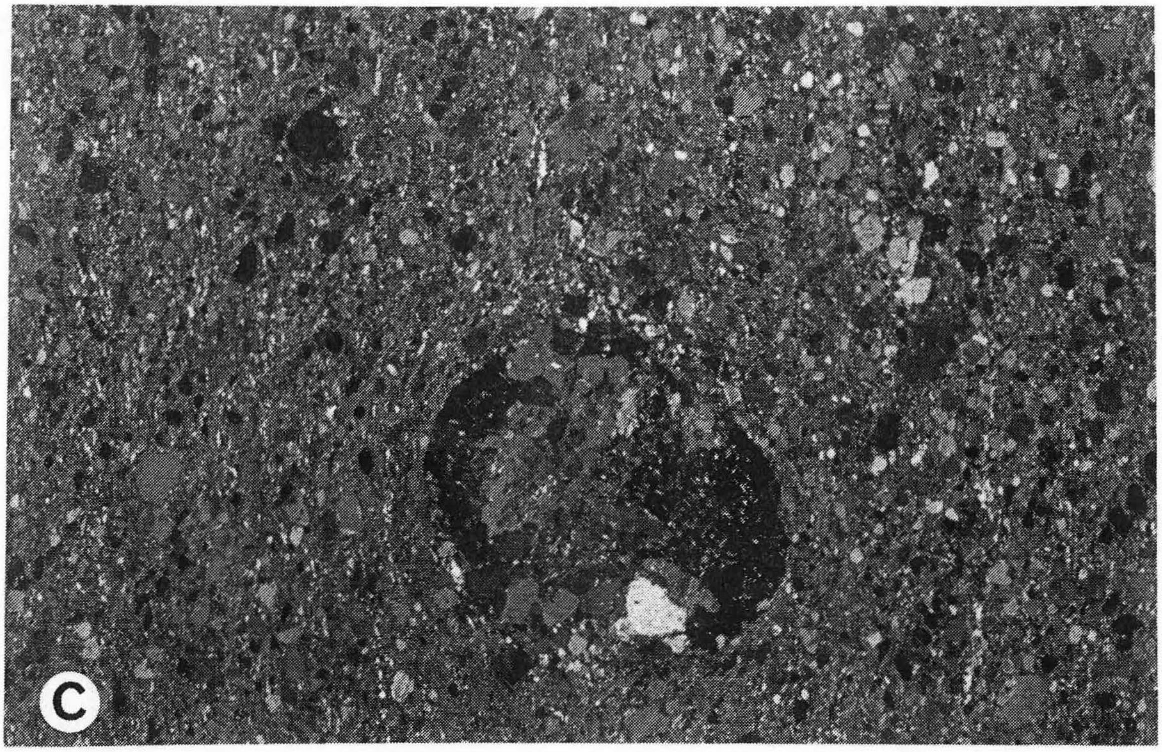
Two sets of thin sections were cut in perpendicular planes. One set is perpendicular to the foliation and parallel to the lineation, the other perpendicular to both foliation and lineation. Photomicrographs of some representative thin sections are presented in plates 8 through 13. They demonstrate a relative increase in strain across the shear zone in the following order: C (plate 8), B (plate 9), and A (plate 10), where thin sections are lineation perpendicular. Plates 11 through 13 show a similar increase in strain: CX (plate 11), BX (plate 12), and AX (plate 13), where 'X' denotes sections cut lineation parallel.

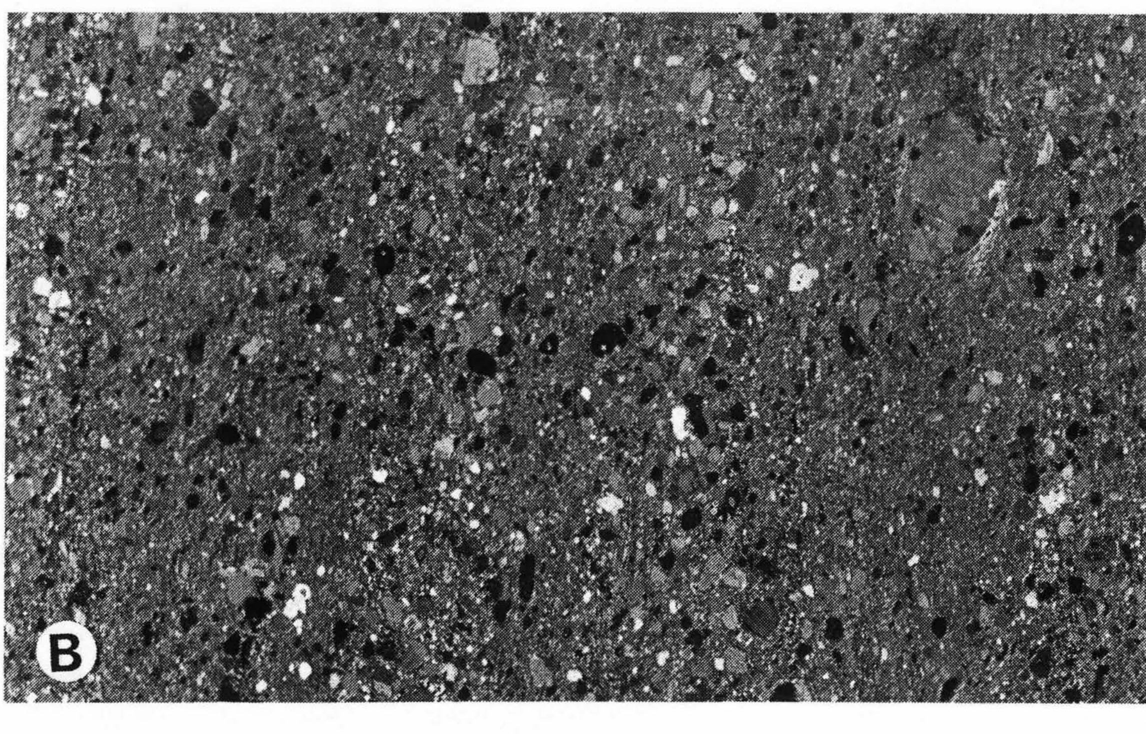
Measurements of grain length, width and long axis orientation to the foliation were taken. These were systematically recorded from the shear zone boundary into the central portion of the zone.

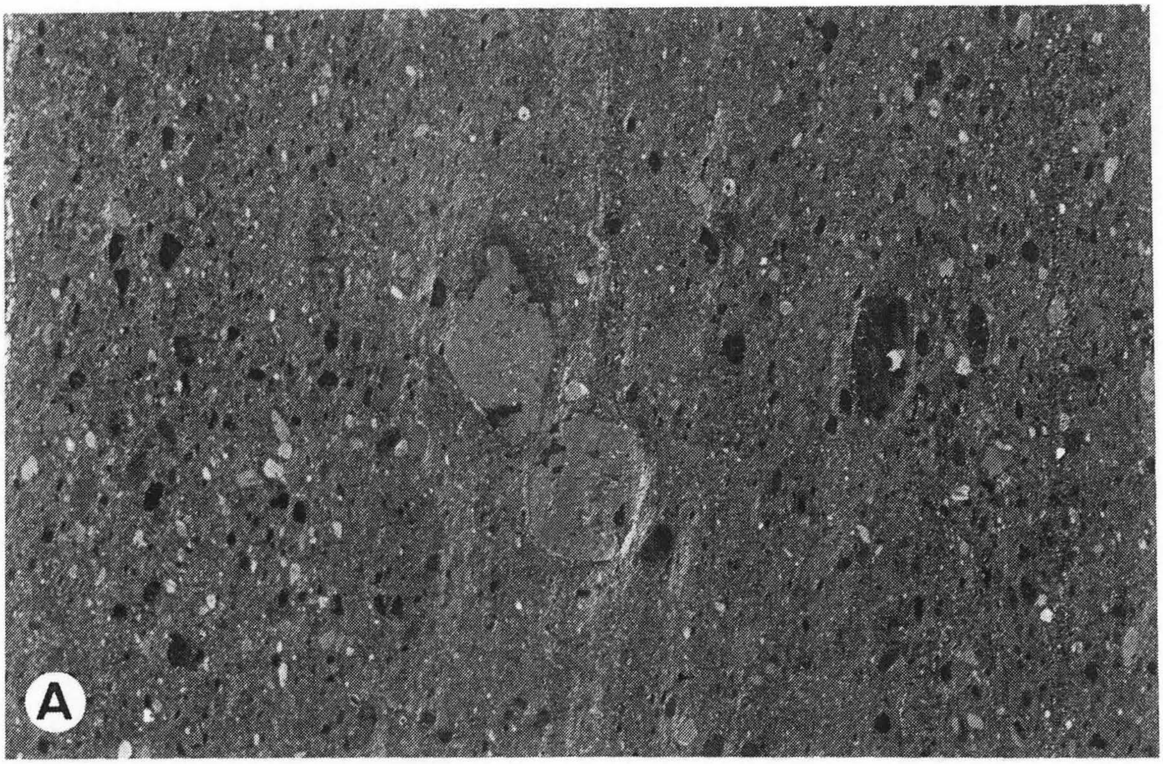
Due to the 'banded' nature of the sheared rock coarse and fine domains are separated. This separation has been made so that the mixing of data from extremely fine and coarse zones will not conceal important information about the distinctive mechanisms occurring in each case (if in fact these occur).

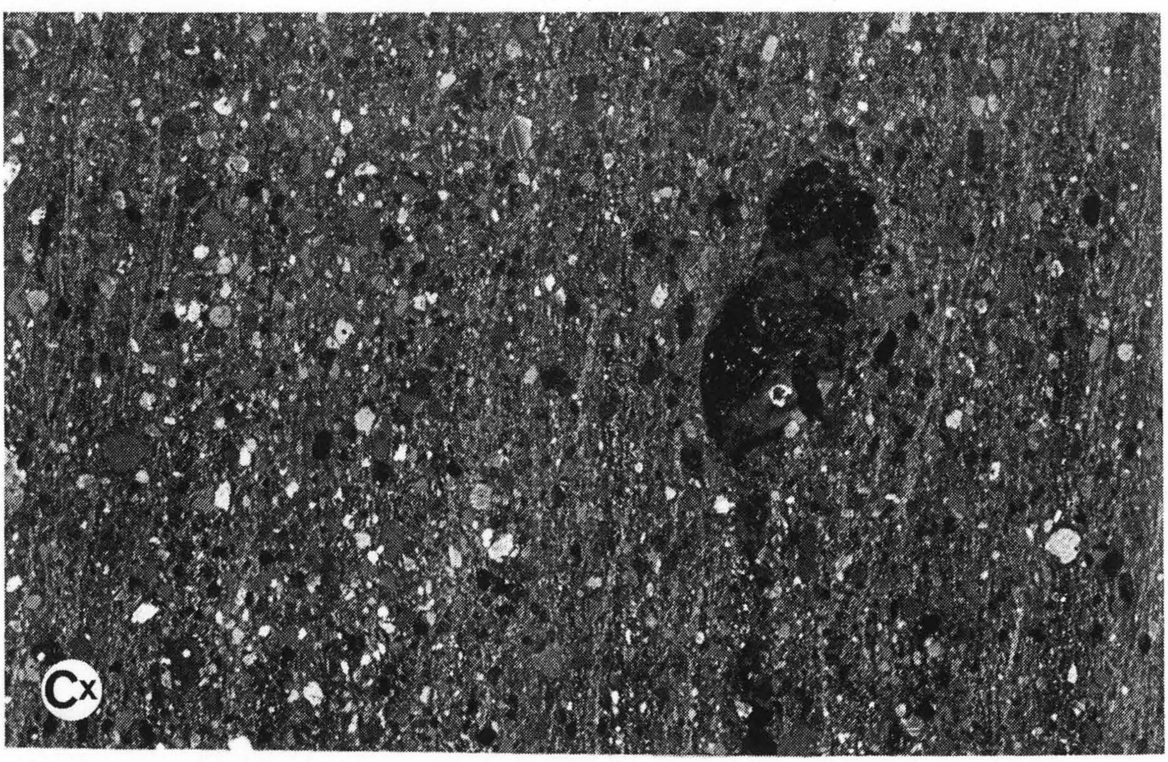
Rf/ø plots were constructed using the grain size and orientation data (Lisle, 1985). The plots to be discussed are selected from the least and most extremely strained areas. Thus, both coarse and fine data from thin sections AX, CX, A and C will be treated. These are labelled as AXF, AXC, CXF, CXC, AF, AC, CF and CC, where the last character in each title denotes the coarse (C) and fine (F) domains (figures 5 through 12).

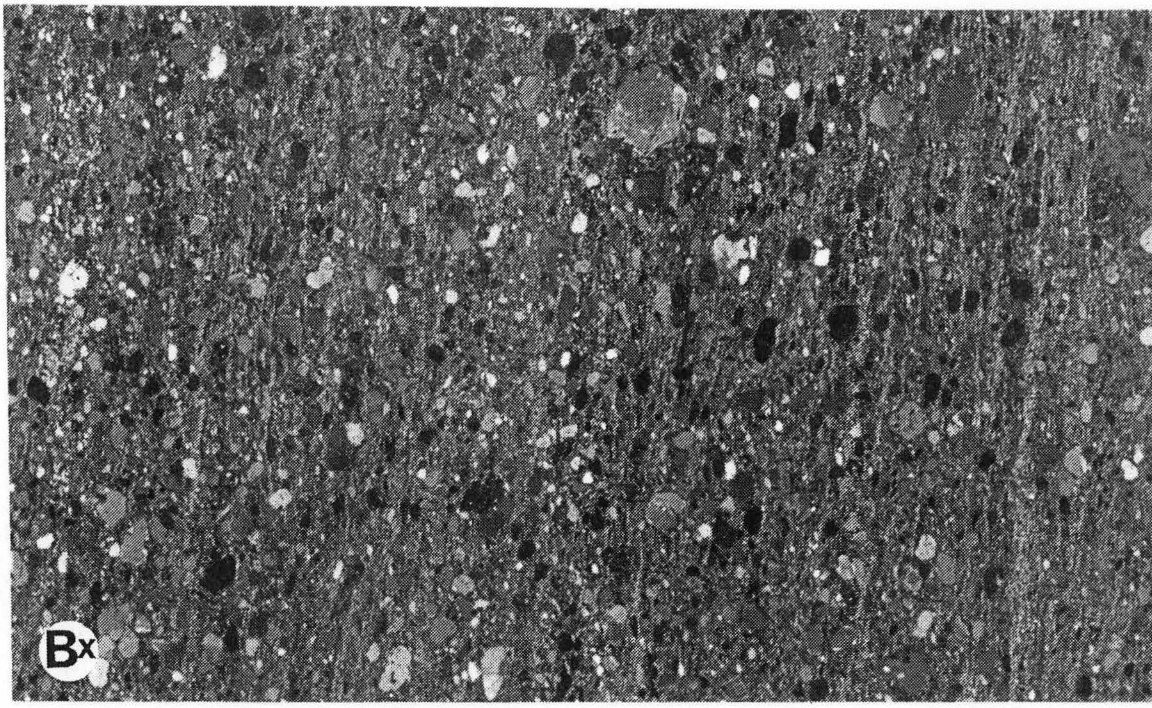
**Plates 8 to 13.** Photomicrographs demonstrate deformation textures across the largest shear zone. The first character in each title represents thin section position relative to the shear zone where 'C' is close to the shear zone boundary and 'A' is from the central portion. 'X' represents sections cut parallel to lineation. The width of the photographs is approximately 1cm.

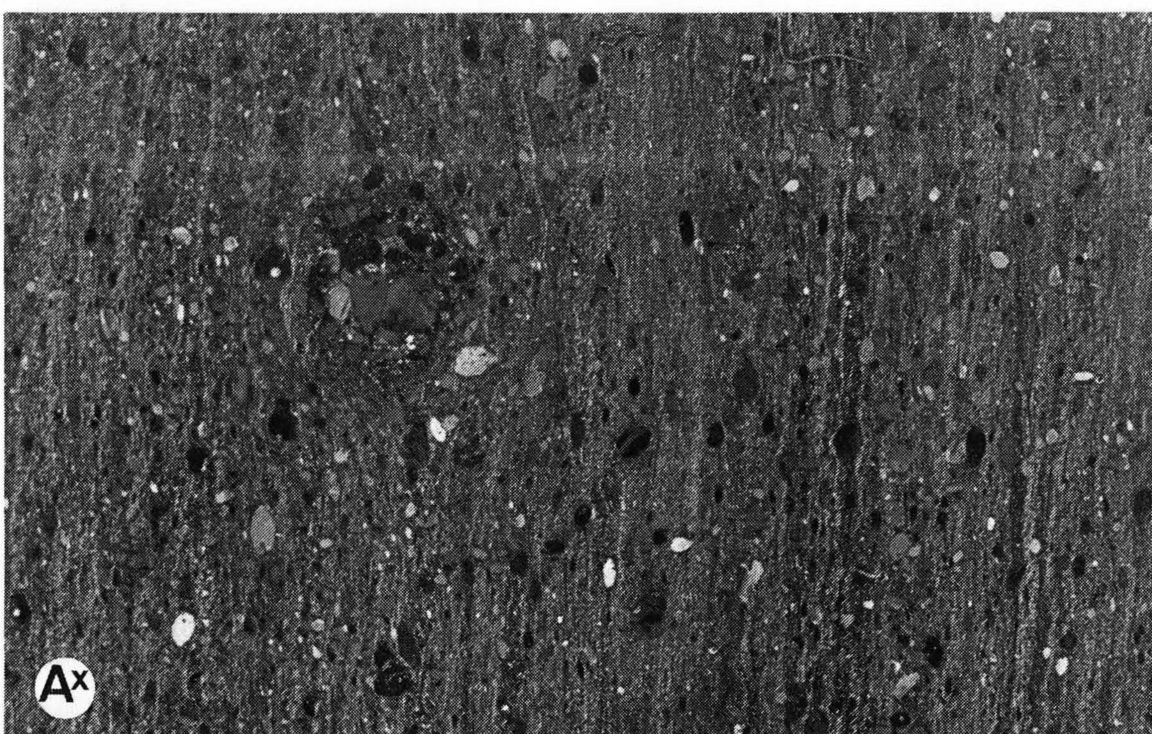

C


B


A


Cx


Bx


Ax

**Figures 5 to 12.** Rf/$\phi$ plots constructed using grain size and orientation data sampled across the largest shear zone. The plots represent the least and most extremely strained areas of the shear zone. The first character denotes high strain (A) or low strain (C) while the last character denotes a coarse (C) or fine (F) domain. 'X' denotes lineation parallel sections. All sections have been cut foliation perpendicular.

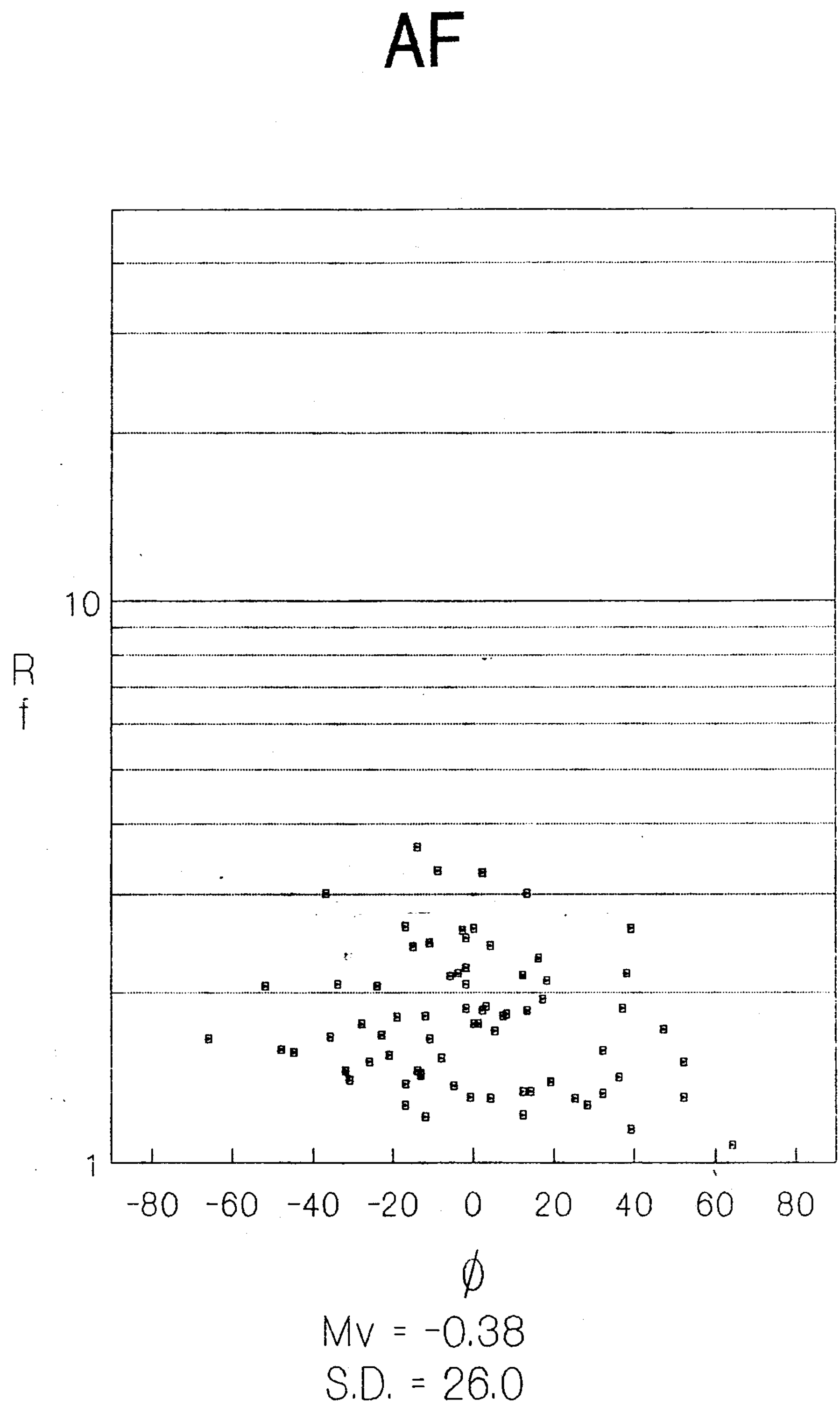



Figure 5. Rf/ø plot AF

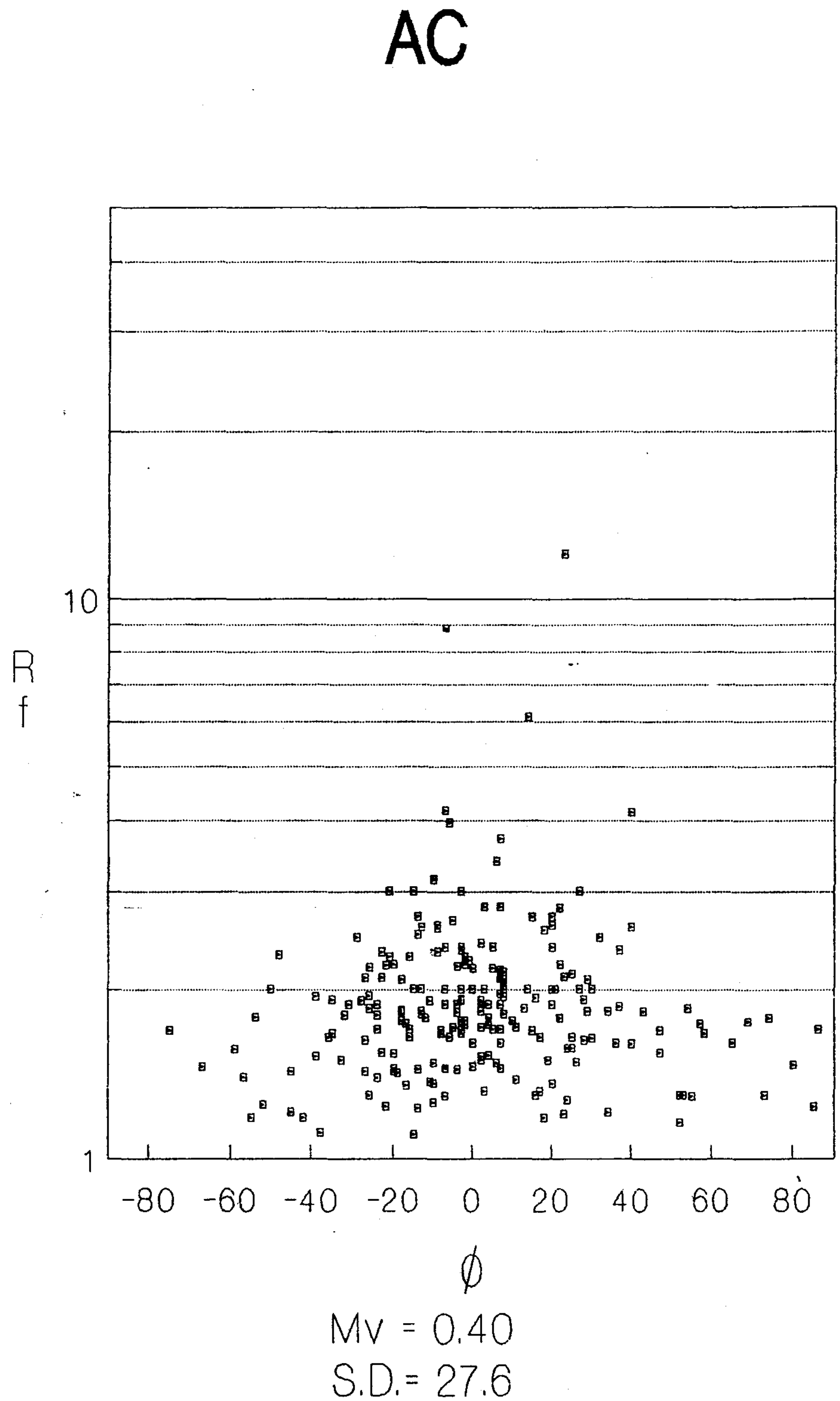



Figure 6. Rf/ϕ plot AC

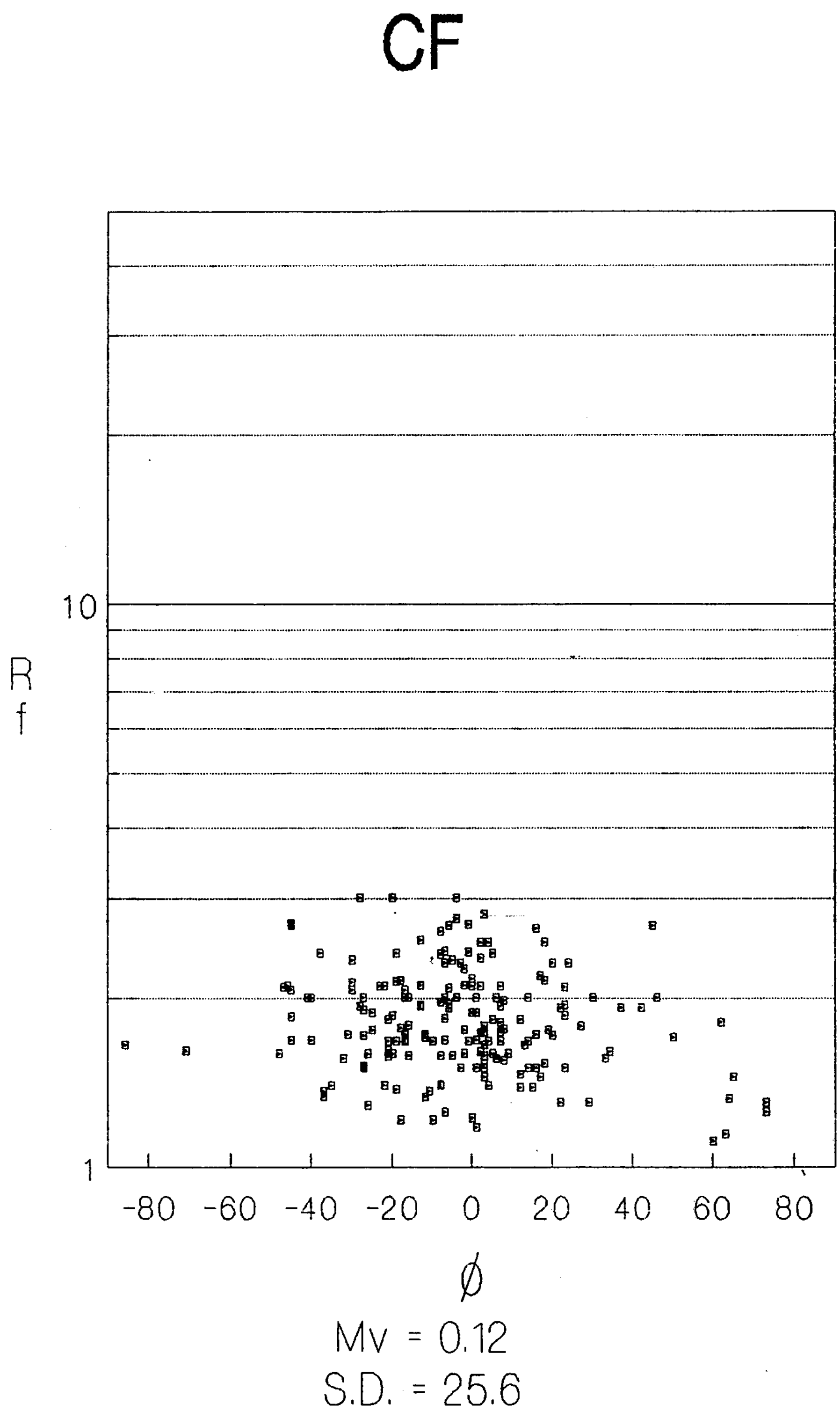



Figure 7. Rf/ø plot CF

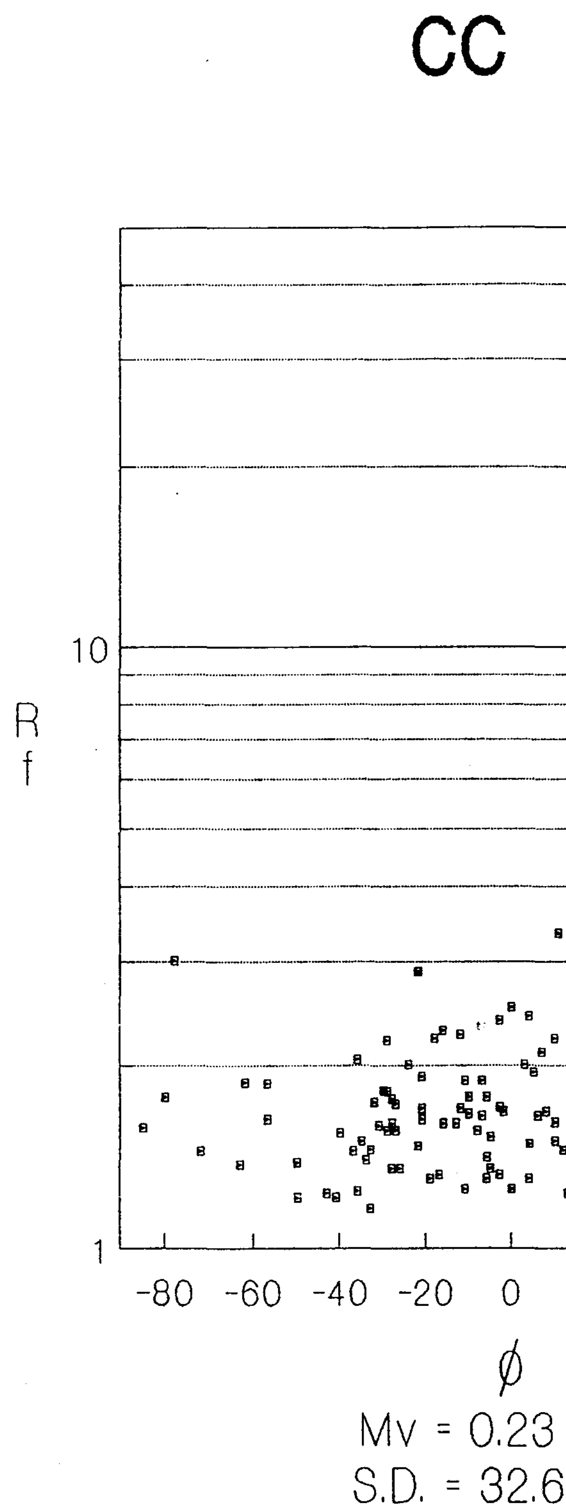



Figure 8. Rf/ϕ plot CC

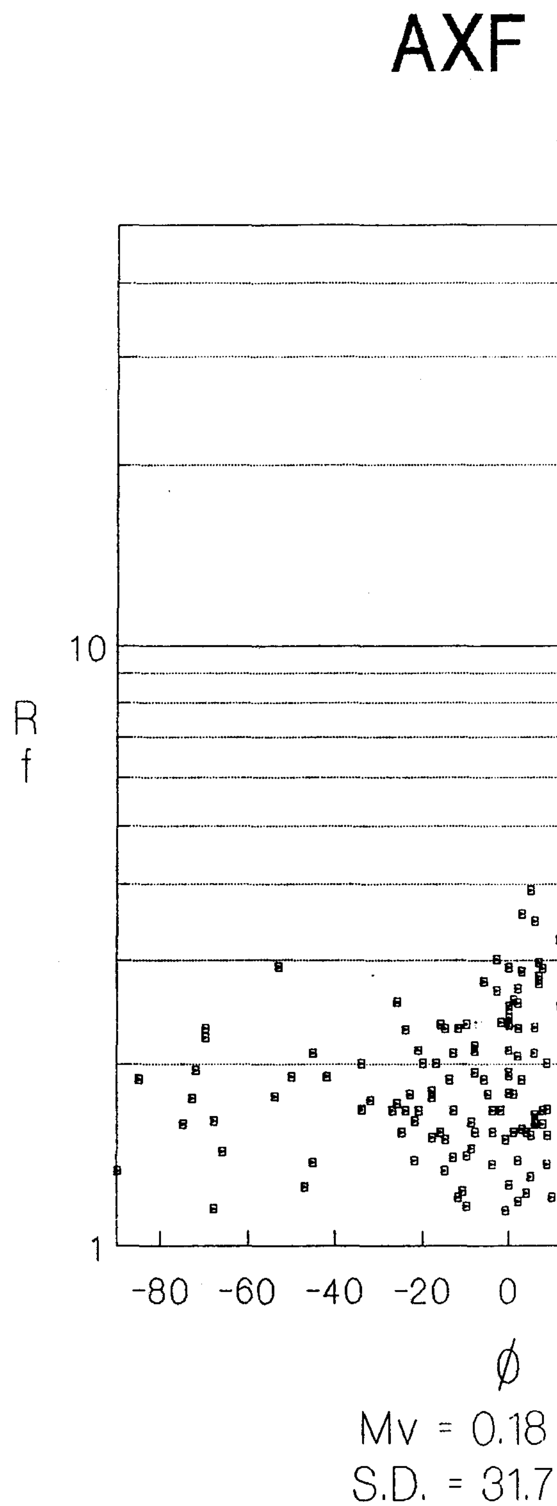



Figure 9. Rf/ø plot AXF

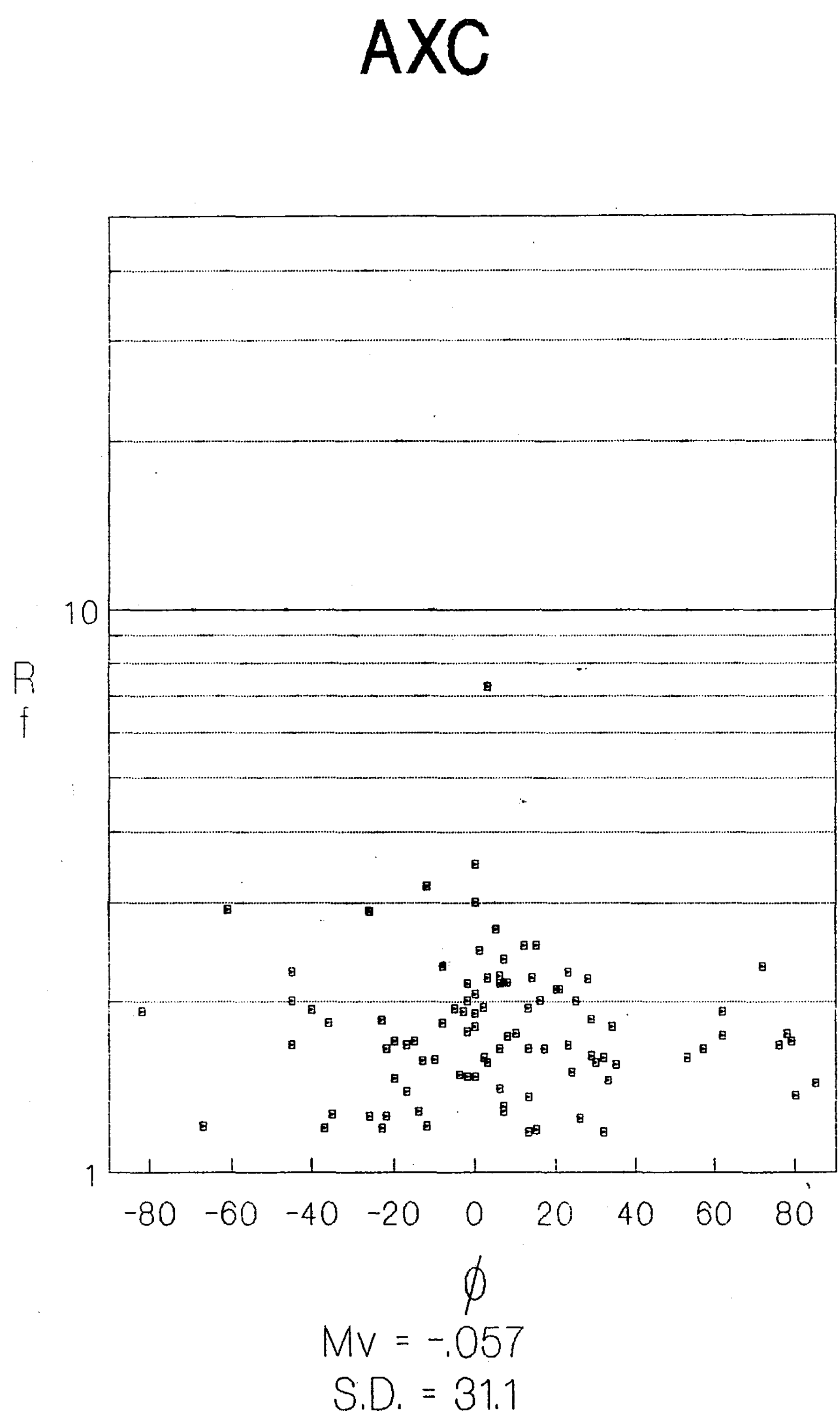



Figure 10. Rf/ϕ plot AXC

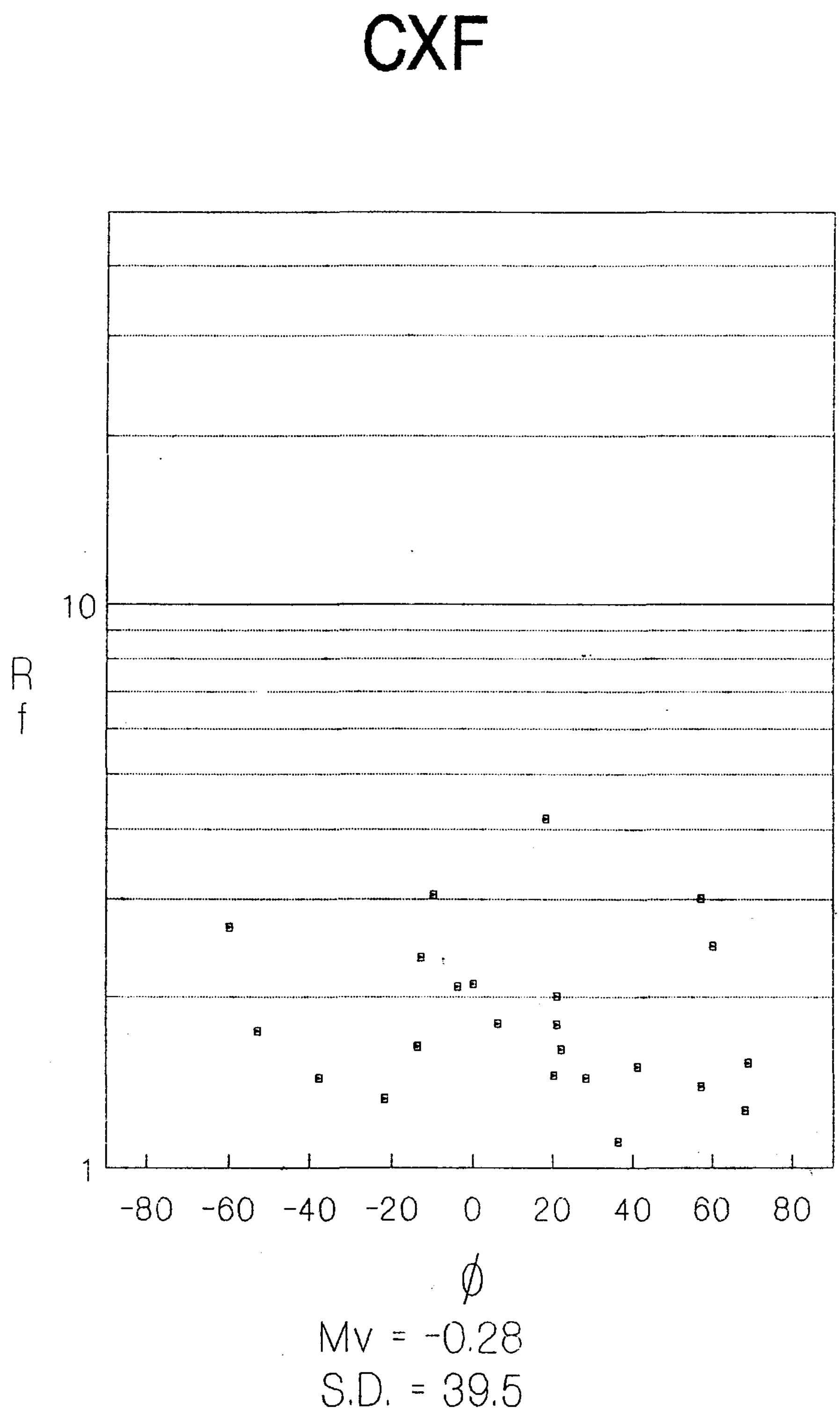
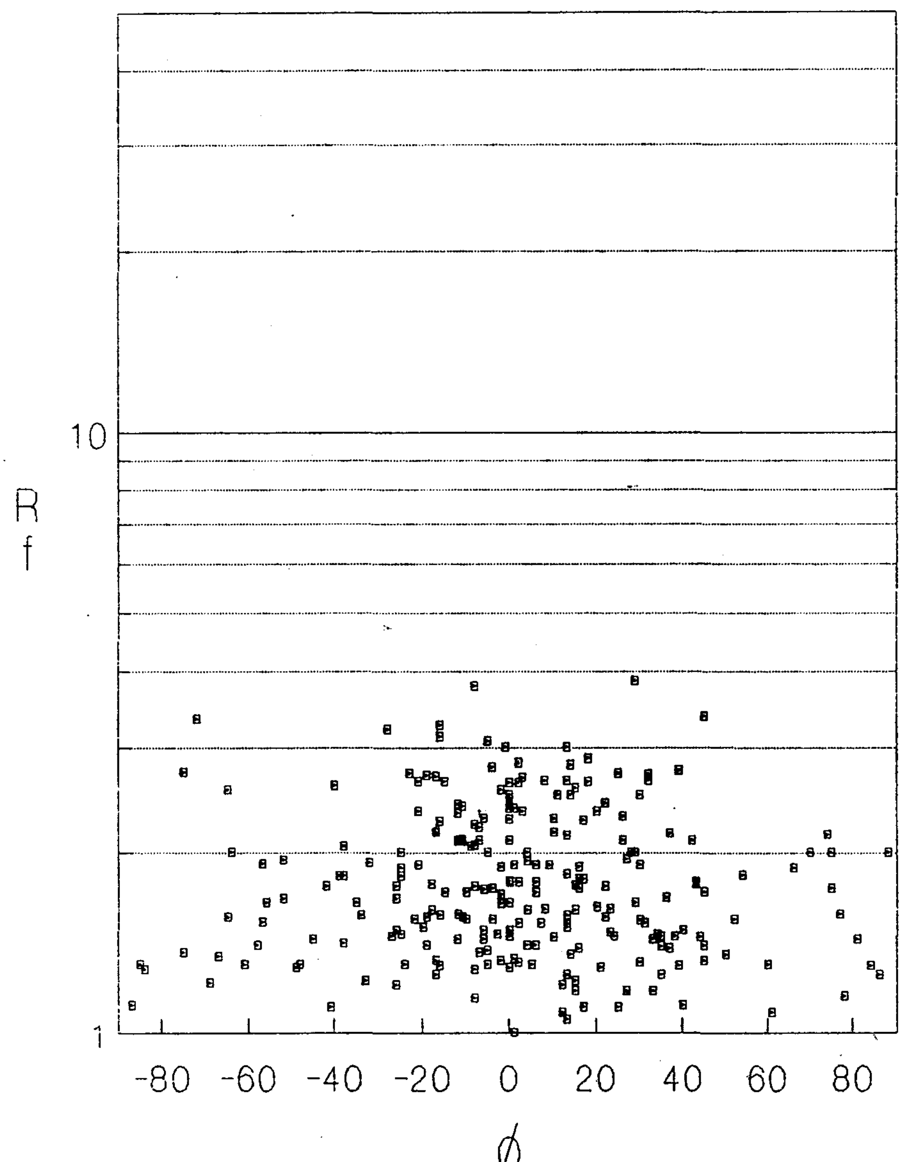



Figure 11. Rf/ϕ plot CXF

# CXC




Mv = 0.49

S.D. = 33.2

Figure 12. Rf/ϕ plot CXC

3.2.2 Microstructures

Before a discussion of the measured grain data commences, a brief microscopic description of the sheared rock is necessary.

The dominant identifiable grain species are rounded feldspar porphyroclasts. These are surrounded by a very fine grained powdery matrix comprised dominantly of quartz and finer equivalents of host rock minerals. A visibly brittle deformation has affected the sheared rock; the porphyroclasts are all rounded and approximate the same shape while the matrix is highly glassy and lacking in typical ribbon shapes.

Due to the fine grained nature of the sheared rock, the determination of length to width ratios was restricted to feldspars as these were the only grains large enough to be measured. Feldspar porphyroclast data plotted on Rf/ø diagrams reveal two things. Firstly, the Rf/ø values for all sections including both coarse and fine domains are very close. These values are summarized in table 1 where the maximum and minimum Rf values are 1.8 and 1.66 respectively.

Secondly, the scatter of long axis orientations to the foliation (ø) about the vector mean is relatively constant across the shear zone. The standard deviations for all data sets are on the order of 30° , therefore no significant changes are occurring as strain apparently increases.

During the ductile deformation of rock, the concentration of long axis orientations tends to increase with increasing strain. Similarly, the value of Rf will increase towards the centre of the shear zone. The conclusion therefore, is that feldspar deformation has occurred by brittle means, fracturing (Andrews, 1984) until attaining an 'equillibrium size' (Boullier, 1980). Aggregates seem to behave in a similar way. An important question follows from this; what processes are a predominantly ductile mineral species such as quartz undergoing?

The quartz is present in the matrix as a fine powder. Its structureless glassy appearance indicates that a cataclastic process has crushed it to its present state. According to Higgins' (1977) classification of cataclastic rocks, however, a true cataclasite exhibits a lack of fluxion structure, indicating the absence of 'rolling' and the predominance of 'crushing' as the main deformation mechanism. Fluxion structure, earlier described as banding in the sheared rock is clearly illustrated in these rocks as the narrow dark and light bands. Typically, as in a protomylonite to ultramylonite, fluxion structure is defined by ductile quartz ribbons and smeared out porphyroclasts. The sheared rock here, in contrast, clearly indicates a brittle deformation mechanism. Quartz is not in ribbons nor are feldspar porphyroclasts smeared out.

How then, can fluxion structure develop in a cataclasite?

An answer to this question is hidden in the ductile 'ghosts' revealed in plane polarized light. Plates 14 and 15 show high strain features (from thin section AX) in both plane and crossed polarized light. In plane polars, the appearance of an earlier ductile matrix becomes evident. Tails around porphyroclasts are subtle, as these have been destroyed by later cataclasis. Fine grained darker areas (probably crushed micas), however, preserve an earlier 'rolling' fluxion structure.

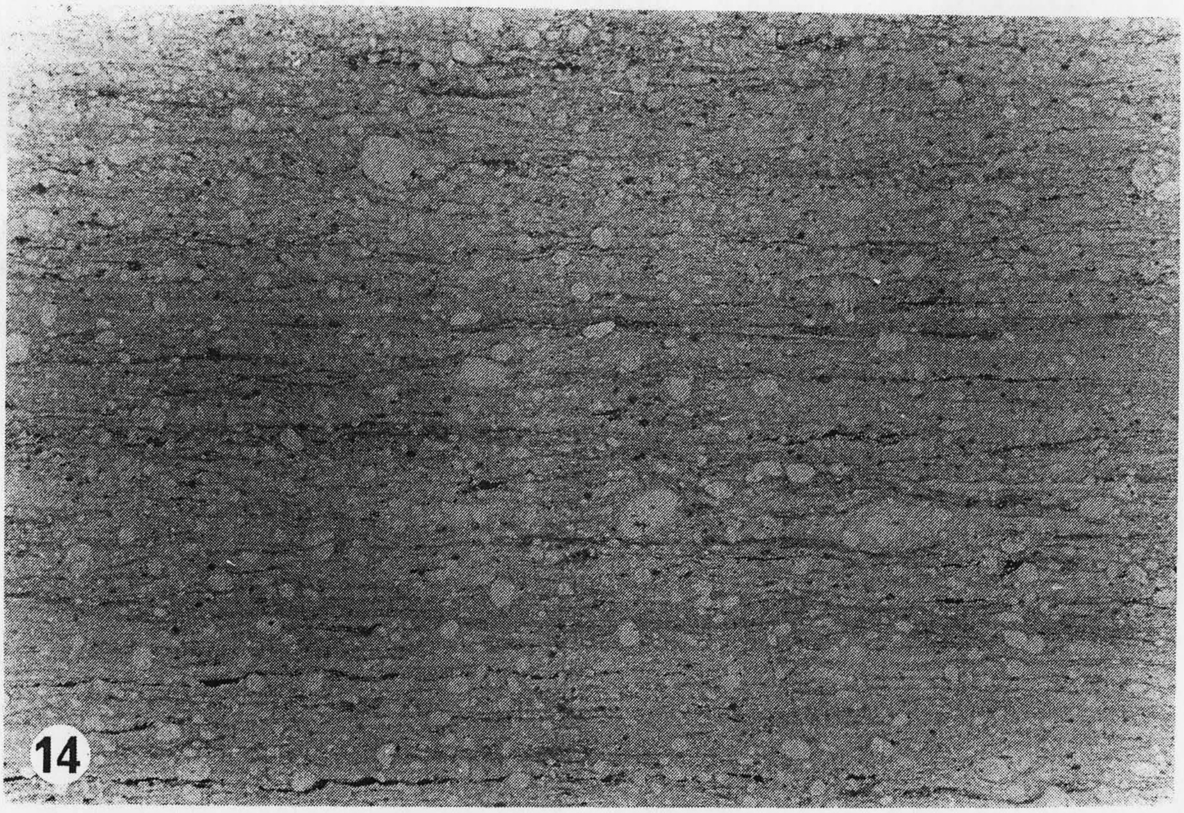
Most likely, during an initial ductile phase of deformation, quartz was deformed into 'ribbons', and porphyroclasts may have been smeared out. This has created the now 'relict' fluxion structure. All prior forms of ductile quartz and feldspar would have immediately responded to a cataclastic crushing. The newly imposed cataclasis has preserved this feature as the fine grained structureless bands. The resultant rock is dominantly a cataclasite, but one that shows evidence of a prior ductile phase. Thus the sheared rock in thin section reveals a brittle - ductile character just as field evidence suggests.

| DOMAIN | Mv | S.D. | Rf |
|---|---|---|---|
| AC | 0.40 | 27.6 | 1.80 |
| AF | -0.38 | 26.0 | 1.72 |
| AXF | 0.18 | 31.7 | 1.74 |
| AXC | -0.16 | 31.1 | 1.73 |
| CC | 0.23 | 32.6 | 1.66 |
| CF | 0.12 | 25.6 | 1.77 |
| CXF | -0.28 | 39.5 | 1.76 |
| CXC | 0.49 | 33.2 | 1.70 |

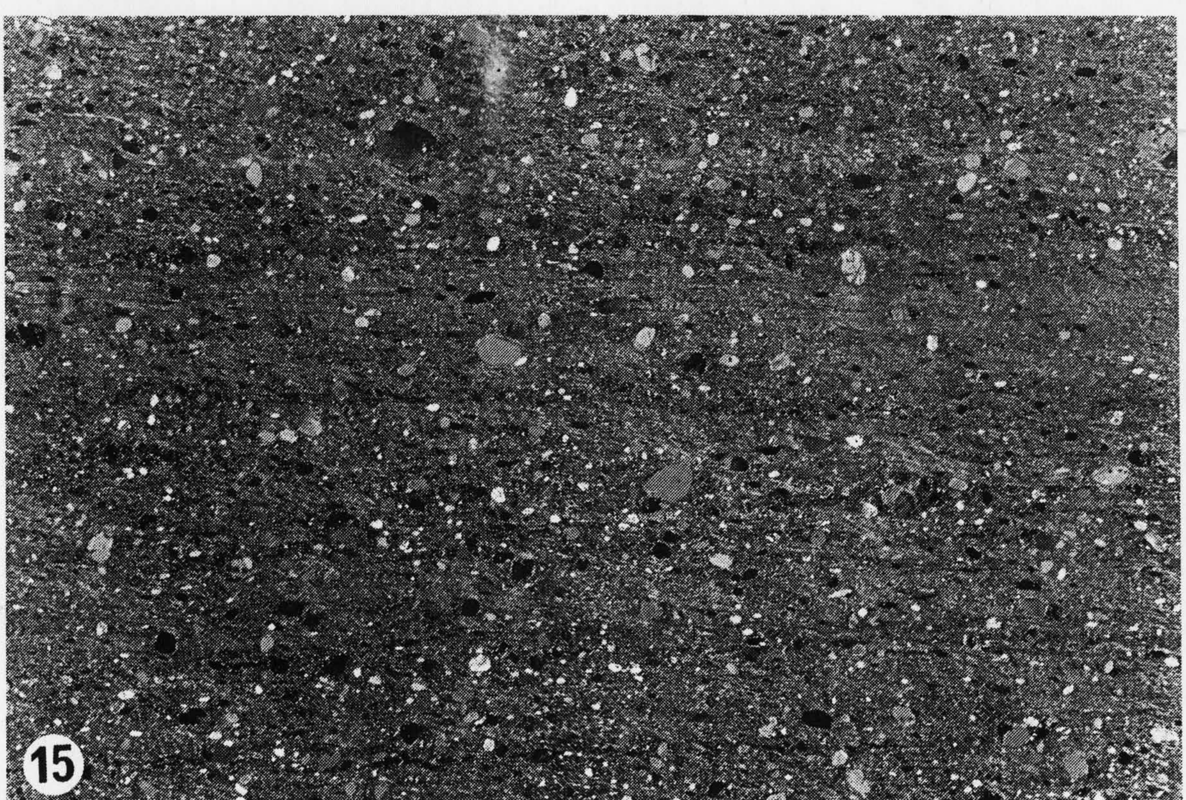
Table 1. Summary of grain size data

**Plate 14.** High strain area photographed under plane polars. A 'relict' fluxion structure is outlined by finegrained dark micas which reveal a prior ductile deformation. The width of the photograph is approximately 2cm.

**Plate 15.** The same high strain area photographed under crossed polars reveals a texture characteristic of brittle cataclasis. The width of the photograph is approximately 2cm.



14



15

### 3.3 Kinematic Indicators

Three types of kinematic indicators are found. These are derived both from outcrop scale and at a microscopic level.

Deflection of foliation and shearing of pegmatites by the shear zones were common features in the field. Asymmetric tails around porphyroclasts are observed in thin section, however these are rare.

Plate 16 shows the deflection of the host rock foliation across an incipient shear. The general sense of motion is sinistral. Small pegmatite dykes commonly show a similar sinistral displacement (plates 17 and 18), however, dextral motion is also observed (plate 19). Note the predominantly ductile behaviour of the sinistral pegmatites through the shear zone, while the dextral offset is taking place along a distinctive break or microfault in the shear. This brittle movement is most likely unrelated in time to those showing ductile deformation. It is possible that the relative movement of shearing has switched from dominantly sinistral during ductile deformation, to dextral during brittle deformation. This, however, cannot be proven based on the present level of research. Microstructures such as asymmetric tails around porphyroclasts are common in mylonites. Due to the overwhelming evidence for cataclasis, however, these features are absent in the Killarney rocks. In plane polarized light, the relict fluxion structures reveal vague tails around porphyroclasts. These are not well defined since subsequent brittle deformation has left an overprint. Both a dextral and sinistral sense of motion is found. This perhaps reflects the prior reworking of the matrix by cataclasis, rendering microstructural kinematic indicators unreliable.

**Plate 16.** Incipient shear displaces foliation sinistrally.

**Plate 17.** Pegmatite dyke showing sinistral deflection across a small shear zone.

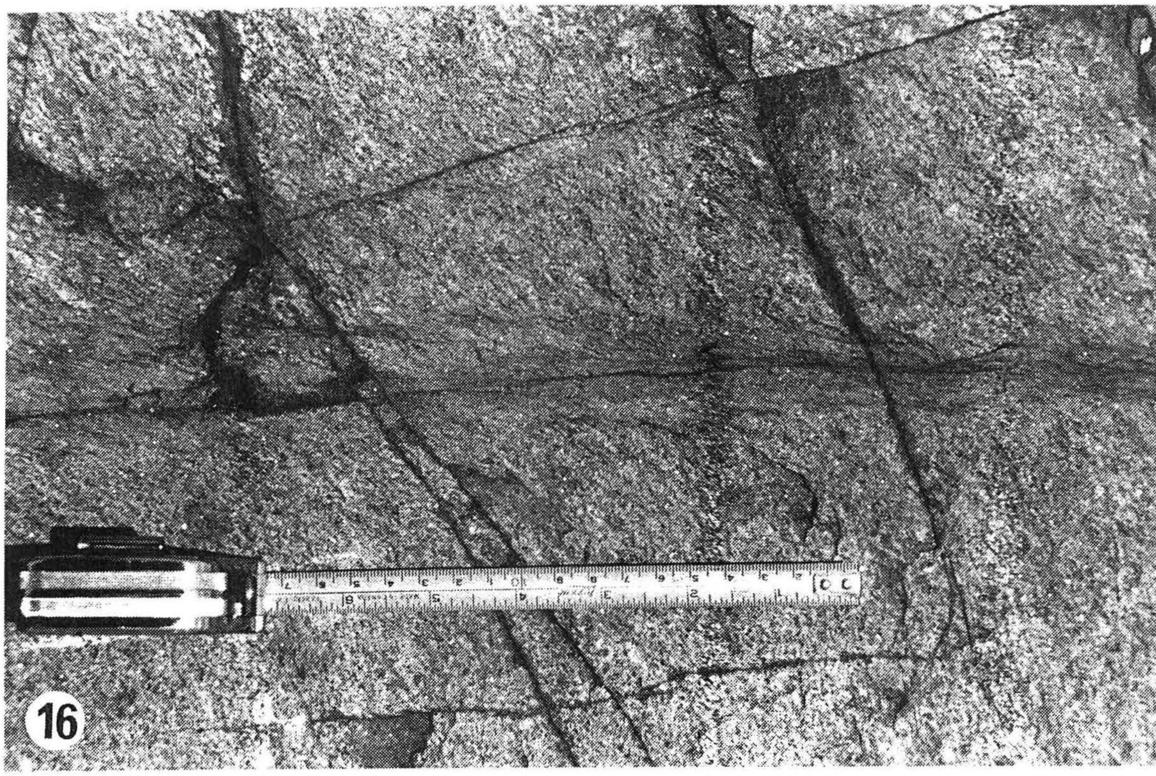
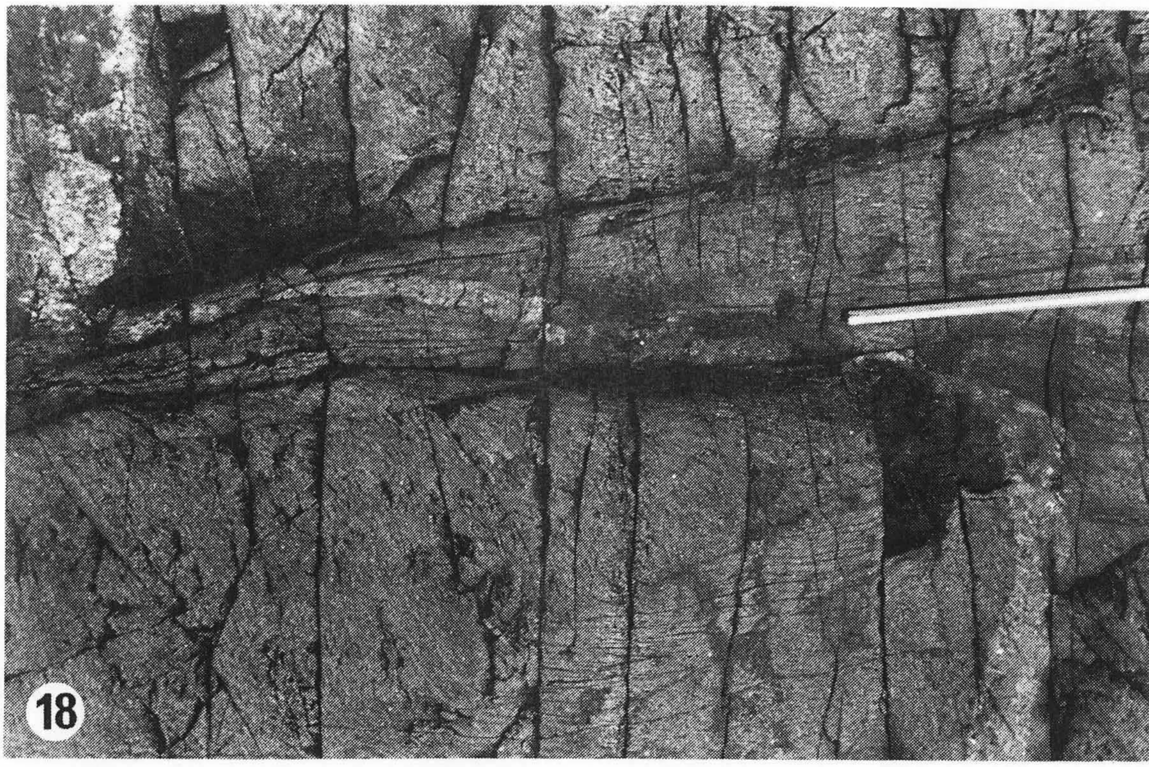


**Plate 18.** Pegmatite dyke showing sinistral deflection across small shear zone.

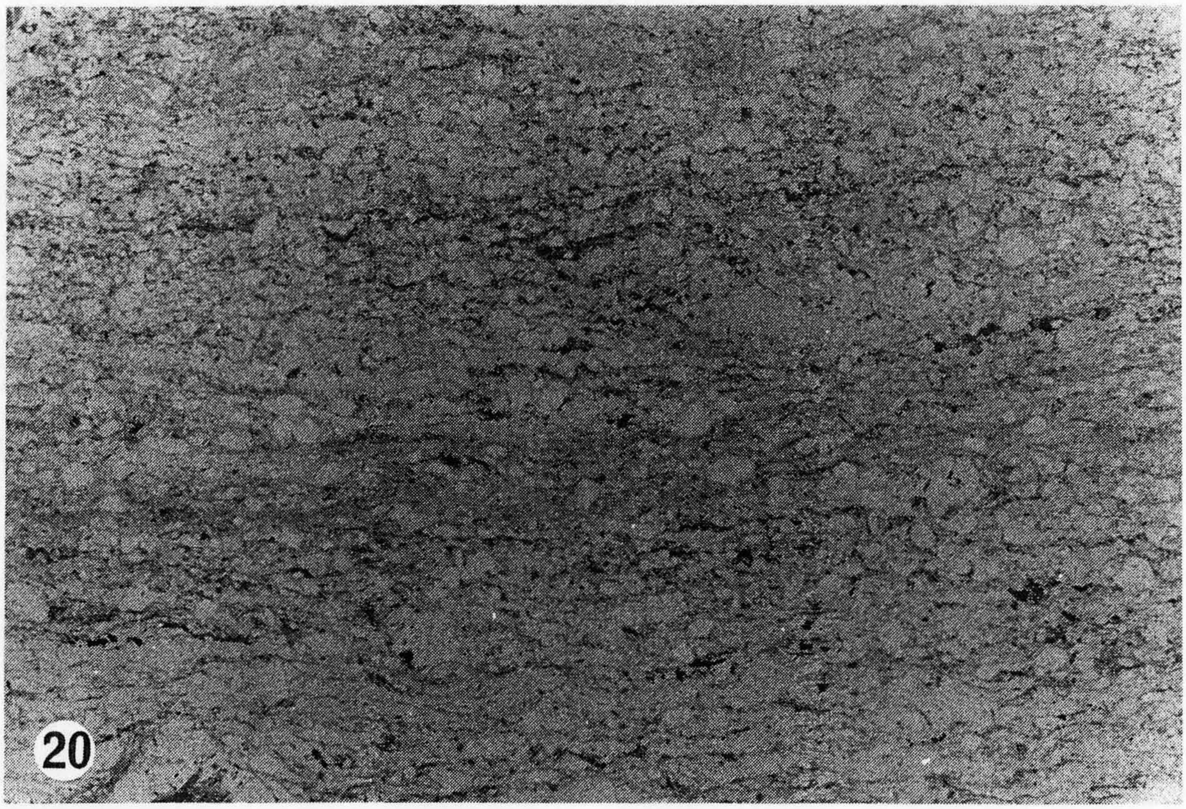
**Plate 19.** Pegmatite dyke brittlely offset by a dextral microfault.



18

19

Plate 20. Feldspar porphyroclast tails demonstrated under plane polars. Both dextral and sinistral indicators are present. The width of the photograph is approximately 2cm.


20

# CHAPTER 4. DISCUSSION/CONCLUSIONS

## 4.1 Strain/Temperature

A combination of ductile heterogeneous simple shear, heterogeneous dilatation normal to the shear zone walls followed by brittle cataclasis has produced the existing shear zone geometries.

To evaluate the shear strain recorded in a shear zone, measurement of strain markers is necessary. Strain markers which may be utilized are outlined by Ramsay and Hubert (1983):

1. Ellipticity of strain ellipses
2. Orientation of strain ellipses
3. Deflection of pre-existing markers such as bedding planes, lithological layering in gneisses, and dykes.

At the outcrop scale elliptical markers are absent in the shear zones being examined due to the scarcity of visible fragmental clasts. As a result of late brittle cataclasis, elliptical markers are also absent on a microscopic scale. Quartz ribbon aggregates, which probably existed prior to the cataclasis have since been destroyed. Furthermore, feldspar porphyroclasts, as previously demonstrated ($Rf/\phi$ plots), lack a systematic increase in axial ratio towards the centre of shear zones. Thus, neither elliptical markers nor their long axis orientations can be applied in the evaluation of shear strain in this particular study.

Pegmatite dykes are common in the area; they are both deflected ductilely across shear zones, and are offset along later brittle faults. These, together with actual shear zone geometries, have been adapted to calculate shear strain. Sinistral offsets are predominant in both dyke and foliation deflections, while dextral offsets are rare. For the purpose of evaluating shear strain, these dextral offsets will be ignored.

Shear strain estimates were derived from three separate dykes (figures 13 through 15) and by analyzing the strain profile across a small shear zone (figure 16) (These figures are tracings from original photographs presented earlier on in this text as plates 17, 18, 6 and 16 respectively). Two dykes show a distinctly ductile deflection across the shear zone while one exhibits a series of brittle offsets. The angle of deflection of pegmatite dykes and foliation across the shear zone for ductile cases (figures 13, 14,and 16) was approximated by a line tracing. The brittle pegmatite was treated as a continuous ductile deflection. The outline of the offsets was approximated by a line tracing (figure 15).

In determining the shear strain profile across the shear zone, shear zone boundaries were averaged as lines. Lines connecting similar crystal fabric orientations were drawn (fabric trajectories). These were subdivided into five areas parallel to the shear zone boundary (figure 16). In each area, the dominant trend of the foliation was marked. These lines were then treated in the same manner as the pegmatite deflections.

Shear strain is evaluated by applying the equation:

$$\cot \alpha' = \cot \alpha - \gamma$$

where the angle of undeflected original pegmatite to a predetermined shear zone boundary is represented as $\alpha$ while $\alpha'$ is the deflected angle of the same pegmatite or foliation to the same trace of the shear zone boundary. The resultant amount of displacement or offset between the two is denoted by $-\gamma$.

Values attained for the strain markers are quite high and range from 14.44 to 10.79 for the ductilely deformed dykes and the maximum value of the shear zone respectively. The brittle dyke displacement, in contrast, shows a much smaller "shear strain" value. Perhaps this is a reflection of a generally lower shear strain for the second, more brittle event.

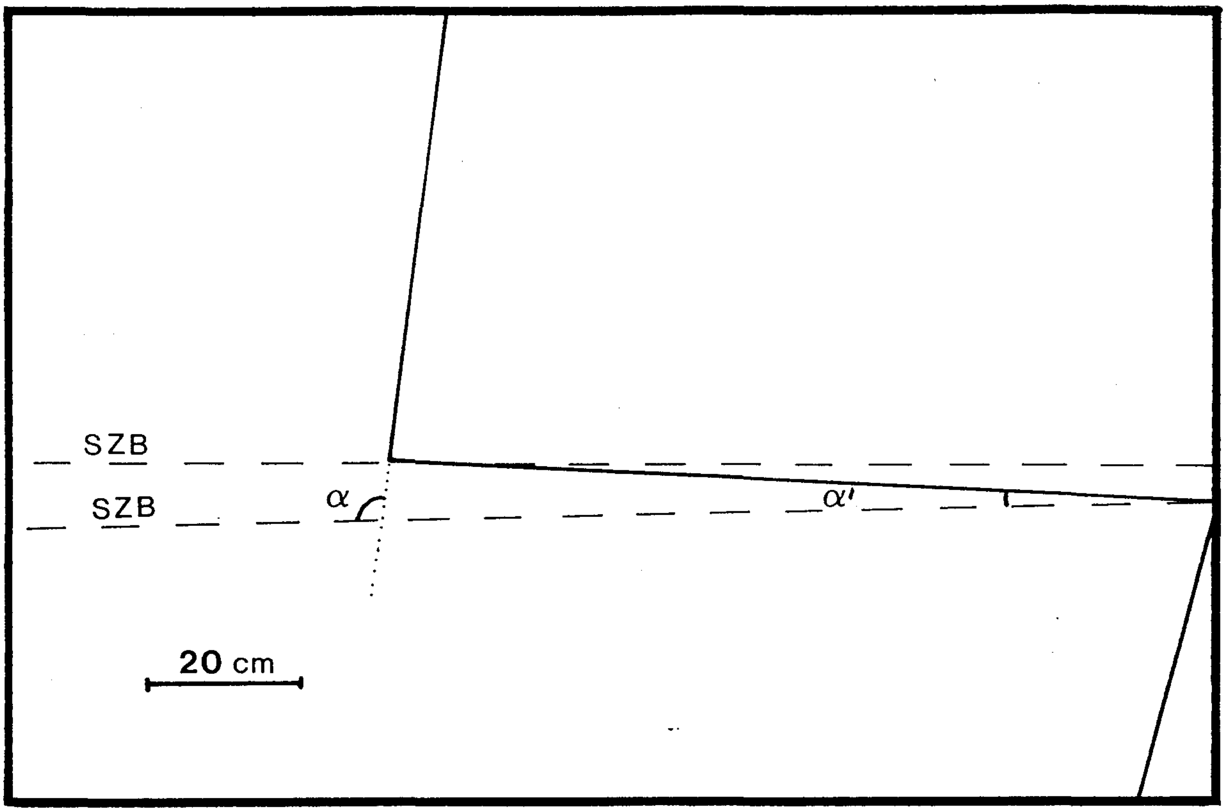
Two distinct temperature regimes have been operating during shearing. Therefore a range of temperatures covering both deformations must be stated in order to represent the complete history of the shear zones.

The only microstructural evidence of ductile deformation is defined by a relict fluxion structure. This structure was once created by the ductile formation of quartz ribbon aggregates and the fracturing and rolling of feldspars. A slight undulatory extinction is present in feldspar grains, however, this is a similar feature to that found in the host rock. Exaggeration of the undulatory extinction has not occurred and therefore no evidence exists which would suggest ductile feldspar deformation. Furthermore, no evidence of a previously existing core and mantle structure is seen. This structure, if present, would give conclusive evidence for the ductile deformation of feldspars (Debat. et al., 1978; Hanmer, 1982)

Thus ductile textures in quartz grains suggests greenschist grade facies (Bouchez, 1977). The lack of ductile behaviour in feldspar, however, indicates that temperatures have not exceeded greenschist facies (White, 1975; Hanmer, 1982; White and Mawer, 1985).

The later, brittle phase of deformation is characterized by the fracturing of quartz grains by cataclasis. Thus temperatures during the later brittle stages of deformation are below lower greenschist facies.

**Figures 13 & 14.** Shear strain constructions from pegmatite dykes ductilely offset by shearing (traced from plates 17 and 18 respectively).


SZB
SZB
α
α'
20 cm

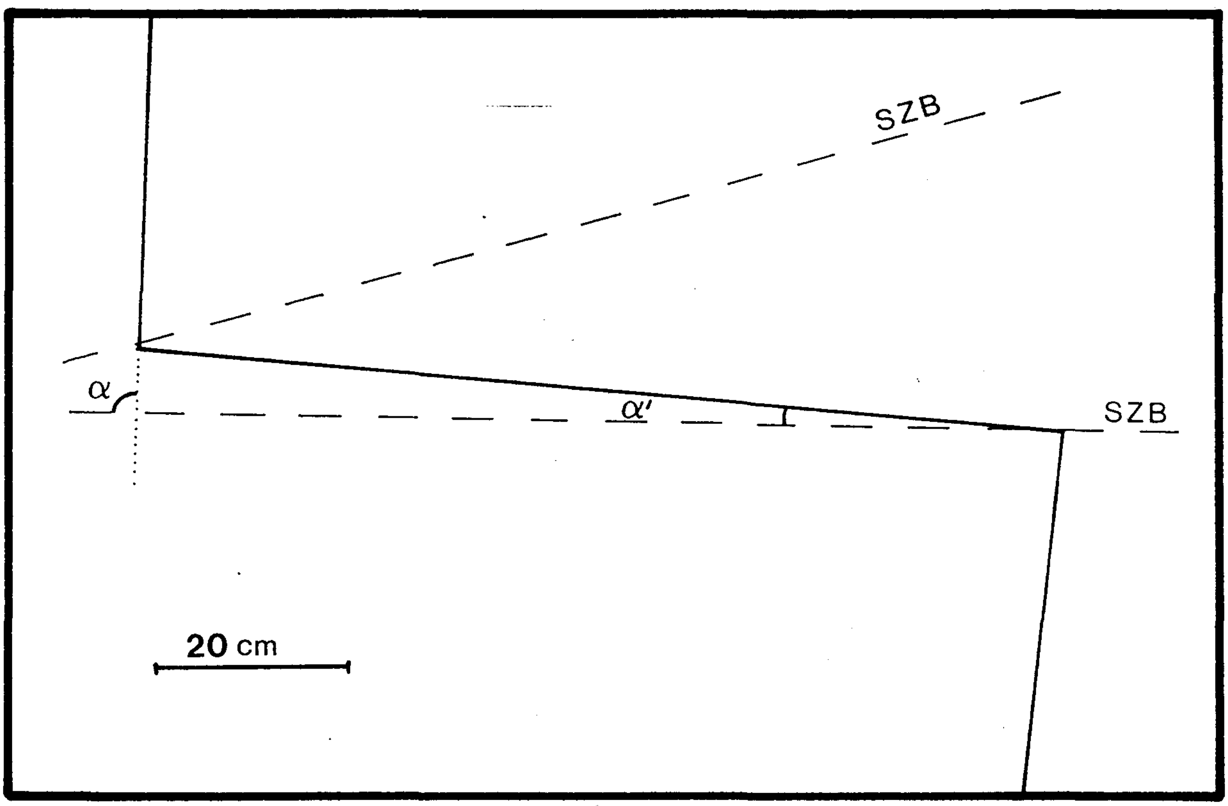
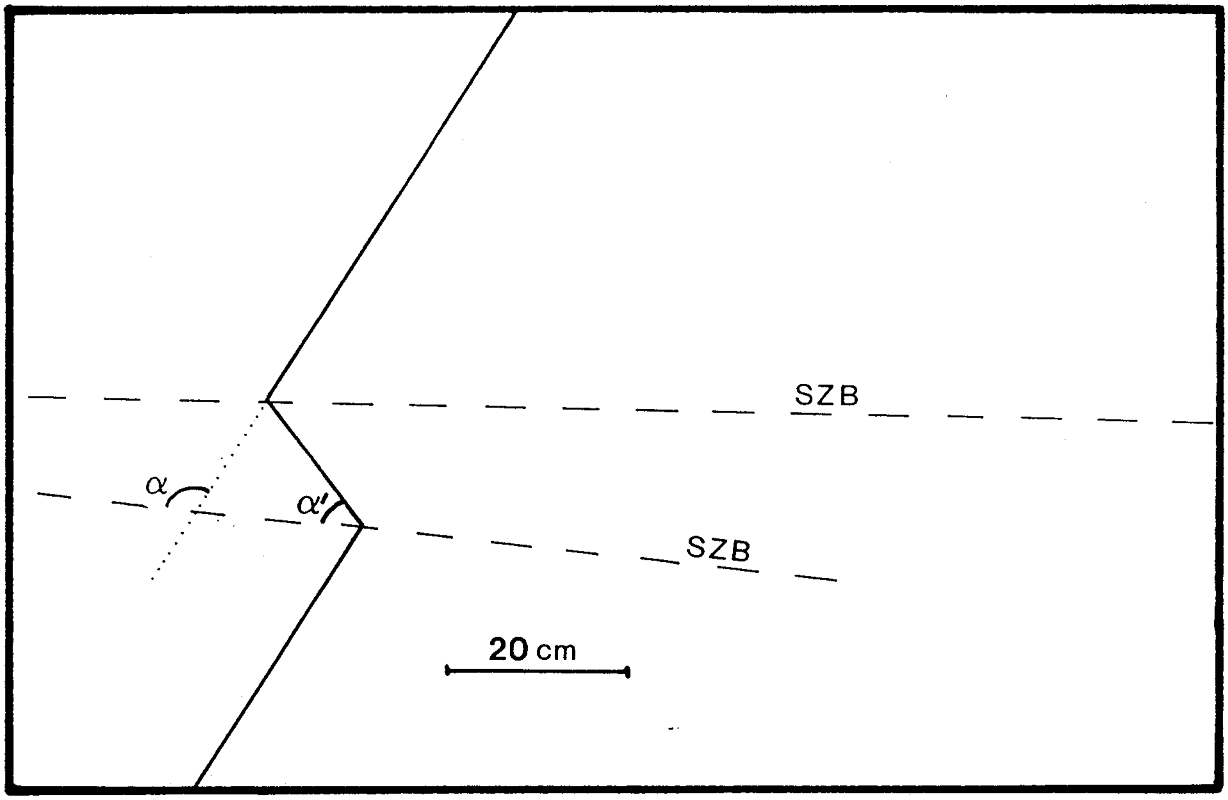
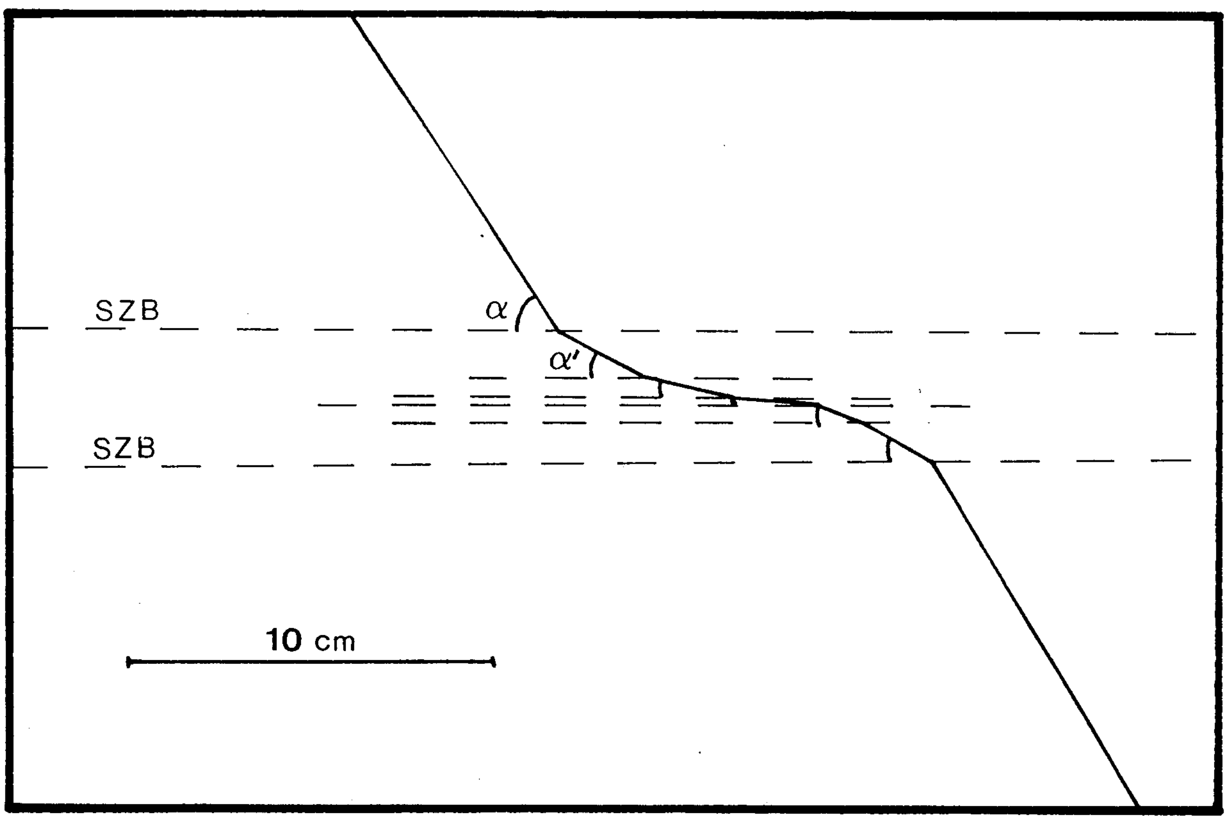

SZB
α
α'
SZB
20 cm

**Figure 15.** Shear strain construction from a pegmatite dyke which has been brittlely offset along several microfaults (taken from plate 6 in text).

**Figure 16.** Shear strain construction from a foliation diverted by a small shear zone (taken from plate 6 in text).


SZB
α
α′
SZB
20 cm


SZB
α
α′
SZB
10 cm

## 4.2 Grenville Relationship

The absence of Grenville Front Tectonic Zone (GFTZ) features within the shear zones prevents direct correlation between these events. However, relative upper and lower age limits can be inferred for the shear zone based on their related geological structures.

The history of the Killarney area can be summarized as follows:

1. Emplacement of granite and porphyry,
   1742 Ma (van Breemen and Davidson, 1988)

2. Flattening and stretching producing a foliation and lineation in the porphyry (Clifford, in press.)
   1625 Ma (Wanless and Loveridge, 1972)

3. Intrusion of pegmatite dykes,
   1400 ± 50 Ma (van Breemen and Davidson, 1988)

4. Sudbury dyke intrusion,
   1250 Ma (Davidson and Bethune, 1988)

5. Grenville event - generation of mylonite zone to the southeast of the Killarney Complex,
   1100 Ma

6. Late brittle - ductile shear zones (MacKinnon and Clifford, 1988)
   Date unknown

The shear zones post-date the foliation developed in the host rock as they cut obliquely across the earlier fabric. They also postdate pegmatite dykes which are deflected by and incorporated into the shear zones. The relationship between the later Sudbury dykes and the shear zones, however, is not known. Direct field evidence relating these two features is lacking. Furthermore, the exact relationship with the GFTZ is not known for similar reasons. The late brittle ductile shear zones described by MacKinnon and Clifford (1988) as en echelon fractures appear to crosscut these shear zones. The time of the formation of the shear zones is therefore constrained to post-dyke emplacement and pre- brittle-ductile shear zone formation. Without a date for this last event one is restricted to inferring that these shear zones definitely occur later than 1400 ± 50 Ma (pegmatite emplacement) and probably later than 1100 Ma (GFTZ).

## 4.3 Conclusions

Based on both field evidence and microstructural relationships two phases of deformation are recognized in the formation of the shear zones. The ductile deflection and brittle microfaulting of pegmatite dykes within the shear zones suggests an initial, predominantly ductile shearing was followed by a second more brittle deformation. Microscopic evidence shows similar brittle-ductile features. Relict 'ghosts' of a prior ductile fluxion structure become obvious in plane polarized light. Crossed polars , however, reveal a late overprint caused by cataclasis.

Values obtained for shear strains were taken largely from ductile strain markers. These produce maximum and minimum values which lie between 14.44 and 10.79. Shearing must have occurred over a range of temperatures in view of the brittle-ductile nature of the shear zone. Greenschist facies conditions were reached during the ductile phase while temperatures dropped below greenschist conditions during brittle deformation. Lower amphibolite conditions were not attained as feldspars do not exhibit ductile features.

Upper and lower age limits for the initiation of shear zones have been determined. They form prior to the emplacement of pegmatite dykes, and up until the formation of brittle ductile shear zones described by MacKinnon and Clifford, (1988) which are post GFTZ. Thus they develop between 1400 ± 50 Ma and approximately 1100 +? Ma. Because a GFTZ overprint is absent in the shear zones it is possible that these could be upper crustal equivalents of the GFTZ. If this is the case, a later brittle deformational event could have been triggered around the time that the brittle - ductile shear zones described by MacKinnon and Clifford were formed.

In order to further constrain the timing of the shear zones relative to the Grenville event, more detailed mapping close to the mylonite zone is required.

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| | FOLIATION | LINEATION |
|---|---|---|
| SHEAR ZONE #1B | 223,86 | - |
| | 229,88 | - |
| | 247,82 | - |
| | 205,88 | - |
| | 224,84 | - |
| | 223,87 | - |
| SHEAR ZONE #2 | 294,83 | - |
| SHEAR ZONE #3 | 278,78 | - |
| SHEAR ZONE #4 | 097,72 | - |
| SHEAR ZONE #5 | 203,79 | - |
| | 173,74 | - |
| | 175,87 | 203,60 |
| SHEAR ZONE #6 | 165,81 | - |
| | 156,76 | - |
| SHEAR ZONE #7 | 183,84 | - |
| | 192,73 | - |
| SHEAR ZONE #8 | - | - |
| SHEAR ZONE #9 | 350,90 | - |

## APPENDIX #1: SHEAR ZONE FOLIATION AND LINEATION DATA



| | FOLIATION | LINEATION |
|---|---|---|
| SHEAR ZONE #1 | 194,86 | 015,15 |
| (SOUTH END) | 197,72 | 340,65 |
| | 261,68 | 315,66 |
| | 252,90 | 198,83 |
| | 253,84 | - |
| | 062,70 | 147,72 |
| | 250,87 | - |
| | 212,96 | - |
| | 254,78 | 331,78 |
| | 253,84 | - |
| | 255,81 | - |
| | 267,83 | - |
| | 257,86 | 328,83 |
| | 257,82 | - |
| | 247,81 | 287,72 |
| | 255,75 | - |
| | 260,70 | 313,70 |
| | 254,86 | - |
| | 218,79 | - |
| | 228,77 | - |
| | 240,77 | 008,75 |
| | 217,76 | - |
| | 228,67 | - |
| | 219,84 | - |
| | 031,86 | 035,56 |
| | 028,78 | - |
| | 027,88 | - |
| | 037,74 | - |
| | 023,82 | 067,68 |
| | 028,87 | - |
| | 033,84 | - |
| | 037,68 | - |
| | 025,72 | 038,36 |
| | 051,78 | - |
| | 023,67 | - |
| | 022,63 | 025,15 |
| (NORTH END) | 057,54 | 078,33 |