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or

The Chief Geologist
Mount Isa Mines Limited
Mount Isa, Qld, 4825.
THE MICROFABRIC OF A DEFORMED QUARTZITE
SEQUENCE, MOUNT ISA, QUEENSLAND

A thesis submitted for
the degree of
DOCTOR OF PHILOSOPHY

in
THE AUSTRALIAN NATIONAL UNIVERSITY

by

CHRISTOPHER JOHN LASCELLES WILSON

August 1970
PART I

INTRODUCTION AND GENERAL GEOLOGY
PREFACE

This thesis incorporates the results of approximately four years work on the structure in a deformed sequence of rocks exposed at Mount Isa north-west Queensland, Australia. The field and laboratory work was carried out by myself and all information compiled from other workers is clearly acknowledged. Any work that has been done in collaboration with other workers, Appendices B and C are accompanied by a statement outlining the nature of the work carried out by each of these workers.

The work on which this thesis is based was begun in May 1966 while I was employed in the Department of Geology and Geophysics, University of Sydney until August 1967 and subsequently in the Department of Geophysics and Geochemistry at the Australian National University Canberra, where I held an Australian National University Research Scholarship. I am grateful to Professors C.E. Marshall and particularly J.C. Jaeger for the facilities they have made available to me in their respective Departments.

The study was supported by Mount Isa Mines Ltd who generously provided accommodation, vehicles and most needs connected with the field work in the Mount Isa area. During these periods of fieldwork they made available their laboratory facilities. The Company also generously provided help by redrawing Map 1, the Geological Cross Sections, Figure 2.2 and also supplied the copies of Map 2. For the drafting and the colouring of Map 1 I would like to especially thank J.L. Cox, D. Hardwick, T. Nettle and Miss L. Payne.

While in Mount Isa I benefited greatly by discussion with members of the staff of Mount Isa Mines Ltd., too numerous to be included here, but I would like to acknowledge the help and information provided during my initial stays in Mount Isa by C.G. Battey, E.M. Bennett, E. Davies, W.J. Smith and Dr P.J. Solomon all of whom have since left Mount Isa. During these and subsequent visits Dr N.J.W. Croxford and Mr R.L. Hewett have provided many fruitful discussions and have extended enormous help to me.
During all phases of the work I have greatly benefited from discussion with M.A. Etheridge, Dr R.B. Farquharson, Dr K.R. Rosengren, Dr R.H. Vernon, Dr P.F. Williams and other colleagues at the University of Sydney and the Australian National University. I would like to add an extra note of thanks to Dr P.F. Williams who spent some time in the field with me and to Dr K.R. Rosengren for the writing of the computer programs used to compile the data presented in this thesis. Messrs. A. Powell and E. Pederson ably made the thin sections, G.T. Milburn provided photographic and drafting advice and the final photographs were processed by the Visual Aids Unit, Australian National University.

Dr B.E. Hobbs originally suggested the Mount Isa district as a suitable area and has supervised the project. Through this enthusiasm and interest in the project he has provided great encouragement and advice both in the field and laboratory which are very much appreciated.

I am grateful to Miss Jill Chapman, Mrs Fiona Aichinger who typed the draft and Mrs Julie Barton for the final typing of this thesis.

Canberra, July 1970

C.J.L. WILSON
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Map 1. Geological Map of an area west of Mount Isa.


Map 4. Distribution of quartz c-axis preferred orientations west of Mount Isa.
This thesis is basically a study of the change of microfabric, which includes microstructure and preferred orientation, with change in metamorphic grade in a deformed Precambrian quartzite sequence at Mount Isa. The thesis is divided into two parts, the first is concerned with establishing the stratigraphic relations, structure and metamorphism in an area immediately to the west of the Mount Isa Mine, the second is a detailed microfabric study of this area supplemented with specimens collected in the same quartzite units but in different structural and metamorphic environments to the east and north of Mount Isa.

Three periods of folding have been recognised. The first are tight folds which transpose bedding locally into an orientation parallel to the axial planes of the folds. A slaty cleavage is the axial plane structure, which at the higher grades of metamorphism is a well developed metamorphic layering. Post-dating this folding are regional faults (tectonic discontinuities) which are folded by second generation folds, characterised by a strain slip cleavage as axial plane structure. These folds are found as either open folds or as zones of intense deformation, which may be associated with large displacements such as the Mount Isa Fault. Accompanying the second generation folding is the intrusion of the main mass of the Sybella Granite. The third generation folds are confined to the zones of intense second generation folding; they are open kink folds which lack any axial plane structure. The faulting accompanying the second generation folding divides the Mount Isa area into four structurally identifiable blocks, each of which contains limbs of larger first generation folds overprinted by second generation folds. A general stratigraphic correlation is possible between these blocks.

The metamorphic grade differs in each one of these structural blocks; increasing to the west with zones containing pelitic rocks characterised by chlorite, biotite, biotite-cordierite and sillimanite. Retrograde metamorphism has also been recognised in the area and is associated with second generation folding.
Both prograde and retrograde microstructural changes accompany this metamorphism, having been studied in the purer quartzites. The retrograde quartzites are characterised by a progressive change from round detrital quartz grains to the development of flat elongate old grains, or portions thereof, surrounded by or completely replaced by abundant new grains, which have a diameter of no more than 5 microns. In the prograde sequence rounded old quartz grains become flat and elongate in the foliation and contain deformation bands, deformation lamellae and many new grains in the chlorite and biotite zones, have polygonal shapes in the biotite-cordierite zone, and abnormally large quartzes (up to 2 cm in diameter) are observed in the sillimanite zone. The successive changes in the microstructure are described. The quartz c-axis patterns accompanying such microstructures change successively from random to peripheral girdles, to crossed girdles to random to either very strong girdles or maxima.

These microfabric observations are interpreted in the light of recent metallurgical theories and the results obtained from hot working experiments and experiments performed on geological materials. It is believed that the variation in microfabric may be attributed to the operation of processes involving either intragranular slip mechanisms or recrystallization. Both mechanisms would be occurring simultaneously but in different parts of the sequence. The role of a host control appears to be significant in all stages of microfabric development and quantitative measurements were obtained in the low grade quartzites where there is also a very strong relationship to the strain.
CHAPTER 1
INTRODUCTION

1. General Statement

Microstructural and preferred orientation studies have been used by many geologists since the classic work of Bruno Sander (1930) in an attempt to unravel the physical conditions existing during the formation of metamorphic rocks. Most studies have been concerned with the minerals quartz, mica, calcite, dolomite, olivine and plagioclase, but quartz has been used in most fabric* studies because of its ubiquitous occurrence in metamorphic terrains and its relatively simple optical properties in comparison to other rock forming minerals. The past decade has seen a considerable number of papers, mainly of experimental nature, which have clarified the interpretation of quartz microfabrics. For the most part, these papers have duplicated the experience gained since 1950 on the deformation and recrystallization of metals. The aim of this thesis is to apply the information gained

*Fabric is used here in the sense of Paterson and Weiss (1961, p. 860) who use it as a concept which includes the geometrical and spatial arrangement of elemental parts of an aggregate at any scale.
in such studies in the hope of arriving at a clearer understanding of physical conditions existing during the deformation and metamorphism of a quartzite sequence.

This investigation of some regionally metamorphosed rocks at Mount Isa, north-west Queensland is an attempt to compare the quartz preferred orientations and microstructure with changes in the mode of deformation and of metamorphic grade. The portion of the sequence studied is dominated by pure, medium to coarse grained quartzite comprising an area which superficially appears to have a very simple structure. It soon became apparent that much of the previous work was inadequate and that detailed mapping and a structural analysis were necessary before the microfabric investigation could begin. The investigation is therefore divided into two parts:

1. General Geology, which includes the stratigraphy, regional metamorphism, and structure.

2. Microfabric study, which involves a study of the changes in (1) Microstructure, and (2) the Preferred Orientation of $c$ axis patterns in the quartzites with changes in metamorphic grade.
A discussion of the background for the microfabric study is given below. This deals first with the means of developing different preferred orientations and then discusses the various types of microstructure and previous investigations of microstructure relevant to metamorphic rocks.

2. The Development of Preferred Orientations

It is not the purpose of this section to provide a detailed review of aspects of preferred orientation development. Rather, the intent is to indicate the important papers in this field and to briefly indicate some of the more important conclusions. A general familiarity with geological and metallurgical treatments of this subject is assumed. Many aspects of these introductory comments are considered in detail in Chapters 5 and 6.

By using simple techniques which include the flat stage, the universal stage, X-ray film (Starkey, 1964; Wenk, 1965a) and counters (Schulz, 1949; Decker et al., 1948) it is possible to determine the fabric of a quartz aggregate within a rock. Most workers have concentrated on establishing the orientation of the c axis in quartz, and only recently have measurements of other
crystallographic directions been made, namely the a axis, Barker et al. (1969), Wenk et al. (1967). The advantage of having this additional information is that it is possible to know the complete orientation of the crystal, so that grain boundary faces may be identified and the orientation of possible slip systems within a given crystal can be determined. Considerable information on preferred orientation in experimentally deformed rocks has emerged since the early work of Griggs, Turner and colleagues (See Griggs and Handin, 1960), but still the picture is obscure and many interpretations still need to be based on analogue with metals.

Analogies between deformed metals and rocks have been quoted by many geologists over the past century, but the variables considered by the geologist may be more nebulous than those considered by the metallurgist, for the geologist is often unsure of the past history of the material. In experiments with known starting material the geologist is aiming at reproducing naturally deformed materials. This involves setting up experimental conditions similar to those of the metallurgist. Although, all the basic experimental work on quartz (Carter et al., 1964; Green, 1966;
Hobbs, 1968) have involved sets of experiments designed for conditions of creep, cold and hot working, with or without annealing, the preferred orientations developed in these experiments and others have been attributed to two different kinds of processes, namely, intracrystalline and recrystallization mechanisms.

(i) **Intracrystalline Mechanisms**

Experiments in which single crystals of known crystallography have been deformed, have established that deformation occurs by slipping or twinning on particular crystallographic planes and in certain directions, only when a critical shear stress in the direction of slip is exceeded (Schmid, 1925). The slipping produces a rotation of the axes of the specimen relative to the axes of the crystal and this rotation ultimately brings extra slip planes into more favourable positions for slipping. von Mises (1928) and subsequently Taylor (1938) have described the general deformation of aggregates of randomly oriented grains which may achieve any imposed shape change by shear on five independant slip systems without any volume change. The Taylor-von Mises criteria for analysis is based on the concept of minimum (internal) work, and involves finding amongst numerous possible slip systems the
combinations that satisfy an imposed strain, in which the sum of the glide shears is a minimum. (See Taylor, 1956). This assumes that the strain is homogeneously distributed and hence the strain undergone by an individual grain is the same as that experienced by the aggregate.

Bishop and Hill (1951) have pointed out that in face centred cubic crystals, if five slip systems are to be chosen from the sets of either six or eight systems available the magnitude and sense of the crystal rotations will depend on which five systems are chosen. If all six or eight systems operate there may be no crystal rotation or the net rate of rotation on certain portions will be slower when several equivalent sets of shears are available, than when the operative slip systems are unambiguously determined.

Taylor's (1938) original treatment assuming homogeneous strain was largely dismissed because:

(i) The work of Barrett and Leverson (1938) showed that Taylor's predictions were not true for at least $\frac{1}{3}$ of the grains they studied and

(ii) It is generally observed that the strains in deformed aggregates depart greatly from homogeneity.
The first of these objections was more or less eliminated by the work of Bishop and Hill (1951) who pointed out that Taylor had been wrong in some of his calculations (see Taylor 1956). The correction of Taylor's work produces predictions which are largely in agreement with Barrett and Leverson's (1938) observations.

The second of these objections has been recently removed by Wonsiewicz and Chin (1970) who have demonstrated that the Taylor approach is also capable of handling the case of inhomogeneous strain within grains.

The Taylor approach (or the Bishop and Hill approach) therefore, now stands as the only physically realistic theory which enables an analytical approach to the plastic deformation of aggregates.

The approach of Bishop and Hill (1951) was to define a plastic yield surface. This being the surface defined for any crystal undergoing a plastic deformation in which the state of stress required to produce any given strain increment must lie. The stress causing that strain increment can then be determined. Chin et al. 1967 and Chin and Mammel (1967) have since shown that the Bishop and Hill approach is equivalent
to Tailors Theory and it is possible to calculate the
crystal rotations undergone during deformation.

If a polycrystalline aggregate is to sustain
a large plastic stain by deformation (Paterson, 1969)
on less than five independant slip system then a
general change of shape cannot occur by glide alone
(von Mises, 1928) without discontinuous behaviour at
grain boundaries. Then other mechanisms such as
cross-slip (Barrett and Massalski, 1966), and
dislocation climb will be involved which generally
occurs at elevated temperatures or slow strain rates.
This has been formulated and treated in detail by
Groves and Kelly (1963, 1969) who have considered
strains in non-metals which can be produced by climb,
by climb in conjunction with unrestricted glide and
by climb in conjunction with glide or specific
crystallographic planes. The effect of an increase in
temperature is to bring about climb an essential
process in the phenomenon of recovery which includes
all the annealing phenomena occurring before the
appearance of recrystallized grains.*

*Recrystallization is used throughout the rest of the
thesis for new grains which have developed as a
result of the migration of high angle grain
boundaries.
Polygonization is the most widely recognised recovery phenomenon and results from the action of cross-slip and/or climb of dislocations. The dislocations rearrange themselves into quite distinct regular polygonal walls which eventually form subgrains with central regions of very low dislocation densities (Vandermeer and Gordon, 1963). Polygonization may be either a thermally activated process, or may occur concurrently with deformation under conditions of low temperature and/or high strain rates. These two processes have been called static and dynamic recovery (Jonas et al., 1968) respectively.

The effect of heating a cold worked material increases the chances that thermally activated processes would occur. This results in movement of dislocations which would normally only move in response to the stress field of other dislocations, and is the basis of static recovery. Dynamic recovery, however, may be produced by both cold and hot working a material.

A number of metallurgists including Hardwick and Tegart (1961) and Ormerod and Tegart (1963) have found that in a number of metals recrystallization generally does not take place during a hot working deformation but only after the deformation. This may be the case in some metamorphic rocks especially those
deformed at low temperatures and/or high strain rates. This behaviour has been rationalized by these authors in terms of the relative stacking-fault energies of the metals, since these control the relative rates of climb of edge dislocations to form sub-boundaries. A pure climb process may be applicable to some silicates, developing a new preferred orientation, but it is not to quartz (Hobbs pers. comm.) which apparently has high stacking fault energy and has many possible paths available for cross slip to take place. This has been verified by McLaren (pers. comm.) who, by use of the Electron Microscope has shown that dislocations in quartz are never split therefore the stacking fault energy must be very high.

On the otherhand it has been shown that microstructures consistent with recrystallization may definitely accompany hot working deformation in some metals (Sellars and Tegart, 1966; Jonas et al., 1969; Sah et al., 1969). A similar recrystallization phenomenon, known as dynamic recrystallization, has also been observed in the course of creep, but at lower strains and strain rates (Gifkins, 1958; Richardson et al., 1966).

The predictions of von Mises (1928) and Taylor (1938) have been strengthened by later workers such
as Bishop and Hill (1951), Chin and Mammel (1967) and it has been shown in metals, (Dillamore and Roberts, 1965; Hu, Cline and Goodman, 1966; and Barrett and Massalski, 1967) that by only using slip it is possible to produce strong preferred orientation. Hobbs (1970) recently presented a formalised method of predicting preferred orientation in quartz aggregates and has also shown that strong preferred orientations may be developed in other silicates by these processes. The development of preferred orientations in carbonate rocks without recrystallization has also been recognised by Griggs et al. (1960b) and this is again attributed to intracrystalline mechanisms such as twinning and translation gliding. Although much of the older geological literature attributes patterns of preferred orientations to slip mechanisms many of these rocks are strongly recrystallized, and therefore any pre-existing preferred orientation would have been modified by grain growth. Also many of these postulated slip mechanisms have not been verified experimentally. (Carter et al., 1964; Christie et al., 1964; Christie and Green, 1964; Heard and Carter, 1968 and Baeta and Ashbee, 1969a and 1969b)

(ii) Recrystallization

Recent geological interpretations have accounted for observed patterns of preferred orientations by
a mechanism of recrystallization where the pattern can be correlated with pronounced strain in individual grains as a result of plastic flow and non-elastic deformation in a stress field. Another theoretical basis for the development of recrystallization fabrics are the Gibbs-Kamb (Kamb, 1959a, 1961) and MacDonald (1960) theories, which have been applied by Brace (1960), Hartman and den Tex (1964) and De Vore (1966, 1969a,b) to predict the types of preferred orientations that can be developed with a particular mineral species due to elastic behaviour.

In the Gibbs-Kamb Theory the predicted preferred orientations are those developed under non-hydrostatic stress for crystals of known elastic compliances which have been subjected to infinitesimal strain in known crystallographic directions. It is believed by these authors that the recrystallization phenomena is generally a solution - growth process on the boundaries of an old crystal, in a polycrystalline material, and only new crystals with the (highest strain energy) will grow at the expense of the less favourable oriented grains. The preferred orientation is believed to be that which minimizes the chemical potential across the plane normal to the maximum stress direction.
These theories appear to be inapplicable to plastically deformed crystals or rocks, as was pointed out by Kamb (1959). These theories also completely disregard the possibility of annealing recrystallization which depends upon the formation and subsequent growth of new grains through a dissipation of the stored energy of deformation. Many experimental studies performed by geologists; Griggs et al. (1960 a, b), Carter et al. (1964), Raleigh (1965), and Hobbs (1968), have been attributed to the latter mechanism where the residual strain energy and the grain boundary energies constitute the potential which drives the recrystallization.

When deformed crystalline materials, which include quartz, are annealed at temperatures higher than those required for recovery, the remaining stored energy of deformation is released through the migration of high angle boundaries (Shewmon, 1966; Gordon and Vandermeer, 1966), resulting in the growth of new unstrained grains at the expense of old deformed host grains. As the high angle boundary migrates into the old strained grain; the old grain is replaced by a new grain of locally differing orientation. The growth of these strain free grains
usually starts independently at many different locations, nuclei, in the deformed material. The ultimate result of this grain boundary migration is the formation of a polycrystalline polygonal structure by a process known as primary recrystallization.* (Beck and Hu, 1966). This sometimes cannot be

*The term primary recrystallization is used in most metallurgical literature to mean isothermal recrystallization of cold worked materials, the process obeying certain kinetic laws. These laws were defined by Burke and Turnbull (1952) for isothermal annealing and can be represented empirically by either the increase of average grain size with time or by percentage of material recrystallized versus time. The kinetics for recrystallization involve an incubation period after which new grains begin to grow at a number of sites. In most of the classical studies of recrystallization (cf. Beck, 1954), both primary and secondary recrystallization are described in terms of such a process, namely the nucleation of new grains and their growth. This assumes that primary recrystallized grains form at the onset of the release of stored energy for grain growth and are independent of any pre-existing stages. The normal grain growth recognised by metallurgists is commonly considered as an intermediate stage in the formation of the equilibrium polygonal microstructure.

In much of the recent metallurgical literature (cf. Hu, 1969; Sellers and Teggart, 1968) many workers now believe recovery processes, which involve the formation of subgrains, may act as nucleation sites for subsequent recrystallization either by a coalescence of adjacent subgrains or a subgrain rotation mechanism (these will be discussed further in Chapters 5 and 6). These then grow by the migration of high angle boundaries. In this thesis such a process will be separated from the classic nucleation ideas for primary recrystallization. It is believed that subgrains formed as a result of recovery, and these

(Continued on next page)
sharply separated from the process known as secondary recrystallization where the boundary migration does not appear to be wholly dependent on the strain energy or work hardening in the old grains, (Dunn and Walter, 1966).

During the past decade there have been a number of excellent reviews on the development of preferred orientation in primary recrystallized metals. Two of these reviews, Wassermann and Grewen (1962) and Dillamore and Roberts (1965), have summarised all data available concerned with deformation "textures" (preferred orientations) while the reviews of Cahn (1966, 1969) have discussed possible mechanisms of recrystallization. Beck and Hu (1966), on the otherhand, discuss the particular question of orientation relationships between the deformed material and the recrystallization nuclei, and the status of the two mechanisms, "oriented nucleation" and "oriented growth" for the formation of annealing textures.

* (Continued) recovered subgrains act as the nuclei for the primary recrystallized grains. Therefore no true nucleation stage exists instead the primary recrystallized microfabric involves migration of high angle boundaries and normal grain growth where grain boundaries have migrated so as to minimize their interfacial tensitional forces to satisfy the equilibrium requirements of a polygonal grained structure (see discussion in Chapter 1, section 3).
Hu (1969) in a recent study of nucleation, in heavily rolled copper, presented strong evidence that nucleation of new recrystallized grains bears a strong relationship to the host grain. But subsequent anisotropic boundary mobility then controls the pattern of preferred orientation. The primary recrystallization preferred orientation then depends on a selective growth mechanism.

The role of this host control phenomena has also been noted by a number of geological workers in the experimental field, both with calcite (Griggs et al., 1960 a, b) and quartz (Hobbs, 1968). Hobbs is the only worker to present quantitative data, which show that significant orientation relationships exist between new and old grains under different conditions of deformation and recrystallization.

The preferred orientation developed by secondary recrystallization generally differs significantly from the preferred orientation in the surrounding primary recrystallized matrix grains but is highly dependent on the matrix fabric. The nuclei for secondary recrystallization must be present in the primary recrystallized matrix and the driving force requirements for their boundary migration has recently been reviewed by Dunn and Walter (1966) and Walter
(1969). The boundary migration appears to be dependent on a number of factors which include the surface energy of the boundaries between the matrix and secondary grains, on the boundary curvature, its angle of disorientation and the possible concentration of solute atoms in the boundary.

In the case of a matrix with a strong single preferred orientation with a low average grain boundary energy, $\gamma_b$, for growth of the secondary to occur,

$$\frac{R_{sec}}{R_m} \frac{\gamma_{b}}{\gamma_{b}}$$

Where $R_{sec}$ = Average grain radius of potential secondary

$R_m$ = Average grain radius in stable matrix

$\gamma_{b}$ = Average energy of boundary between primary and secondary.

In the case of a strong matrix preferred orientation $R_{sec}$ would be fairly large. The orientation with the highest driving force is that of the average orientation of the matrix but since the mobility is probably quite low for such an orientation, the rate of growth of any potential secondary will be small. With a stable matrix of weak preferred orientation, these will be of a high average angle of disorientation, high average
mobility and high average driving force. Under these conditions a potential secondary grain would have a better chance of growing provided its growth was not inhibited by inclusions.

3. Development of Ideas on Microstructure

In 1960 Voll showed the possible application of physical metallurgical principles to a study of microstructure in silicate and carbonate rocks. The first effective application of such techniques was that of Stanton (1964 a, b), who conducted a quantitative study on natural metamorphic sulphide rocks and presented experimental results on synthetic sulphide aggregates. Detailed studies of silicates have recently received attention from such workers as Hobbs (1966), Kretz (1966a, b, 1969), Read (1965) with a very comprehensive study being undertaken by Vernon (1965, 1968) who described the microstructures in granulite and amphibolite grade rocks from Broken Hill. This latter work was continued by Ramsom (1969) with a detailed study of recrystallization of quartz and mica in retrograde zones.

Most of the above work has been primarily concerned with describing interlocking aggregates of polygonal grains, in rocks resulting from primary
recrystallization using the general guiding principles and ideas presented by Harker and Parker (1945) and Smith (1948, 1953, 1954 and 1964). The theory developed by these authors governs the arrangement of randomly oriented three dimensional "domains" (cells) so that they (i) completely fill space and (ii) produce a minimal boundary free energy.

These authors have been concerned with the relationship between grain size, shape, and distribution to grain growth in polycrystalline aggregates of either one or two phase material. Their theory, which is guided by certain topological principles, only applies to ideal cells or grain configurations; (Smith also applies the general principles to such diverse aggregates as foams and organic cellular structures), where the grain boundary interfacial energy does not vary significantly with relative grain orientation.

Smith (1964) has emphasised the importance of past history on the present appearance of polycrystalline aggregates. He also considers that there are two basically different but dependent modes of growth: (i) those growing in effective isolation, and (ii) those growing under the influence of other
crystals. The first type of growth could be regarded as a forerunner to the second. In the early stages of growth of stable phases, new grains nucleate at favourable sites in the old grain of the unstable phase. Once grain growth begins then it will possibly continue until its source of free energy is expended or subsequent growth is inhibited by another phase.

The grain shapes these authors were concerned with are those found in annealed aggregates, and consist of primarily recrystallized polygonal grains of fairly low interfacial energy. As a result of extensive study, the origin and nature of the driving forces for recrystallization are now understood. A review of this work has been given by Wassermann and Grewen (1962), Dillamore and Roberts (1965), and Beck and Hu (1966). The factors which determine the mobility of the grain boundary are, however, not nearly so well understood and have been reviewed by Lücke and Stüwe (1963) and Aust (1969). Nor are the processes involved in recovery which precede the development of primary recrystallization. These processes have been reviewed by Vandermeer and Gordon (1963) who show that recovery can exert an influence on the kinetics of recrystallization and the rate of grain boundary migration.
Secondary recrystallized aggregates are characterised by "discontinuous grain growth" or "coarsening" of a relatively few grains which become very large, and absorb the finer polygonal primary recrystallized matrix. The microstructure is characterised by a marked grain size contrast between the coarser grains and the matrix and has been called "duplex structure" (Beck, 1954). Very few detailed descriptions exist in the geological literature of secondary recrystallization (Carstens, 1966; Wenk, 1965b, 1966), but it is commonly described in many of the glaciological works on microfabric e.g. Shumskii (1964) and Rigsby (1968).

It is on the basis of the above information that the microfabric of the Mount Isa metamorphic sequence is interpreted. A necessary prerequisite then are the ideas and theories formulated in the voluminous metallurgical literature. As the complexities of the deformational processes under geological conditions must be enormous, a geologist may, then, have difficulty in applying all the knowledge obtained through experimental work, or through an extensive study of the metallurgical literature. But a detailed knowledge of the underlying metallurgical principles
will not force him into making assumptions with little justification or questionable or inclusive evidence as has been done in the past.

This study then, is intended to illustrate any microfabric developments, and their interpretation, that may occur with different conditions of deformation and temperature in a quartzite sequence. To the knowledge of the present author this is the first attempt to establish such a relationship in a prograde regional metamorphic sequence. Although many authors have recognised different orientation patterns at varying grades of metamorphism, many of which recur repeatedly in different tectonic and metamorphic environments, none have been related to differing physical conditions that may exist during the deformation and metamorphism through a comparison of microstructure.
CHAPTER 2
STRATIGRAPHY AND GENERAL GEOLOGY

1. **Introduction**

The majority of the rocks discussed in this thesis outcrop to the west of Mount Isa in north-west Queensland, Australia, at latitude 20°43' S and longitude 139°30' E (Figure 2.1). A mine was established here by Mount Isa Mines Ltd. (MIM) in 1931 to extract silver, lead, zinc and subsequently copper, and has since become the largest mine in Australia. But this study has avoided any of the local geological problems associated with the mine and has concentrated on an area to the west and south-west of the mine. However, as this is the first detailed structural study to have been undertaken in the Mount Isa District many of the problems have only been resolved by using the detailed knowledge and unpublished results gathered in the Mine area and adjacent areas by geologists from MIM Ltd. Although the author cannot be completely in accord with ideas held by MIM geologists their work has served as an excellent basis for the present work. Two recent general accounts of the geology and mining methods
Figure 2.1

Limits of outcropping Precambrian and area of Map 1
at Mount Isa have been published by Bennett (1965) and Davies (1967). A modified version of the map of the Mount Isa district (Battey, 1962), which was included with Bennett's paper is also included in this thesis as Map 2.

Mapping was carried out on three sets of aerial photographs, Mount Isa 1956, Mount Isa 1960 and Big Toby Creek 1962 at scales of 1:9000 and 1:4500, 1:9600 and 1:10000 respectively. A suitable base map was drawn at a scale of 1:9000 by means of a radial line plot, and all information was compiled at this scale. Good control of the scale exists north from 28000 S and east of 15000 W, the rest of the map is uncontrolled. The grid co-ordinates are the same as Mount Isa Mines grid and the Australian National Grid. The four figure numbers refer to specimens and sections in the collection of the Department of Geophysics and Geochemistry, The Australian National University.

The terms "psammite", "psammopelite" and "pelite" are used in a compositional sense throughout this thesis for metamorphosed sandstones, argillaceous sandstone, and argillite. There is therefore a gradation from a quartz rich rock to a mica rich rock. The term "quartzite" has been reserved for almost pure siliceous
rocks with $\geq 90\%$ quartz, a usage similar to that of Skolnick (1965) for both sedimentary and metamorphic quartzites.

The majority of this chapter is devoted to a detailed description of the area west of the Mount Isa Fault and its relationship to the previously established stratigraphy. The detailed stratigraphy which follows was found necessary in order to establish the macroscopic structure which will be discussed in Chapter 4.

2. Stratigraphic Basis for Study

(i) Stratigraphy of the area east of the Mount Isa Fault

The first significant stratigraphic study of the Mount Isa area was published by Carter, Brooks and Walker (1961). They established the stratigraphic relationships between the rock units of the Mount Isa district and those in other parts of the Pre-cambrian shield in north western Queensland. Bennett (1965) summarised previous work by Murray (1961), Hiedecker (1961), Cordwell, Wilson and Lord (1963), Powell (1963) and O'Dea (1964), in areas both to the east (Fig. 2.2a) and west of Mount Isa (see Fig. 2.2b) and established subdivisions within the Mount Isa group sediments.
NOTE:
Detailed plane table mapping in the vicinity of the Mount Isa Mine, by officers of M.I.M. Ltd., also exists but this is not shown.

Figure 2.2a
Recently Robinson (1968), after regional mapping in the Eastern Creek Volcanics, established four new formations which are shown in Table 2.1. Good descriptions of the rocks in the Mount Isa sequence are non existent, except for the very detailed work of Croxford (1962) in the Urquhardt slates* of the Mount Isa Group, and the more recent unpublished work of Van den Heuvel (1969). From the few published descriptions, the sequence is apparently characterised by quite rapid changes in thickness and lithology of individual units, especially laterally. The sequence has been subdivided (Carter et al., 1961) into several units:

- Mount Isa Group 3,000m +
- Myally Beds 6,000 to 9,000m (The Judenan Beds are their Western equivalents)
- Eastern Creek Volcanics 6,100m +
- Mount Guide Quartzite 2,450m +
- Leichhardt Metamorphics
- Yaringa Metamorphics

*Murray (1961, p.130) recognised the metamorphic nature of the Mount Isa Group and pointed out that the Urquhardt Shales should be called slates. But he and later workers retained the name shales because of its accepted usage and because bedding and cleavage were commonly parallel. Slate is used here to describe the rock.
<table>
<thead>
<tr>
<th>Mount Isa Group</th>
<th>Sequence west of the Mount Isa Fault</th>
</tr>
</thead>
<tbody>
<tr>
<td>Magazine Formation</td>
<td>210 m</td>
</tr>
<tr>
<td>Kennedy Formation</td>
<td>310 m</td>
</tr>
<tr>
<td>Spear Formation</td>
<td>170 m</td>
</tr>
<tr>
<td>Urquhart Formation</td>
<td>910 m</td>
</tr>
<tr>
<td>Native Bee Formation</td>
<td>800 m</td>
</tr>
<tr>
<td>Breakaway Formation</td>
<td>1040 m</td>
</tr>
<tr>
<td>Moondarra Formation</td>
<td>1220 m+</td>
</tr>
<tr>
<td>Phyllite and Quartz-muscovite-carbonaceous Schist</td>
<td>120 m+</td>
</tr>
<tr>
<td>Quartz-muscovite Schist</td>
<td>330 m+</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Myally Beds</th>
<th>Sequence west of the Mount Isa Fault</th>
</tr>
</thead>
<tbody>
<tr>
<td>Quartzite marker</td>
<td>6000-9000</td>
</tr>
<tr>
<td>Myally Beds</td>
<td>6000-9000</td>
</tr>
<tr>
<td>Foliated Quartzite</td>
<td>330 m+</td>
</tr>
<tr>
<td>Quartzite</td>
<td>390 m+</td>
</tr>
<tr>
<td>Muscovite-(biotite) Schist</td>
<td>450 m</td>
</tr>
<tr>
<td>Quartz-muscovite Schist</td>
<td>120 m</td>
</tr>
<tr>
<td>Quartzite</td>
<td>60 m</td>
</tr>
<tr>
<td>Quartz-muscovite Schist</td>
<td>90 m</td>
</tr>
<tr>
<td>Quartzite</td>
<td>60 m</td>
</tr>
<tr>
<td>Quartz-chlorite Schist</td>
<td>60 m+</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Eastern Creek Volcanics</th>
<th>Mount Guide Quartzite 2450 m+</th>
</tr>
</thead>
<tbody>
<tr>
<td>Pickwick Beds</td>
<td>760 m</td>
</tr>
<tr>
<td>Lena Quartzite</td>
<td>610 m</td>
</tr>
<tr>
<td>Paroo Beds</td>
<td>1830 m</td>
</tr>
<tr>
<td>Cromwell Beds</td>
<td>2140 m</td>
</tr>
<tr>
<td>Amphibolite</td>
<td>220 m</td>
</tr>
<tr>
<td>Schists</td>
<td>200 m</td>
</tr>
<tr>
<td>Amphibolite</td>
<td>370 m</td>
</tr>
<tr>
<td>Schists</td>
<td>420 m</td>
</tr>
<tr>
<td>Amphibolite</td>
<td>610 m</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Mount Guide Quartzite 2450 m+</th>
<th>May Downs gneiss</th>
</tr>
</thead>
<tbody>
<tr>
<td>Mount Guide Quartzite</td>
<td>1000 m</td>
</tr>
<tr>
<td>May Downs gneiss</td>
<td>1000 m</td>
</tr>
</tbody>
</table>
The thicknesses quoted are those determined by previous workers in the type sections. It is probable that these thicknesses are not true thicknesses as no account of structural duplication existing in the type sections has been considered. Therefore these figures are liable to give an inaccurate picture of the correct thickness (c.f. Cloos, 1947).

The Yaringa Metamorphics consist of an isolated outcrop of gneiss and mica schist with some quartzite, conglomerates and migmatites. They represent the oldest rocks found in the Mount Isa District, occurring 30 kilometres to the west of Mount Isa, and have been labelled as possible Archean basement by Carter et al. (1961). The Leichhardt Metamorphics are found between Mount Isa and Cloncurry and consist of moderately metamorphosed pelitic and psammitic rocks with some metamorphosed acid lavas. The overlying Mount Guide Quartzite consists of quartzite, arkose and conglomerate. The thickness of this unit is extremely variable and Carter et al. (1961) believe that the lower boundary is probably disconformable. The Eastern Creek Volcanics are composed of interbedded metabasalts and metasediments and contain minor sills and stocks of gabbro. The sequence has been subdivided into four
units by Robinson (1968) (See Table 2.1). The Myally Beds have a variable thickness and apparently overlie the Eastern Creek Volcanics conformably in some places and disconformably in others. They consist of quartzite, siltstone and shale with variable thicknesses of conglomerate. A district quartzite member is very prominent in some upper portions of the section. The Judenan Beds are believed to be equivalent to the Myally Beds, and occur west of the Mount Isa Fault. The Mount Isa Group consists of a conformable sequence of low grade regionally metamorphosed shales, siltstones and bedded carbonates. The sequence has been subdivided into seven formations by Bennett (1965) (See Table 2.1).

(ii) Stratigraphy of the area west of the Mount Isa Fault

The rocks of the area to the west of the Mount Isa Fault are very similar to their eastern equivalents. Quartzites and quartz mica schists predominate with metamorphosed basic rocks prominent near the base.

The area to the west of the Mount Isa Fault is cut by three major strike faults. The Mount Isa Fault occurs in the east and has corresponding parallel faults, Faults I, II and Fault III, occurring successively westwards. (See Figure 2.3). These divide the area into four zones, and stratigraphic correlation between
LOCATION OF THE MAJOR FAULTS IN THE AREA WEST OF MOUNT ISA

THE MOUNT ISA AND FAULTS 1 to 3 ARE STRIKE FAULTS

FAULTS a to h ARE PROMINENT CROSS FAULTS

Figure 2-3
these is not possible except by arguments based on the
general nature of the rocks, their position in the
local stratigraphic columns and analogy with the
stratigraphic column established in the east of the
Mount Isa Fault.

The succession in descending order, but not
geographical position is:

---

**Rocks between the Mount Isa Fault and Fault I**

<table>
<thead>
<tr>
<th>Depth</th>
<th>Description</th>
</tr>
</thead>
<tbody>
<tr>
<td>120m</td>
<td>Black carbonaceous slates and phyllites with minor quartz-muscovite schists.</td>
</tr>
<tr>
<td>330m</td>
<td>Quartz-muscovite schist</td>
</tr>
</tbody>
</table>

**Foliated Quartzite**

<table>
<thead>
<tr>
<th>Depth</th>
<th>Description</th>
</tr>
</thead>
<tbody>
<tr>
<td>330m</td>
<td>Well bedded quartzites with minor lenticular conglomerate towards the base.</td>
</tr>
</tbody>
</table>

**Rocks between Fault II and Fault III**

<table>
<thead>
<tr>
<th>Depth</th>
<th>Description</th>
</tr>
</thead>
<tbody>
<tr>
<td>840m</td>
<td>Quartzite and quartz-muscovite (biotite) schist.</td>
</tr>
</tbody>
</table>

**Rocks between Fault I and Fault II**

<table>
<thead>
<tr>
<th>Depth</th>
<th>Description</th>
</tr>
</thead>
<tbody>
<tr>
<td>390m</td>
<td>Quartzite, quartz-muscovite, quartz-chlorite schist with minor actinolite schists.</td>
</tr>
</tbody>
</table>

**Rocks west of Fault III**

<table>
<thead>
<tr>
<th>Location</th>
<th>Depth</th>
<th>Description</th>
</tr>
</thead>
<tbody>
<tr>
<td>Eastern Creek</td>
<td>1820m</td>
<td>Metabasalts (amphibolites), hornblende-plagioclase schists, quartz-chlorite schist, quartzite, quartz-mica schists, quartzite, cordierite schists and cordierite-entophyllite schists.</td>
</tr>
<tr>
<td>Mount Guide</td>
<td>1200m</td>
<td>Quartzite with minor schist and gneiss</td>
</tr>
<tr>
<td>May Downs Gneiss</td>
<td></td>
<td>Gneiss, schist and quartzite</td>
</tr>
</tbody>
</table>
Intruding this stratigraphic sequence is the Sybella Granite, with associated pegmatitic and aplitic dykes, and a number of basic intrusions.

**The May Downs gneiss**

The May Downs Gneiss is composed of what appear to be the oldest rocks in the area consisting of an alternating assemblage of foliated grey quartz-microcline-muscovite-(biotite) gneiss, pink and white coloured microcline gneiss, quartz-muscovite-biotite schists and quartzite. The outcrop is generally poor and the unit occurs on the western margin of the area. (See Map 1).

The gneisses have an average grain size of approximately 0.3 mm. and are composed of varying proportions of microcline quartz, plagioclase, biotite, muscovite and, rarely, sillimanite. They are generally equigranular rocks consisting of polygonal shaped grains of quartz, microcline and plagioclase with planar to smoothly curved interfaces and with micas of similar dimensions sitting in grain boundaries. However, in many of the more microcline rich rocks large muscovite mica blades up to 3 mm. across are occasionally observed. The predominant rocks are quartz-microcline and the quartz-biotite gneisses, the former containing up to 50% microcline, 30% quartz, and
10% biotite and muscovite mica, whereas the latter contain approximately 80% quartz and 15% biotite with minor amounts of microcline and plagioclase. The mica schists interbedded with these gneisses occur as rather thin layers of up to 20 cm. thick and are composed of muscovite, biotite and quartz with minor quantities of plagioclase and microcline. They are extremely well foliated. The quartzites generally have a higher mica and feldspar content than those occurring in the overlying units but have a similar equant grain shape with polygonal grain boundaries between which are numerous mica inclusions.

The presence of these gneisses and schists were noted by Joplin (1955 p. 39) and Brooks and Shipway (1960) but have appeared on all published maps as Sybella Granite. Carter et al. (1961, p. 60) recognised the presence of these rocks and pointed out that they may be Archaean basement similar to the Yaringa Metamorphics.

The Yaringa Metamorphics have been assigned to the Archean on a basis of their apparently higher metamorphic grade and a marked variation in strike from the regional north-south trend of most of the Mount Isa-Cloncurry area. The basis of this age assignment is unjustified until further detailed structural and
stratigraphic work is carried out in the area. From field observations, the present author believes that there is a strong mineralogical similarity between these rocks and the Yaringa Metamorphics but no field relationship could be established. No obvious similarity was found between these and the Leichhardt Metamorphics, the latter being very much richer in plagioclase.

**Mount Guide Quartzite**

This unit was first mapped to the west of the Mount Isa Fault by Blanchard and Hall (1942, p. 37, Fig. 5) and later by Brooks and Shipway (1960, Fig. 1). The unit consists of interbedded massive coarse grained quartzite with subordinate quartz-mica schist. Bedding is the only distinctive sedimentary structure. Bedding thicknesses vary from 1 cm. to 8 m. and junctions between adjacent beds, especially the thicker units, are impossible to recognise on the ground. Individual dips on such junctions are unobtainable and only the trend of the units may be observed on the aerial photographs.

The more massive quartzite which makes up the greater proportion of the unit is very pure with only minor quantities of microcline and muscovite mica. These rocks are very coarse recrystallized quartzites with grains varying from 0.2 mm. to 3 cm., this grain
size being partly controlled by impurity content. (See Chapter 5). Interbedded with these are very minor quartz-muscovite-biotite schists and occasional mappable gneissic layers. The gneisses are either quartz-
muscovite-microcline gneisses with or without plagioclase. They are generally well foliated with a layering of approximately 2 mm. thick consisting of alternating quartz feldspar versus mica rich layers. They are very similar to the gneisses found in the May Downs gneiss. Some of the interbedded gneisses have been very intensely folded, especially in the vicinity of grid reference 17250 W 9700 S with many second generation folds having variable plunges and axial plane directions. In such areas these gneisses resemble migmatites, for they are also intensely veined with coarse quartz microcline rich layers which follow the compositional layering in the rock.

The upper boundary of this unit appears to be conformable with the Eastern Creek volcanics, the dominant rock type on the junction being microcline rich quartzites. The boundary strikes almost north-south, but there is a gentle swing to the west in the northern portion of the area. The lower boundary is quite variable in strike and has been subjected to both first and second generation folding. Internally the whole
unit has been extensively folded and many large first and second generation structures have been mapped.

The most prominent of the first generation structures is a refolded anticline north of Lena Creek. Second generation folds are found occurring west of Beryl North and as a dome shaped structure near Yaringa Creek. The junction between this unit and the Sybella Granite in the vicinity of the large dome structure is discordant, with mappable units being truncated by the Sybella granite. Irregular basic intrusions are found to the south of the Crystal Workings and also as a series of sills within this unit.

The Mount Guide Quartzites to the west of Mount Isa differ from the type section described by Carter et al. (1961) in that there is a complete lack of any basal conglomerate or any interbedded extrusives, although what they have called metabasalts may equally be basic sills. In their map of the Mount Isa area Carter et al. (1961) include this unit in the Eastern Creek volcanics (p. 66) but point out "that it may be Mount Guide Quartzite because it underlies a succession of metabasalts". They also suggest that it may be a quartzite lens within the Eastern Creek volcanics or an Archaen unit. Bennett (1965) and the present author on the basis of lithological similarities and
stratigraphic position believe that this unit is the metamorphic equivalent of the Mount Guide Quartzites east of the Mount Isa Fault. There is no evidence for this unit being Archaean.

The Eastern Creek Volcanics

The Eastern Creek Volcanics consist of interbedded metamorphosed lavas, pyroclastics and sedimentary rocks, which each form layers from as little as 5 cm. up to some hundreds of metres in thickness. The metasediments and pyroclastics are generally fine grained rocks in which the original layering is usually very well preserved. The layering which varies from a few millimetres up to 50 cm., is very regular, and is due to slight differences in grain size and composition, and is accentuated by marked colour variation from white to pink to green and black, depending on whether the dominant mineral is muscovite, chlorite, microcline, epidote or hornblende.

These volcanics occur as a north-south trending belt the western boundary of which, and also the strike of the very prominent foliation, taking a westerly swing in the northern portion of the area. This western, or lower boundary appears to be conformable with the underlying Mount Guide quartzite within the
area mapped. The Eastern boundary adjoins the Judenan Beds along the Fault III. From approximately 8000 S to 16000 N there is a thin zone of quartz-
muscovite schist separating the amphibolites from the quartzites and the approximate position of the fault is thought to lie within or near this unit which has not been shown on the map.

This unit was first mapped by Blanchard and Hall (1942) and subsequently by Brooks and Shipway (1960), Carter et al. (1961) and O'Dea (1964). No subdivisions were made of these metabasalts and metasediments and the rocks have been briefly described by Joplin (1955), O'Dea (1964) and to the north of the area at Slaughter Yard Creek by Doust (1966) (See Figure 2.2b)

There is quite a marked difference in metamorphism across the Mount Isa area. Most of the rocks in the type section of the Eastern Creek Volcanics, (Carter et al., 1961; Robinson, 1968) consist of low grade regional metamorphics. Quartz, albite, epidote, prehnite, chlorite, carbonate and sphene occur as authigenic minerals. Whereas the rocks to the west of the Fault are higher grade and may contain cordierite, anthophyllite, andalusite, and hornblende.
The sequence found to the west of the Mount Isa Fault is typically developed from grid reference 11380 W 12180 N going eastwards perpendicular to the strike of the rocks to grid reference 5700 W 15960 N. The rocks found in this traverse are typical of the units further south. In this type section, the Eastern Creek Volcanics may be subdivided into three quite distinct units, two are amphibolite units and the other is a schist. From 8000 S to 20000 S it is possible to recognise an additional amphibolite and schist unit, but south of this the relationships between the units are obscured by the intrusion of the Sybella Granite and its associated pegmatites. Along the eastern margin of the Sybella Granite there is only one schist and one amphibolite member.

In the type section the lowest, most western unit consists of a well foliated and strongly lineated amphibolite with minor quartz-biotite-schist, quartzite and carbonate rich layers. Overlying this with a gradational boundary are quartz-muscovite-biotite, biotite-hornblende and muscovite-cordierite schists with very minor quartz-epidote and quartzite layers. The upper unit consists of amphibolite which lacks a strong foliation, although it is extremely well lineated, and some outcrops are
composed of a series of very fine pencils. Within this unit are also some quite minor marble layers.

Within the Eastern Creek Volcanics, especially in those west of the Mount Isa Fault are rocks which may have originated as lava flows; although they form layers up to several hundred metres in thickness concordant with the bedding of the metasediments, no unequivocal evidence has been found to show that all of these rocks are of extrusive origin and it is conceivable that they were emplaced as sills. However no discordant boundary between these rocks and the metasediments has been seen, and on the basis of the evidence available. An extrusive origin seems to be more likely. Many of the metabasalts include layers containing numerous deformed amygdales.

The metabasalts, which are the predominant rock type, are now hornblende-plagioclase-quartz amphibolites in which there is a poorly defined foliation. In the field this foliation appears as very fine light and dark layers. This is a result of a variation in both the microstructure and the relative proportions of the minerals, with an alternation of hornblende-plagioclase layers versus quartz-plagioclase rich and hornblende poor layers. The layering varies in width from 0.5 mm. to 1.5 mm.
In the amygdaloidal basalts a lineation may be observed as an elongation of the amygdales, which are infilled with quartz and calcite, with the longest axes of the deformed amygdales parallel to the mineral lineation in the deformed amphibolite or adjacent metasediment, and also parallel to the first generation fold axes in the area.

Epidote rich nodules and veins commonly occur both in the metasediments and metabasalts along joint cracks. These joints are often found either crossing the foliation and or lensing out within the foliation. Microcline and quartz-plagioclase rich veins are also common in many of these rocks and are very similar to the distribution of veins described in an altered basic lava from the central west of New South Wales by Smith (1968).

It has been firmly stated in all the literature concerned with these rocks that they may be correlated with the Eastern Creek Volcanics to the east. The major difference being that they belong to the amphibolite facies (See Section 7) unlike those further east which are placed in the lower greenschist facies (Robinson, 1968). A number of authors, Brooks and Shipway (1960), Carter (1959), Carter et al. (1961), Joplin (1955), O'Dea (1964)
and Walker (1958) have also ascribed their higher grade of metamorphism to a thermal effect from the Sybella Granite rather than to their position in a regional metamorphic sequence.

Robinson (1968) has subdivided the Eastern Creek Volcanics, to the east of Mount Isa into four formations (See Table 2.1), but no attempt has been made to correlate the units established to the west of the Mount Isa Fault with those of Robinson's. The major differences between the two sequences are that the basal beds to the west are dominated by pyroclastics with some labile sediments and contain very little obvious extrusives, while Robinson's Cromwell Beds are composed almost exclusively of extrusives. The schist unit in the west is devoid of any obvious metabasalt which dominate the Paroo Beds. This unit to the west is then overlain by massive amphibolite the greater proportion of which appears to have once been extrusive. The Lena Quartzite and Pickwick Beds are also absent, although they may exist in the sequence but have been faulted out of the area west of Mount Isa.

The Judenan Beds

This name was applied by Carter et al. (1961, p. 100) to include all rocks between the Mount Isa
Fault and the Western equivalents of the Eastern Creek Volcanics. They were equated with the Myally Beds in the east. The Judenan Beds have been strongly faulted and folded and a complete sequence does not exist in the area due west of Mount Isa. The Judenan Beds are composed of quartzite and quartz mica schists with subordinate chlorite and actinolite schists. They strike almost north-south throughout most of the area, except for a slight swing in strike between Lena and Mica Creeks around the northern end of the Sybella Granite.

Bedding/cleavage intersections, fold vergences and sedimentary structures indicate that nearly all units are the right way up. They are not extensively overturned as believed by other authors, such as, Brooks and Shipway (1960), Zimmerman (1961), Carter et al. (1961) and Bennett (1965). Heidecker (1961, p. 16) was the first to recognise that facing criteria indicated that these beds were the right way up. However his final analysis of the structure required bedding to be overturned and dip west (see Heidecker Fig. 6, p. 87). He believed that these facings were the result of a series of wedge-like blocks between numerous strike faults parallel to the cleavage.
The Judenan Beds have been divided into three separate blocks by the strike faults I, II and III (Fig. 2.3). These blocks will be discussed individually as they represent quite different rock types, and direct correlation between the three is not possible. The overall dip of the rocks in the eastern block is easterly although most local dips are to the west because of intense isoclinal folding of the units parallel to the axial plane of the folds. The rocks in the other two blocks dip in a westerly direction, with the most westerly block apparently underlying the rocks between Faults I and II. On stratigraphic arguments an attempt is made to correlate these units with those existing in the east and it is believed that the easterly block (between the Mount Isa Fault and Fault I) contains the youngest rocks found west of the Mount Isa Fault, and the rocks found in the central block are the oldest.

Rocks between the Mount Isa Fault and Fault I

These two faults extend the whole length of the area mapped as two parallel linaments between which are sandwiched three units. Extremely good exposure of all three units may be observed in any east-west traverse between King Gully and south of Lena Creek. A well bedded quartzite unit is the most conspicuous
rock type in this zone and is overlain to the east by a quartz-muscovite schist and a thinner quartz-muscovite-carbonaceous schist and phyllite unit. All three formations have been truncated, at different localities, along their eastern boundary by the Mount Isa Fault, as shown on Map 1.

(a) The Foliated Quartzite

The foliated quartzite attains a maximum thickness of approximately 330 metres between Mica and Lena Creeks but further north thins to some 120 metres. To the south of Mica Creek at 650 W 41380 S the unit is truncated by the Mount Isa Fault as it changes strike from a north-south direction to a north-west direction in a second generation fold closure. The unit consists of well bedded white quartzite and minor schists within which are well preserved sedimentary structures. These include current bedding, lode casts, ripple marks, mud cracks and flame structures (See Plate 2.1). Bedding varies from 5 cm. to 1 m. in thickness and appears to dip in a westerly direction at a fairly high angle. Facings within these beds are observed to change suddenly in some localities but no fold closures are evident. Junctions between this and the overlying unit are frequently observed to have quite different dips to
(a) Current bedding typical of the thicker beds of quartzite in the foliated quartzite unit at 480 W 2900 S.

(b) Fossil mudcracks on the surface of a bedding plane at 3050 W 16620 S. Ruler, to centre of hinge, is 6" long (15 cm.)
the mesoscopic* bedding seen in outcrop. A good example of this may be observed by looking south from 900 W 16000 N to the southern side of King Gully, the junction dips east in contrast to the westerly dipping mesoscopic bedding. This is an extremely good example of transposition and will be discussed in Chapter 3.

In the lower part of this unit a number of lenticular conglomerates occur which vary from 2 m. to 10 m. in thickness and have strike lengths up to 2 km. (See Map 1). The pebbles in the conglomerate consist of massive white and grey quartzite varying from 5 mm. to 18 cm. across (See Plate 2.2).

Within this unit there is a very common structure which resembles current bedding. This occurs as a series of finely spaced joints varying from 3 mm. to 1 cm. apart, which may be stained with secondary limonite. They frequently lie oblique to the bedding, in much the same way as the current bedding, but may not terminate within a unit, and, instead, cut across the bedding to terminate in an adjacent unit. This

*The terms macroscopic, mesoscopic and microscopic will be used in this thesis in the sense of Turner and Weiss (1963).
PLATE 2.2

From approximately 3050 W 166000 S at the base of the foliated quartzite unit

(a) Interbedded psammitic and pelite layers with small flame structures separating load casts in the psammites at the psammitic–pelite junction. Intruding the pelite layers are also small thin dykes of psammitic.

(b) Irregularly shaped conglomerate pebbles.
is considered to be a type of fracture cleavage. In the more micaceous quartzites (≥5% ≤20%) this structure occurs as a series of parallel zones which resemble a mylonite.

The quartzites consist of recrystallized and strongly deformed detrital quartz approximately 0.1 to 0.2 mm. in grain size, and may have a muscovite mica content varying from much less than 2% to rarely greater than 10% by volume. Other accessory minerals found in these rocks include zircon, apatite, tourmaline, and some secondary limonite. The rocks generally have a strong planar fabric with the quartz grains tending to be flattened in a plane which is parallel to both bedding and a slaty cleavage. The slaty cleavage is only evident in the more micaceous rocks.

Interbedded with the quartzite are minor pelitic beds now composed of quartz-muscovite schist which may have a strongly developed slaty cleavage. In several localities boudins consisting of prolate ellipsoidal masses of quartz-biotite schist have been observed lying in the slaty cleavage. These boudins are invariably oriented with their major and intermediate axes lying in the slaty cleavage and their minor axis
is also parallel to a faint mineral lineation defined by mica blades lying on the cleavage surface.

(b) The quartz–muscovite schist unit

This overlies the foliated quartzite to the east and lenses out against the Mount Isa Fault in the area south of Mica Creek. It extends northwards to reach a maximum thickness of 330 metres in the vicinity of 22000 S, and 80 metres in the King Gully area.

The unit is composed of light brown quartz–muscovite–(biotite) schist and minor darker grey quartzites. These rocks have a grain size of approximately 0.1 mm. and generally consist of quartz–muscovite schists with minor quartz–muscovite–biotite or quartz–biotite schists. They have an extremely strong preferred orientation of the (001) cleavage in the micas, with the quartz occurring between the micas as small polygonal grains. Some of these pelites also exhibit a tendency to form a metamorphic layering consisting of muscovite mica layer versus a muscovite quartz rich layer.

Interbedded with this unit are minor quartzites which occur as thin layers, varying from 3 to 20 cm. thick, and are generally lenticular. The thicker beds may be
continuous along the strike for distances up to 50 m. while the thinner layers may only be continuous for 6 m. or less. The bedding is very well preserved in this unit and is generally transposed parallel to a strong slaty cleavage which is extremely well developed in the pelites. Most quartzites have been completely recrystallized and now consist of small polygonal quartz grains or larger quartz grains flat and parallel to the slaty cleavage. In some of the thicker beds detrital quartz grains appear to be preserved along with detrital micas and are set in a matrix of finer grained recrystallized quartz and mica.

(c) The quartz-muscovite carbonaceous schists and phyllites

This is a grey coloured unit which overlies the quartz muscovite schist and has previously been included in what has been known as the "Mount Isa Shear Zone" (Powell, 1963). It is a very thin (120 metres) continuous unit which can be mapped from King Gully in the north to the White blow at approximately 23000 S, where it is truncated by the Mount Isa Fault. The unit is strongly folded with many second generation folds of variable plunge trending roughly in a north-south direction. Most vergence
observations in this unit indicate an easterly direction of facing (See Chapter 4).

The unit consists predominantly of a light grey to black phyllitic rock with minor light coloured siliceous layers. Bedding is very well preserved varying in thickness from less than 1 cm. to 1 metre. The phyllites and schists are composed of fine grained, approximately 0.03 mm., carbonaceous material, muscovite mica, biotite mica and quartz with very minor limonite, zircon and tourmaline. The variation in colour between the phyllites is caused by the abundance of quartz, muscovite mica and carbonaceous material. An abundance of the carbonaceous material also appears to inhibit grain growth of the micas. In such rocks biotite is generally completely absent and the muscovite is finer grained than in the non-carbonaceous rocks which are generally schists rather than phyllites. There is an extremely strong preferred orientation of the micas in all rocks and the quartz occurs as elongated single grains or lenticular aggregates in the slaty cleavage.

The quartzites occur as very fine beds, with bedding no more than 15 cm. in thickness, but generally only 1 cm. thick, and are composed almost
entirely of fine grained interlocking quartz grains, with minor quantities of muscovite mica.

**Rocks between Faults I and II**

Both Faults I and II are strike faults striking the whole length of the area, but there is no structural evidence for Fault II continuing south of Mica Creek. The rocks between these two faults may be divided into three blocks which are separated by the east-west faults e and f (See Fig. 2.3). The best representative area for the sequence between these faults is in the vicinity of the Tailings Dams. The other two blocks contain only the higher members exposed in the Tailings Dam area.

The rocks in the Tailings Dam area represent only a small portion of the Judenan Beds and total approximately 390 metres of quartzite, quartz mica, quartz-chlorite and actinolite schists. The lowest unit is a well foliated quartz-chlorite schist which abuts against the foliated quartzite in the east and is overlain by quartzite in the west. The unit also contains minor layers of quartz-muscovite schist and numerous thin layers rich in actinolite and some larger lenses of coarsely crystaline actinolite and fine grained quartzite. These larger lenses have been mapped by Heidecker (1961) but are not shown in the
present work. Chlorite schists make up the bulk of this unit and vary from pure chlorite schists to rocks having as much as 50% quartz. Beds vary from 1 cm. to 1 m. with very fine laminations in many of the more quartz rich material. Ripple marks and cross bedding are also abundant in the quartz rich material (See Plate 2.3). The light green actinolite rich layers are very lenticular varying from 2 cm. to 30 cm. and are sometimes interbedded with a purple coloured quartz-magnetite schist. The larger lenses of actinolite lie parallel to the strike of the major units and vary from 10 to 120 metres and consist of coarse grained actinolite, up to 1 cm. across, but averaging 4 mm., and minor magnetite and quartz. The lenses of quartzite have similar dimensions but there appears to be no direct relationship between the different rock types. The quartzite is extremely fine grained, the average grain size varies from 0.05 to less than 0.02 mm. and the rock is composed of quartz (60%), biotite (20%), magnetite (15%) and minor plagioclase and tremolite.

The overlying massive poorly bedded quartzites outcrop in a north-north-west trending belt which is truncated against the Fault I west of Kennedy Creek and by Fault II in the west. It also outcrops as a thin slither against Fault I from 7000 S to Lena
PLATE 2.3

From north of King Gully within the quartz-chlorite unit at 900 W 15950 N

(a) Coarse grained actinolite rich layers interbedded with fine grained quartz-magnetite layers. Scale on ruler is inches.

(b) Small symmetrical, transverse ripple marks. The long edge of the photograph is approximately 2 metres.
Creek. There are no obvious sedimentary structures in these rocks and bedding thickness varies from 30 cm. to 2 metres. The unit is highly jointed and it is often difficult to distinguish between joints and bedding surfaces, for compositional differences between different bedded units are insignificant and it is only very infrequently that there may be a thin veneer of pelitic material. The unit has been gently folded, generally as a series of large amplitude second generation folds, a good example of this may be seen in the area to the west of Kennedy Creek.

The quartz mica schist unit is composed of finely bedded light brown quartz-muscovite schist, which has a distinct sandy appearance in outcrop, minor muscovite schist and quartzite. Bedding within these rocks varies from less than 5 mm. to greater than 2 metres in thickness. The unit extends from adjacent to Fault I at 6000 S through No. 5E Tailings Dam to be truncated by the Fault II at approximately 8370 N. The same unit is also found in the two southern blocks. In the Lena Creek area the western boundary is obscured by a thick cover of alluvium and "bull dust", but widens quite considerably in the core of a faulted synclinal structure. To the south of 2000 N to Mica Creek the mapped outcrop of this unit is lenticular
with the eastern boundary being truncated by Fault II. The western boundary in this southern block consists of an irregularly shaped mass of light green chlorite schist with amphibolite pods. This chlorite schist is a retrogressed basic intrusion which also intrudes this same unit further north. Along the eastern boundary in this southern block these schists appear to have been extensively altered by contact metamorphic effects from the intrusion. Mica and chlorite form large decussate radiating aggregates and small irregular knots consisting of quartz, chlorite and muscovite. The strong slaty cleavage which characterises the unit elsewhere has been completely obliterated along this margin.

The rocks within this unit are generally dominated by a very strong slaty cleavage ($S_1$) which in places has been overprinted by a crenulation cleavage ($S_2$).

The overlying quartzite unit also consists of a massive white quartzite with more abundant evidence of bedding than the underlying quartzite and is composed of 1 metre thick beds of white quartzite devoid of sedimentary structures except for some good occurrences of ripple marks at grid reference 2500 W 2500 S. Interbedded with this unit are thin layers of quartz-muscovite schist and occasional
lenses of quartz-chlorite and chlorite schist, these latter rock types probably represent retrogressed intrusives. Bedding is generally very hard to positively identify and may be confused with closely spaced jointing. Individual beds vary in thickness from 5 cm. to greater than 2 metres.

This unit outcrops from grid reference 1200 W 12000 N to the Fault e where it thickens quite considerably in the nose of a faulted syncline. It outcrops again in the southern block extending from Fault f to south of Mica Creek.

The overlying quartz-muscovite schist is the highest portion of the sequence exposed between Faults I and II and is not complete, as the upper portion has been faulted out. It outcrops from Fault e northwards to approximately 7100 N as a north-north-west trending belt where it is truncated by the Fault II. The unit consists of interbedded quartz-muscovite schist, muscovite schist and minor quartzite. It is generally finely bedded with individual units being no more than 1 metre in thickness.

**Rocks between Faults II and III**

Two units have been recognised: the upper and the most westerly Judenan unit is a quartzite, while the lower consists of mica schists. The schists
extend from the extreme southern end of the mapped area increasing in outcrop width in the vicinity of Fault f, but thin out again in the north. Quartz-muscovite schists are the dominant rocks but there are also minor quantities of quartz-muscovite (magnetite) schist, quartz-muscovite-chlorite schist, andalusite schist and occasional layers of quartzite. The whole unit has a strongly developed slaty cleavage which is often folded by folds with an associated crenulation cleavage.

The overlying quartzite outcrops as two distinct portions, a southern belt, which extends south from the Fault h to south of Mica Creek. The northern portion terminates against Fault III at grid reference 9050 W 9500 S and extends north of Lena Creek as a belt of variable width. The unit is composed of massive well bedded quartzite with minor quartz-mica schist. Beds vary from 6 cm. to 6 metres with very few sedimentary features except for some rare current bedding and ripple marks, (e.g. at grid reference 12200 W 23200 S). In most of this unit bedding is unrecognisable and may be confused with jointing.

The quartzites are coarse grained rocks (average grain size 0.2 mm) and contain less than 7 percent mica. Quartz grains meet in triple point junctions (Voll, 1960, p. 526) and the small percentage of
mica has a reasonably strong preferred orientation. Minor micaceous quartzites also occur with a modal percentage of mica varying from 7 to 20%. In these rocks there is generally a good mica preferred orientation, the quartzes still retain their polygonal shapes and there is only a slight tendency for them to be elongate parallel to the foliation $S_1$.

3. Discussion

The sequence described to the west of the Mount Isa Fault probably corresponds to part of the sequence developed to the east. The possible correlation is illustrated in Table 2.1.

Overall there is a broad stratigraphic correlation between the major units with significant differences only existing when units are further sub-divided. These differences may be attributed to an insufficient knowledge of the stratigraphy in the east or they may be a function of rapid changes in the original sedimentary environment. The presence of shallow water sedimentary structures such as mud cracks, ripple marks, current bedding, pull apart and the presence of rapid changes in lithologies and numerous unconformities in the sequence, established by Carter
et al. (1961), appear to support the idea that the depositional basin probably represents shallow water sedimentation. In the past the depositional basin in the Mount Isa area has been referred to as a geosyncline (Carter et al., 1961) the implication being that sedimentation occurred in deep water. In the author's opinion until further detailed work is carried out to establish the sedimentary and structural history over a greater area the term geosyncline should be dropped from its current application to this area, since the present day usage of geosyncline implies the connotation of a particular type of sedimentation (Bouma, 1962). There is no evidence for this type of sedimentation in the Mount Isa region.

A quite significant difference in stratigraphic thicknesses exists between the areas east and west of the Mount Isa Fault. The thickness of the sediments to the west of Mount Isa, excluding the May Downs gneiss, is 4,830 metres in comparison to a thickness of approximately 12 kilometres found in the equivalent units to the east of the Mount Isa Fault. The published figure for the total thickness of all the eastern units is in the excess of 18 kilometres. These differences in thickness may be attributed to insufficient structural information in the defined
type areas, or may represent very major differences in stratigraphic thicknesses within a very small region. The author believes that the former is probably the reason. It would serve to explain, had 8 km+ of sediment existed in the Mount Isa region (see Turner, 1968) why the grade of metamorphism of some of the lower units in the east, viz the Mount Guide Quartzite, is not appreciably higher.

4. Basic Intrusions

Many small lens shaped basic intrusives, the majority of which are amphibolite, have been mapped in an area to the west of the Mount Isa Mine. The distribution of the larger of these intrusions is illustrated in Map 1. These bodies generally occur as small flat lenses varying from 30 to 300 metres in length. Some may be traced as a series of discontinuous lenses for up to a kilometres, while others may occur as circular bodies up to 300 metres in diameter. The junctions between these and the adjacent rocks are generally sharp and the majority of the amphibolite bodies appear to be concordant with the layering in the adjacent metasediments although some appear to be cross-cutting sheets. Indistinct contacts are visible in some amphibolite
bodies, these are surrounded by an irregular zone of chlorite schist which generally possesses a very poorly defined slaty cleavage.

Outcropping in the area also, are a number of other basic bodies of gabbro, associated metagabbros and unmetamorphosed basaltic dykes. The western greenstones (Bennett, 1965) are another belt of deformed basic rocks now consisting of chlorite schist which outcrop discontinuously along the eastern margin of the Mount Isa Fault. Descriptions, and field and chronological relationships between the basic intrusions will be discussed in Appendix A.

5. **Sybella Granite**

The northern nose of a large irregular oval shaped body of Sybella Granite (Joplin, 1955; Joplin and Walker, 1961; Carter et al., 1961) has been mapped as a large lenticular mass terminating at a point approximately 15500 W 24000 S. The longer dimension runs north-south for over 25 kilometres; the maximum width from east to west is 6.5 kilometres. Minor sills and stocks of granite also outcrop within the southwestern portion of the area, in the Mount Guide Quartzite and the May Downs gneiss. The granite is only part of the more extensively outcropping Sybella
Granite, which also outcrops intermitently to the north-west and west of the area. The area of granite outcrop mapped by Carter et al. (1961) also includes many gneisses and quartzite sediments.

The contacts of the major mass of granite are usually sharp and parallel to trends in the neighbouring sediments, except at its northern end where the granite truncates the units. Here elongate xenoliths of amphibolite and quartzite contaminate the granite. Other large "rafts" of sediment are also found along the eastern margin.

The orientation of foliations and lineations within the granite mass are parallel to similar lineations and foliations developed within the country rocks, and variation of foliation and lineations in the country rocks and the granite is sympathetic. Contacts between the smaller intrusions and the country rocks are sharp and definitely intrusive (See Plate 2.4).

Preliminary accounts of the petrology of the Sybella Granite and other granites in the Pre-cambrian shield area of north-west Queensland have been published by Carter et al. (1961), and Joplin (1955, 1961). Doust (1967) has also examined a small
(a) Contact between a small intrusion of Sybella Granite and quartz-microcline gneiss within the May Downs gneiss at 29360 W 37380 S.

(b) Quartzite and smaller amphibolite xenoliths in a strongly foliated and lineated granite at 15500 W 24000 S.
intrusion of Sybella Granite in the Slaughter Yard Creek area.

The Sybella Granite is a quartz-microcline granite and is composed predominantly of two phases; a foliated granite and porphyritic granite. Mineralogically there is very little variation between the two phases. Except in microstructure, both contain large phenocrysts (blastophenocrysts) or aggregates of microcline set in a matrix of microcline (50% to 30% by volume), quartz (40%), plagioclase (20%) biotite (5%), and muscovite mica (2%). Hornblende may also be present in some phases and accessories include, magnetite, ilmenite, apatite, flourite, zircon and sphene.

The foliated granite is a coarse grained rock with a fabric equivalent to the foliation $S_2$, and the mineral lineation in the neighbouring rocks (See Section 6, Chapter 3). The foliation occurs as broad layers and is defined by lenticular biotite aggregates between lenses of quartz and microcline. The outer edge of the main body of granite has a prominent foliation which occurs as a wide zone and extends for many hundreds of metres in width grading imperceptibly into the massive phases. Strongly foliated phases are not restricted to the margins,
but also occur in many of the smaller sill like intrusions of granite. The relationship between the massive and foliated phase appears to be gradational with no marked change in mineralogy and no evidence that one type of granite intrudes another. In places the foliation is weak and although the rocks may show some deformational effects in thin section, they appear massive in outcrop.

The microcline in the lenses or augen like eyes, occurs as large interlocking grains up to 4 mm. across, but where any quantity of biotite is present the grain size is reduced to approximately 1.5 mm. The microcline is generally euhedral, well twinned and adjacent grains meet at triple point junctions with gently curved interfaces, both at microcline-microcline and microcline-quartz interfaces (See Plate 2.5b). There are generally no perthitic lamellae within the microcline, but albite commonly occurs as small exolved blebs or stringers. Quartz inclusions also frequently occur as small circular grains within the microcline and grains are commonly penetrated by coarse areas of myrmekite.

The plagioclase is usually confined to the groundmass where it occurs as irregular grains with curved interfaces meeting at triple point junctions.
Larger grains, approximately 2 mm. in diameter, are sometimes present. The curved interfaces of most feldspars has been attributed to a metamorphic feature (Vernon, 1968) in the bulk of the granite and is not always the result of resorption as suggested by Joplin and Walker (1961). The composition of the plagioclase, optically determined by the normal "a" method, is An_{22}. Myrmekite is also very common and is particularly so on boundaries where the microcline is in contact with plagioclase.

The quartz occurs as large irregular cuspat grains up to 7 mm. in diameter, (Plate 2.5a) which nearly always exhibit undulose extinction, defined by a series of misoriented subgrains which differ from each other by as much as 2 to 4 degrees. Grain boundaries are curved and grains frequently project into feldspar phases. In areas of higher mica content the grain size is significantly less with a maximum diameter of only 3 mm. Grain shapes in such regions tend to be polygonal and more regular and do not protrude into other grain aggregates as is the case with the coarse grained quartz.

The biotite is very dark brown to greenish yellow, is strongly phleochroic, and occurs as either single euhedral blades or aggregates. In
(a) A large elongate quartz grain in a coarse grained phase of the foliated Sybella Granite. The quartz has distinctly curved grain boundaries undulose extinction and protrudes into the microcline-quartz-biotite phase of the normal granite. (Specimen 7894; scale: 1 mm; crossed nicols)

(b) Shape of interfaces in the foliated phase of the Sybella Granite. The interfaces typically curve into triple-junctions giving the grains a polygonal shape. (Specimen 7666; scale: 0.5 mm.; crossed nicols)
a number of altered rocks the biotite is interleaved with chlorite. It is finer grained than the other phases, being up to 1 mm. in length parallel to the (001) cleavage, and generally less than this, normal to (001). There is usually an obvious preferred orientation of biotite in most specimens.

Muscovite is never as abundant as the biotite but has a similar grain size. It is usually euhedral in shape, but some larger grains may be subhedral or anhedral, and may contain small bleb like inclusions of quartz.

The accessories, ilmenite and apatite, occur as rounded grains up to 0.5 mm. in diameter and are irregularly scattered. Some large magnetites with square outlines are also found, these only occur in certain phases of the granite such as at 15000 W 28000 S where they occur as 5 mm. diameter grains. Flourite is found rarely and generally occurs as aggregates of large crystals up to 6 mm.

Foliated granitic rocks containing numerous xenoliths occur at the northern nose of the main intrusive body and in many of the smaller intrusions. There is a strong linear fabric with the xenoliths of amphibolite and quartzite (Plate 2.4b) plunging to the north at a high angle; the same as the
mineral lineation, in the country rocks. The composition of these rocks does not differ markedly from the rest of the Syabella Granite except for a higher biotite content (>10%) and a greater percentage of a dark green-brown hornblende (5%). Some amphibolite xenoliths have been modified with the development of biotite at the expense of the hornblende. The quartzite xenoliths may show extreme deformation without mineralogical changes taking place, except for the addition of some potash feldspar.

The massive unfoliated phase of the granite is typically found in the southern and central portion of the large granite body south of Mica Creek and in many of the smaller intrusions. Mineralogically it is very similar to the foliated phase except for the spatial distribution of minerals. The biotites do not have an extremely strong preferred orientation nor are there pronounced blastophenocrysts with strong lenticular shapes. Phenocrysts of single microcline crystals are found instead; they are euhedral or subhedral and may lack curved interfaces and triple point junctions. There is also no marked distinction between mica rich versus mica poor areas.

Minor intrusions of microgranite have been observed, occurring as small stocks and dykes. They
are identical to the microgranites described by Joplin (1955, p. 56), except they contain few xenoliths and may sometimes have a poor foliation. Thin dykes of microgranite are frequently observed intruding the coarser porphyritic granites and they have also been observed intruding parallel to the axial plane of second generation folds (See Plate 2.6).

Intensely foliated granites have been described by Carter et al. (1961) to the south of the area mapped, but none have been found in the present study. Lister (pers. comm.) from the area west of Mount Novit has found a number of granites possessing a strong foliation so that they appear as augen-gneisses and blasto-mylonites.

6. Pegmatites

The numerous pegmatites shown on Map 1 are genetically related to the Sybella Granite and are of two types, those occurring within the granite and those within the surrounding metamorphic rocks.

Irregular bodies of pegmatite, a few centimetres to up to a metre in width, grade imperceptibly into the granite, or they may form sharp walled dykes or veins which cut across the foliation of the granite
PLATE 2.6

(a) A thin dyke of microgranite intruding folded Eastern Creek Volcanics at 13700 W 111405 S. The dyke is almost parallel to the axial plane of a B2 fold, and displaces the limbs, in strongly foliated amphibolites. The light layers in the amphibolite are quartz-plagioclase rich layers and the dark layers are hornblende rich and quartz-plagioclase poor layers. The base of the dyke is 25 cm. wide.

(b) The irregular contact between coarse grained pegmatite and May Downs gneiss at 29350 W 37380 S.
or country rocks (See Plate 2.6). A coarse grain size, lack of foliation, and an almost monomineralic composition characterise these bodies. Mineralogically, however, they closely resemble the granite itself except for a very low biotite and a higher microcline content. The sharp walled pegmatites may be over 50 metres wide and are strongly linear with constant widths in two dimensions, and have been observed to cut some of the more irregular pegmatite bodies. The orientation of these pegmatites with respect to the foliation and jointing in the granite and neighbouring metamorphic rocks suggests that these pegmatites filled joints in the granite body.

The largest of the pegmatites in the metamorphic rocks is the mass which extends south from the Big Beryl Mine for a distance of 2 kilometres, has a maximum width of 1 kilometre, and at the southern end adjoins and intrudes the Sybella Granite. Elsewhere the pegmatites occur as individual tabular or lenticular bodies which range up to 800 metres in length and 100 metres in width. In most instances they have been intruded parallel to the planes of schistosity. There is a marked absence of pegmatites in the schists and quartzites east of the Fault III except in the vicinity of 10000 W 22000 S, and
a nearly complete confinement of the pegmatites to the Eastern Creek Volcanics, Mount Guide Quartzites and May Downs gneiss.

Most of these pegmatite bodies are composed of microcline, quartz, albite and muscovite with minor amounts of beryl, black tourmaline, garnet, tantalite-columbite minerals, cassiterite, monazite and fluorite. The distribution and occurrence of these accessory minerals, which have been mined from a series of small open-cut pits or shallow underground workings have been described by Brooks and Shipway (1960) and O'Dea (1964). The microcline, which is the predominant mineral (over 90% of most pegmatites) occurs as crystals up to 60 cm. across, but averages 5 cm. Much of the quartz is locally intergrown with the microcline to give the feldspar a graphic appearance. The muscovite also occurs as very coarse plates or booklets randomly oriented and widely scattered throughout the pegmatite bodies.

Most contact relationships between the pegmatites versus quartzite or amphibolite are sharp, whereas contacts with the schists are not as marked. The schists away from the pegmatite bodies are well foliated medium grained, composed of quartz (30%) and muscovite, the latter occurs as well-oriented blades
of up to 1 to 2 mm in length. Biotite also occurs in well oriented blades of about the same size but is not as abundant as the muscovite; apatite, magnetite and zircon occur in minor amounts.

Schists adjacent to the contacts show numerous differences from those of the unaltered mica schist; the width of these zones is very variable from 6 metres to 10 centimetres. In general the foliation of the altered schist is poorer, the grain size coarser (See Plate 2.7a), the mineral composition may be different and more variable. One of the most outstanding differences, is the great abundance of tourmaline and sometimes chlorite in the altered schist. In outcrop the tourmaline appears as tiny black crystals ranging from a few milimetres to a few centimetres and is generally concentrated in the chlorite rich portions. Most tourmaline crystals are euhedral and lie across the foliation which is defined by blades of muscovite and biotite.

Tourmalinization is not confined to the pegmatite margins but is also found in many of the quartz rich schists. In such cases there may be a local wholesale replacement of the micas to form a rock composed of tourmaline and quartz. The most easterly example of this feature was found at 2000 W 5000 N which is
(a) Aggregate of muscovite grains in the coarse schist adjacent to the Big Beryl Pegmatite mine, (Specimen 7945). Most boundaries are of the "rational inpingement" type (Vernon, 1968), the interfaces being parallel to (001) of one of the adjacent grains. A number of exceptions are present, and no obvious preferred orientation exists in this rock. (Scale: 0.5 mm.; crossed nicols).

(b) Quartz inclusions in the muscovite grains from the above schist. The quartz-muscovite interfaces are commonly curved, and many inclusions are elongate either parallel or normal to the (001) muscovite cleavage. The doublet on many quartz interfaces is an optical effect, as the interface is tilted to the plane of the section at an oblique angle. (Scale: 0.15 mm.; crossed nicols).

(c) Quartz-chlorite relationships. The effect of anomalously large narrow chlorite grains is to form planar interfaces parallel to chlorite (001) planes. The quartz-quartz interfaces are gently curved and meet chlorite (001) interfaces at approximately 90 degrees. (Specimen 7615a; Scale 0.5 mm.; crossed nicols).
some 5 kilometres from the nearest outcropping granite or pegmatite.

The chlorite schist adjacent to the pegmatites are very coarse grained and consist of chlorite, quartz, muscovite and talc. Most of the chlorite occurs as lenticular aggregates with poor preferred orientation or as decussate aggregates. In quartz rich portions the grains may be short rectangular blades, up to 0.5 mm. in length, including polygonal quartz grains or as long slender single blades up to 4 mm. in length. These large blades inhibit the grain growth of the quartz and gives rise to a microstructure with planar interfaces (Plate 2.7c). The muscovite occurs as small stubby grains while the talc occurs as large aggregates ranging up to roughly 1 cm. in diameter of extremely fine grained material. Many of the layered silicates in these rocks are kinked and in some cases there is an extremely good crenulation cleavage present.

Several amphibolites adjacent to the pegmatites have suffered mineralogical readjustment with the hornblende being made over to a dark brown biotite mica and the formation of garnet. Large books, up to 1 metre across and 30 cm. thick, of both muscovite and biotite mica occur occasionally along the margins of many of the pegmatites.
Quartz blows occur as large lenticular outcrops, ranging from one metre to hundreds of metres in length, of clear white quartz. They are extremely common in the vicinity of the pegmatites and were mapped in detail by O'Dea (1964) between Mica and Lena Creeks. They have a distinct north-south alignment, parallel to the strike of the rocks, occur in most horizons and may be observed postdating many of the pegmatite bodies.

Well within the pegmatites are numerous large rafts of schist up to several metres across. They show intense "tourmalinization" and "muscovitization" and the boundaries of the fragments are, in many cases, vague and gradational. These xenoliths show a great diversity in orientation of the schistosity and in many they consist of a series of random micas, some of which also possess a crenulation cleavage.

7. Metamorphism

Joplin (1955) and Carter et al. (1961) have reported the major metamorphic variations in the Mount Isa-Cloncurry area in two reconnaissance surveys. Joplin (1955) recognises five different types of metamorphism:

(1) A regional metamorphism in which small areas of high grade rocks were considered
to be part of the basement and were separated from lower grade metamorphics by a "Metamorphic Unconformity" (See also Carter et al., 1961, p. 167).

(2) Retrograde rocks which are confined to narrow zones such as faults; these postdate the regional metamorphism, but pre-date the intrusion of the granites.

(3) Granite intrusions into the regional sequence, and the formation of high grade rocks as a contact phenomena.

(4) Metasomatism where the rocks have acted as an opensystem with the addition of soda accompanied by chlorine or boron; resulting in "scapolitization" and "tourmalinization" respectively.

(5) Addition around the margins of the granite where the sediments have had the addition of such minerals as microcline.

One of the examples cited by Joplin for "magmatic" addition was in the schists and gneisses around the margins of the Sybella granite. In the present study there appears to be no evidence of addition of either microcline or any other mineral. Any veins of quartz or microcline present are either pegmatites; or
a series of ptygmatically folded veins. The latter are believed to have originated as a result of the development of a differentiated layering in many of these schists and gneisses (See Chapter 3). There is only limited evidence of "metasomatism" at Mount Isa (See Section 6), this is the presence of tourmaline crystals in the chlorite schists and the "sericitization" of the plagioclase in the chlorite schists adjacent to the pegmatites. Rare and very localized occurrences of scapolite are found in the diopside-garnet marbles west of Mount Isa. This scapolite is believed to have formed as a result of an isochemical metamorphism similar to the occurrences recorded by Ramsay and Davidson (1970) at Mary Kathleen.

The gabbros in the vicinity of 20000 W 27000 S are the only rocks to show any extensive contact metamorphic effects from the Sybella Granite (See Appendix A). Other contact effects are ill defined; the most obvious and only change observed appears to be a general increase in grain size in some of the schists near the granite-country rock boundary.

The only significant metamorphic events, to the west of the Mount Isa Fault, are the regional and retrogressive metamorphisms. The regional metamorphism is ubiquitous and varies from very low
greenschist to amphibolite facies in a distance of 5 kilometres perpendicular to the regional strike of the area. The marked increase in metamorphic grade observed can be attributed to the presence of second generation strike faults (See Fig. 2.3 and Chapter 4) which divide the area into 4 recognisable metamorphic zones whose characteristic mineral assemblages are summarized in Table 2.2 and distribution are shown on Fig. 5.2a. This also explains the large increase in metamorphic grade recognised by Joplin (1955) and Carter et al. (1961) who believed that a "metamorphic unconformity" existed.

The retrogressive metamorphism is localized in narrow highly deformed zones in which the grade is lower greenschist. The best example of such a retrogressive zone is the Mount Isa Fault; here and in many of the other retrograde zones east of the Mount Isa Fault, the contrast between metamorphic grade from the adjacent metamorphics to that in the fault zone is not great. To the west of the Mount Isa Fault these retrogressive zones involve the localized deformation of higher grade rocks; where progressive changes from coarse grained amphibolite facies rocks can be traced into a zone composed almost entirely of greenschist facies rocks. Excellent examples of the
<table>
<thead>
<tr>
<th>Location of Metamorphic Zones</th>
<th>Pelite and Pasmopilites</th>
<th>Dolomitic and Carbonate rocks</th>
<th>Basic Igneous and Sedimentary rocks</th>
<th>Metamorphic Zones</th>
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<tr>
<td>East of the Mount Isa Fault</td>
<td>quartz-chlorite-albite</td>
<td>dolomite-quartz-potash feldspar</td>
<td>quartz-albite-chlorite</td>
<td>Chlorite Zone</td>
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<td>quartz-chlorite-calcite-albite</td>
<td>dolomite-quartz-phlogopite-chlorite</td>
<td>quartz-albite-chlorite-biotite</td>
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<td></td>
<td>muscovite-chlorite-quartz (=magnetite-basemite)</td>
<td>dolomite-quartz-chlorite-muscovite</td>
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<tr>
<td>between Mount Isa Fault and Fault I</td>
<td>quartz-muscovite</td>
<td>dolomite-quartz-chlorite-muscovite</td>
<td>Absent</td>
<td>Biotite Zone</td>
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<td></td>
<td>quartz-muscovite-biotite</td>
<td>calcite-actinolite-quartz (=magnetite)</td>
<td>hornblende-plagioclase(An12)</td>
<td>Biotite Zone</td>
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<td></td>
<td>quartz-muscovite-chlorite</td>
<td>actinolite-calcite-albite-quartz</td>
<td>hornblende-quartz-albite-epidote (=biotite)</td>
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<td>quartz-muscovite-magnelite</td>
<td>hornblende-quartz-plagioclase-calcite-sphene (=magnetite)</td>
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<td>quartz-magnelite</td>
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<td>quartz-muscovite-biotite-cordierite</td>
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<td></td>
<td>quartz-biotite-muscovite (=sillimanite)</td>
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<tr>
<td>between Faults I and III</td>
<td>quartz-plagioclase(An30)-microcline</td>
<td></td>
<td>hornblende-plagioclase(An30)</td>
<td>Amphibolite Facies</td>
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<td></td>
<td>hornblende-quartz-plagioclase-magnelite-sphene</td>
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<tr>
<td></td>
<td>quartz-plagioclase(An30)-biotite-sillimanite</td>
<td></td>
<td>hornblende-orthoclase-quartz</td>
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</tr>
<tr>
<td>West of Fault III</td>
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<td></td>
<td>cordierite-anthophyllite-chlorite-plagioclase(An30)</td>
<td>Stillmanite Zone</td>
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<td>calcite-actinolite</td>
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<td>cordierite-antophyllite</td>
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<td>calcite-actinolite- magnetite</td>
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<td>cordierite-antophyllite-biotite</td>
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<td>calcite-diopside-fosterite</td>
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retrogressive metamorphism are found in the amphibolites in which the amphibole-plagioclase assemblage has been retrogressed to a quartz-chlorite schist in localized zones.

The changes observed in the prograde regional sequence will be described briefly in the following, for it was not the purpose of this thesis work to carry out an extensive study of the metamorphism. A complete understanding of the metamorphism will only be possible when detailed chemical and mineralogical data becomes available. For a detailed description of the microstructural changes observed in the pelites and quartzites with increase in metamorphic grade the reader is referred to Chapters 3 and 5 respectively.

The four regional metamorphic zones distinguished are: (1) a chlorite zone, (2) a biotite zone, (3) a biotite-cordierite zone, (4) a sillimanite zone.

The Chlorite Zone

This is an extensive zone of regional metamorphism in the Mount Isa district, the limits of which are not known, but it extends for at least 8 kilometres to the east of Mount Isa as an area of low grade dolomitic, pelitic and volcanic rocks. The pelites and psammopelites have previously been referred to as "shales" and "siltstones" (cf. Bennett, 1965) but many
rocks rich in layered silicates, such as muscovite and chlorite, possess a very strong slaty cleavage. The characteristic metamorphic assemblages in the non-mineralized dolomitic rocks are dominated by dolomite-quartz-potash feldspar (see Croxford, 1964, p. 41). The formation of any other metamorphic phase may be controlled both by bulk composition and the partial pressure of $CO_2$ (Bailey, 1964). Until the unpublished results of the work such as Van der Heuvel (1969) become available little will be known about the metamorphism of these dolomitic rocks.

Robinson (1968) mapped the distribution of basic volcanics in this zone; they are characterised by abundant chlorite, epidote, and a very sodic plagioclase and in appearance have many of the characteristics of basic volcanics altered during burial metamorphism (Smith, 1968). A description of the western volcanics which are typical of the chlorite schists found in this zone is given in Appendix A.

**Biotite Zone**

There is a predominance of quartzites over the pelitic rocks in this zone which is confined to the area between the Mount Isa Fault and Fault I. The schists found in this zone are composed of quartz-muscovite assemblages with minor biotite; intercalated
with these are quartz-biotite schists. Of very minor abundance is dolomite which is found in association with chlorite and muscovite.

**Biotite-Cordierite Zone**

This consists of two distinct areas, the first is between Faults I and II and the second is between Faults II and III. In both areas there is a large proportion of quartzite interbedded with the schist. The pelitic schists are coarsely crystalline, possess a good slaty cleavage and are composed of quartz-muscovite-biotite with large cordierite porphyroblasts lying within the cleavage (See Plate 3.7), but randomly oriented in this plane with respect to any of other linear fabric elements. The cordierite porphyroblasts are elongate crystals often including trains of fine quartz grains parallel to the foliation S₁. In some schists, especially those deformed by second generation folds, small knots of quartz and chlorite are found which are believed to be retrogressed cordierite porphyroblasts. Magnetite commonly occurs in schists deficient in biotite as euhedral octahedrons. Actinolite rich rocks are restricted in this zone to small lenses or pods within quartz-chlorite-muscovite schists.
Rare and very localized sillimanite is found in this zone adjacent to the pegmatites and occurs as elongate mats parallel to the mineral lineation found elsewhere in these schists. It is believed that this sillimanite is related to the intrusion of the pegmatites and is a result of localized alteration of the quartz-muscovite-biotite schists, an origin similar to that envisaged by Zwart (1963) for sillimanite in the Central Pyrenees.

Basic volcanics occur as intrusives (See Appendix A) and are characterised by a green-blue hornblende, quartz and a sodic plagioclase feldspar.

No marked change in metamorphic grade has been observed across Fault II and it is believed that the two areas represent the same metamorphic grade. The metamorphic event being contemporaneous with and postdating the first generation folding, but predates the second generation deformation which is responsible for the retrogression of the cordierite.

**Sillimanite Zone**

The schists and gneisses of pelitic composition are characterised by the assemblages muscovite-biotite-sillimanite, and hornblende-plagioclase-quartz in rocks of basaltic composition. The latter are generally basic intrusives described in Appendix A. There is no
evidence for the existence of cordierite, but there is an abundance of plagioclase and microcline. The initial bulk composition of these gneisses then differs significantly from the lower grade schists. In the gneisses there is apparently a greater abundance of alkalis whereas the schists presumably have a higher percentage of iron and magnesium which would account for the abundance of cordierite and magnetite.

The sillimanite in this zone is not ubiquitous, it occurs as needle-like crystals within a coarse aggregate of polygonal shaped quartz plagioclase, microcline with minor biotite and muscovite; it is also extremely common as sheaf-like aggregates in second generation folds (See Section 4.1, Chapter 4).

The basic volcanics have previously been described (Section 2.11, Chapter 2). The hornblende-plagioclase rich rocks are all equivalent to metamorphosed extrusives and some sediments, whereas the hornblende-quartz-orthoclase and cordierite rich assemblages are all derived from the interbedded sediments. In the cordierite rich rocks there is generally an enormous variation in assemblages even in the same outcrop. The variation in the carbonate rocks is also very localized being confined to small lenticular zones within the basic volcanics. Fosterite marble has only
been found at one locality (12500 W 15000 N) and much of this has been retrogressed to serpentine. In general retrogression has only been noted in the hornblende-plagioclase rich amphibolites and is of two types. The first is associated with the zones of intense second generation folding, and the second has only been noted in the vicinity of the pegmatites and consists of irregular zones composed of albite and randomly oriented chlorite. These are presumably associated with "metasomatic" effects from the pegmatites.

Geochronological data (Appendix C) defines two major metamorphic events, one at approximately 1930 m.y. and the other at 1665 m.y.; which are the regional and retrogressive metamorphisms, respectively. The microgranite and pegmatites are dated at 1560 m.y., which is in agreement with the observation that the pegmatites post-date the retrogressive zones and are apparently responsible for the very localized alteration of the adjacent schists.
1. Introduction

The interpretation of any mesoscopic structure in a deformed sequence of rocks is dependent on the nature of axial plane structures and on overprinting relationships existing between the various structural surfaces. The criteria presently used to define these surfaces and different generations of folds is an outcome of the pioneering work of Bruno Sander (1930), which was brought to the attention of English speaking geologists by two different schools: firstly, the American, through the early work by such people as Knopf and Ingerson (1938) and Fairbairn (1949), who were primarily concerned with the microscopic features in a deformed rock, and secondly, the British workers, such as Phillips (1937) and McIntyre (1951), who were concerned with both microscopic and field applications of Sander's work.

In this chapter, the approach developed by these workers is used to classify the various structural elements observed in the area into three main groups which are thought to represent a chronological sequence. These are labelled first, second and third
generation structures and a summary of these features is presented below:

\[ S \quad \text{a lithological layering which is equivalent to bedding} \]

First generation structures

\[ S_1 \quad \text{a slaty cleavage foliation or a metamorphic layering} \]

\[ I_M \quad \text{mineral lineation contained in } S_1 \]

\[ B_1 \quad \text{the first recognisable folds} \]

Second generation structures

\[ S_2 \quad \text{strain slip cleavage} \]

\[ B_2 \quad \text{folds, which overprint } B_1 \text{ structural elements} \]

Third generation structures

\[ S_3 \quad \text{axial plane trace or a strain slip cleavage which post-dates the } S_2 \text{ cleavage} \]

\[ B_3 \quad \text{open kink folds, restricted to zones of intense second generation folding.} \]

This chapter begins by defining the bedding foliation, \( S \), and this is followed by description and definition of the relationships between the foliations and lineations associated with the mainfold groups and their variation with different rock types. These fabric elements will be related to the macroscopic structure in Chapter 4.

All structural terms used here are as defined by Turner and Weiss (1963), and geographical north, as opposed to mine co-ordinates, is used as datum for all
direction measurements. Data is plotted on the lower hemisphere using equal area projections.

2. **Foliation S**

This is bedding, and it is the first recognisable lithological layering: it is generally planar, but may be lenticular, sometimes contains sedimentary structures and occurs as regularly alternating units of different thicknesses. It is frequently folded into an orientation, which is parallel to the axial plane of either first or second generation folds and it is commonly highly modified by metamorphism. This, in the extreme case, may result in the strong development of a metamorphic layering which tends to obliterate the bedding.

3. **First Generation Structures**

First generation structures are characterised by a slaty cleavage foliation known as $S_1$, an association linear structure $B_1$ and mineral lineation $L_M$. They are the first recognisable fabric elements in the area post-dating the sedimentary layering $S$. The morphology of foliation $S_1$ changes markedly both with metamorphic grade and rock type throughout the area and these changes are described below.
(i) **Foliation S₁**

$S_1$ is a slaty cleavage, or metamorphic layering defined in most psammitic or pelitic rock types by a preferred orientation of (001) of mica or other layer silicates, the extent of its development and morphology is dependent on the metamorphic grade and bulk composition. There is also a marked difference between the morphology of the foliation $S_1$ found in the pelitic rocks in comparison to that found in the Eastern Creek Volcanics. Below is a discussion of the morphology of the slaty cleavage in the area and the variation in different rock types.

**Pelitic Rocks**

Most micaceous rocks west of the Mount Isa Fault commonly possess a good slaty cleavage, whereas those to the east have a very poor development of the cleavage. In the majority of dolomitic rocks, east of the Mount Isa Fault, layer silicates are rare and hence there is little development of slaty cleavage. Rosengren (1968), however, recognised the existence of a very good slaty cleavage lying at a low angle to bedding in the Spear-Kennedy and Magazine Formations. In most of these rocks this slaty cleavage is parallel to bedding and has previously been believed to be a bedding facility. Evidence that the cleavage found
in the mine area is associated with the first generation folding was recognised by Williams (1969), who found one fold in which slaty cleavage is parallel to the axial plane of the fold. This is a tight intrafolial fold which is confined to one zone of bedded layers, which themselves are enclosed by other almost parallel layers folded by the more open folds typically found in the Mount Isa Group (See Appendix B). Williams (1969) believes that the parallelism of the cleavage and bedding in the mine area is a result of transposition.

In the chlorite zone, most of the large quartzes, and mica blades in the pelites are detrital and merge into a fine matrix of extremely fine quartz and randomly oriented layered silicates. In the more quartz rich rocks (quartz to layer silicate ratio of 50 : 50) there may be a dimensional preferred orientation of the apparent long axes of the quartz grains and this may be accompanied by the alignment of the trace of (001) of mica and chlorite parallel to the direction of quartz elongation. In these rocks small "mica beards"* may also form at the ends of these

*"Mica beards" may be observed in thin sections perpendicular to the cleavage, and are quartz mica intergrowths in which the (001) cleavage of the layer silicate meets the surface of the detrital quartz grain at a very high angle. The trace of (001) is usually parallel to the direction of quartz elongation and helps to define a poor foliation.
quartz grains. These larger quartzes retain what could be interpreted as a detrital shape (Plate 3.1a), and the effect of the "mica beards" on the quartz grain boundaries is to give them a ragged and ill defined appearance.

The pelites within and immediately adjacent to the Mount Isa Fault consist of a quartz-muscovite assemblage with an average grain size of 0.03 mm. and have strong preferred orientations of the basal planes of mica. The mica occurs as a series of parallel trains between which are small quartz grains as illustrated in Plate 3.1b. Individual mica grains tend to be poorly defined and the ends are commonly shredded especially at quartz mica junctions. There is little evidence for the presence of detrital micas in these slates, but there are the occasional coarse blades of mica oriented with the trace of (001) at high angles to the foliation. These micas tend to have squarish outlines, are generally larger than the other strongly oriented micas and are commonly composed of a biotite and an interlayered chlorite-biotite mixture.

West of the Mount Isa Fault, in the biotite zone, quartz-mica junctions in the schists become very well defined, quartzes occur as individual elongate grains with definite grain boundaries, and are surrounded by
(a) Detrital quartz grains surrounded by fine metamorphic micas. Note ragged appearance of quartz boundaries although in this microphotograph there are no true "mica beards". (Specimen 7904; Scale: 0.1 mm.; crossed nicols)

(b) Slaty cleavage in 7779b from slate in the Mount Isa Fault zone. This exhibits good preferred orientation of (001) in muscovite in a thin section which is approximately 10 microns thick. In thicker thin sections the microstructure of these fine grained foliated rocks are obscured by the presence of more than one layer of minerals. The darker portion of the section is composed of fine quartz grains. (Scale: 0.02 mm.; crossed nicols)
sheaf like aggregates of mica (See Plate 3.2a). In the biotite-cordierite metamorphic zone many of the micas become partially included in the quartz grains and have characteristically rounded ends (See Plate 3.2b).

In the sillimanite zone the fabric of the pelitic rocks consists of large single mica blades and/or aggregates of interlocking mica blades, dominated by a strong preferred orientation of their (001) cleavage trace (See Plate 3.3). In some of the schists adjacent to the pegmatites both the muscovite and the minor quantities of biotite are extremely coarse grained interlocking blades (See Plate 2.7a). The preferred orientation of these micas also varies from randomly oriented to very strongly oriented. However, these rocks do not appear to be typical of the higher grade schists.

Psammitic Rocks and Quartzites

In the biotite and biotite cordierite zones, quartz rich rocks with a quartz to mica ratio of 70 : 30, have grain sizes which are generally coarser (0.1 mm.) than the pelites (0.03 mm.) and the metamorphic mica occurs as very much smaller blades with two distinct morphologies. The first is an anastomosing structure with discontinuous mica trains outlining a series of
(a) Elongate sheaf like aggregates of muscovite mica with lenticular and shredded ends in schists immediately to the west of the Mount Isa Fault. (Specimen 7780; Scale: 0.5 mm.; crossed nicols; Locality 1480 W 14900 S)

(b) Well defined muscovite mica partially included by quartz and ends of micas are well rounded. The quartz occurs as elongate grains between the mica. (Specimen 7712b; Scale: 0.1 mm; crossed nicols; Locality 9111 W 11290 N)
Large sheaf-like aggregate of muscovite mica with a strong parallelism of individual blades.
(Specimen 7844; Scale: 0.5 mm.; crossed nicols)
almost parallel augen shaped aggregates of quartz. The quartz in these augen like masses tends to be elongate parallel to the intersection of the mica trains, and elongate between adjacent mica trains, giving the rock a distinct linear fabric, or mineral lineation $L^M\times$. A large number of these quartz grains are detrital quartzes; with strong undulose extinction and small new grains on their grain boundaries (Plate 3.4a).

The second case is applicable to the more quartz rich rocks where the mica fabric is variable in its development; here individual mica blades are dispersed between, and are sometimes included by large quartz grains. A foliation commonly exists as elongate quartz grains or aggregates of quartz parallel to the trains of mica (See Plate 3.4b).

In the biotite-cordierite and sillimanite zones, micas may have a significant effect on the microstructure developed. The micas can occur as distinct tabular shaped grains sitting along part or the whole of straight quartz-quartz grain boundaries. Such rocks are often characterised by polygonal shaped quartz grains. On the other hand in rocks with a higher percentage of mica, the quartz grains are often markedly elongate in one particular direction, which is parallel to the trace of the (001) cleavage in the micas.
Variation of morphology of $S_1$ in the psammites

(a) Anastomosing cleavage defined by irregular trains of muscovite and chlorite which outline large detrital quartz grains.
(Specimen 7936a; Scale: 1.5 mm; crossed nicols)

(b) Individual mica blades dispersed between polygonal and sometimes elongate quartz grains.
(Specimen 7884; Scale: 1.5 mm; crossed nicols)
These two features are often observed together.

In the almost pure quartzites the micas have only a slight influence on the fabric. The quartzes occur as polygonal shaped grains with mica sitting along part of the whole of the quartz-quartz boundary. In some sillimanite zone quartzites micas are often found completely included within large quartz grains (See Chapter 5).

There is then a marked change in the morphology of the foliation $S_1$, which apparently accompanies changes in the quartz to mica ratio. It is only in the rocks with a quartz to mica ratio of approximately $40 : 60$ that there is a strong development of the anastomosing cleavage, if a rock is any more micaceous then only a strong planar fabric will generally develop, and no linear fabric is generally observed. There are occasional exceptions, with mica schists containing small lenses of quartz and large porphyroblasts lying across or within the foliation and elongate parallel to the mineral lineation found in the more quartz rich rocks. In the quartzites the development of a slaty cleavage is not common and a mineral lineation is also rare.

The quartz versus mica rich layering associated with the development of the slaty cleavage $S_1$ may be
obvious on a microscopic scale but this is generally not observable on a mesoscopic scale. Most of the layering observed in the lower grade rocks appears to be inherited, but may be exaggerated by the slaty cleavage development, especially in limb regions of $B_\perp$ folds, where bedding has been transposed parallel to the axial plane slaty cleavage.

The layering parallel to $S_\perp$ tends to develop preferentially in the gneisses and is especially well developed in the more pelitic varieties. It is defined by alternating lenticular layers of coarse grained quartz-microcline versus biotite rich layers and is best observed in hinges of $B_\perp$ folds (Plate 3.5). Individual layers vary in thickness from 1 mm. to 1 cm. and often grade into one another without any clear demarcation. The micas in both types of layers have a strong preferred orientation, parallel to the $S_\perp$ layering or gneissosity.

**PLATE 3.5**

Differentiated layering in quartz-mica rich layers and axial plane to a first generation fold which also contains small disharmonic folds. The lighter layers are microcline rich and the grey layers are quartzites and do not possess a differentiated layering.

(Locality 17000 W 5000 N; Scale: 30 cms.)
The gneissosity throughout the May Downs gneiss is commonly parallel to the major lithological boundaries, and in the very quartz rich varieties can only be distinguished from bedding with difficulty. Commonly parallel to the $S_1$ layering in these gneisses and quartzites are thin monomineralic layers rich in microcline, pegmatites, which thicken and thin markedly along strike. In many of the quartz rich rocks that do not possess a strong $S_1$ layering these pegmatitic veins are ptygmatically folded.

**The Amphibolites**

There is a layering developed in the amphibolites which is equivalent to $S_1$ in the pelitic rocks. This layering varies considerably with composition. Many of the deformed amphibolites show no pronounced mesoscopic layering or parting, they appear to be completely massive, both in thin section and outcrop. Others show a pronounced layering. In general the amphibolites of recognisable igneous derivation show a poor layering, except for some of the metamorphosed basic intrusions (See Appendix A). Those of probable sedimentary origin generally have a very conspicuous layering.

The bulk of the Eastern Creek Volcanics are of apparent igneous derivation and the layering when
present is defined by an alteration of dark hornblende rich and plagioclase poor layers with light plagioclase-quartz and hornblende poor layers or lenses, typically 0.2 to 2 mm. wide. Common blue-green hornblende and plagioclase, An 30-50, are the most important minerals, hornblende forming over 60% of the rock. In detail the boundary between layers is ill defined and ragged, it is therefore difficult to resolve in thin section.

A distinct compositional layering has been imposed on many of the bedded amphibolites. It is variable in thickness and may be accompanied by a strong linear fabric. This linear fabric is best developed in many of the quartz rich sediments which are interbedded with the amphibolites and is defined by a biotite or epidote mineral streaking (See Plate 3.7). The layering is usually less than 1 cm. in width, of variable length and is more pronounced than in the igneous type amphibolites. There is also an increase in modal quartz and the appearance of potash feldspar. These two minerals also lighten the overall colour of the rocks and help to define the feldspar rich layers.

In the cordierite-anthophyllite rocks $S_1$ development is also dependent on the mineralogical composition. Rocks rich in anthophyllite consist of numerous large sheaf-like aggregates or single needle-
like crystals oriented so that their long diagonal is parallel to the foliation $S_1$. Between adjacent trains of anthophyllite are aggregates of polygonal shaped quartz grains (See Plate 3.6). Other rocks may contain high percentages of biotite which also has a strong preferred orientation, defined again by blades of biotite and anthophyllite of varying size, with a strong dimensional orientation parallel to the length of the mica.

In rocks depleted in mica and anthophyllite the cordierites occur embedded in an aggregate of fine polygonal grains without the development of an augen structure. The bulk of the rock is composed of quartz and albite aggregates, the grains commonly having a polygonal shape, with gently curved grain boundaries meeting at triple point junctions. The grain size range within these aggregates appears to be relatively small (Plate 3.6).

(ii) Mineral Lineation $L_M$

This lineation lies within the foliation $S_1$ and is seen mesoscopically as a mineral streaking or rodding which is commonly associated with and parallel to first generation fold axes. The best development of the lineation is within the biotite zone mica rich
(a) Cordierite-anthophyllite rocks with a very strong preferred direction of elongation of the anthophyllite crystals. Between the anthophyllite are elongate cordierite and quartz grains. (Specimen 7638; Scale: 1.5 mm.; crossed nicols)

(b) Cordierite-biotite-anthophyllite rock in which biotite and minor anthophyllite have a very strong preferred orientation. The cordierite occurs as large poikiloblastic crystals, some of which are twinned, and also contains numerous quartz inclusions. (Specimen 7859b; Scale: 0.5 mm.; crossed nicols)

(c) Outcrop of cordierite-anthophyllite rock. A prominent quartz sheeting also lies in the foliation S₂, and many of these layers are highly folded. (Locality 12220 W 11660 S, same as specimen 7638)
quartzites and the sillimanite zone amphibolites; it is generally absent in the higher grade quartzites.

The lineation $I_M$ is not very conspicuous in the low grade quartzites, but when present, occurs as a very fine ribbing on the foliation $S_\perp$. It is defined by a series of quartz aggregates surrounded by anastomosing trains of mica; in sections normal to $I_M$, the aggregates are elongate parallel to the foliation and in sections parallel to $I_M$ the aggregates occur as elongate strips.

In some pelites large porphyroblastic biotite mica blades lie within the foliation plane $S_\perp$. These occur as either single plates or interlocking grain aggregates which have a strong dimensional preferred orientation parallel to the length of the aggregate. Many of the higher grade schists also contain porphyroblasts of cordierite, but these seldom define a strong preferred mineral elongation, and lie randomly in the plane of the slaty cleavage, $S_\perp$ (See Plate 3.7b). In such rocks there may also be a very faint biotite lineation which is generally parallel to the $B_\perp$ fold axes, but is wrapped around the cordierite porphyroblasts.

The lineation $I_M$ in the amphibolites is defined by an elongation of the amphibole grains, within a matrix of finer randomly oriented plagioclase,
(a) Strong mineral lineation in quartzite interbedded with amphibolites in the Eastern Creek Volcanics. The lineation is defined by a strong dimensional orientation and a concentration of biotite and epidote crystals into rod shaped lenses on the surface $S_1$ (Locality 13750 W 26150 S)

(b) Quartz-cordierite-biotite schist with large cordierite porphyroblasts lying in the foliation $S_1$. The lineation on the $S_1$ surface is a bedding-cleavage intersection and a parallel mineral elongation of small biotite plates (Locality 9900 W 9800 N)
orthoclase and quartz. The schists and quartzites interbedded with these amphibolites also have a strong development of the lineation $I_M$; defined by a concentration of small lenses of epidote or biotite in the plane of the foliation (Plate 3.7a).

(iii) $B_1$ Folds

$B_1$ folds are not abundant, but have been observed in most units in the Mount Isa district, and are generally characterised by a strong development of $S_1$ as their axial plane structure. The folds occur on all scales from small mesoscopic folds to over two kilometres in amplitude. However, there is an almost complete absence of any recognisable folds in the foliated quartzite unit and the amphibolites of the Eastern Creek Volcanics.

In the psammitic units of the Judenans and some of the gneisses the folds generally vary from isoclinal to open concentric folds (See Figure 3.1 and Plate 3.8). In the case of the fold profiles depicted in Figure 3.1b there has been a small displacement parallel to $S_1$, probably contemporaneous with folding, for there is now an absence of the opposed fold hinge. Other folds, such as the one shown in Plate 3.9a, have suffered very extensive plastic deformation with the width of the zone occupied by the hinge being much
FIGURE 3.1

Profiles of $B_1$ folds

(a) $B_1$ fold in Specimen 7758 from 1540 W 17150 S

(b) Tight $B_1$ closures at 150 W 12000 N separated by a small fault parallel to the axial plane slaty cleavage.

(c) $B_1$ fold in interbedded quartzite and schist at 18500 W 9000 S.

(d and e) at approximately 1000 W 37000 S.
In (e) the psammitic layering lenses out within the quartz muscovite schist.

(f) $B_1$ fold in quartz gneiss with the development of a very strong metamorphic layering parallel to the axial plane of the fold. From approximately 18000 W 4150 N.

(g) $B_1$ fold in quartz microcline gneiss in specimen 7881a from 12500 W 6000 S.
Parallel to the axial plane are two quartz microcline muscovite pegmatitic veins.

(h) $B_1$ fold in interbedded quartzite and gneiss from 27000 W 38000 S.

(i) Open $B_1$ folding in quartz chlorite schist in specimen 7826 from 0760 W 11920 N.
(a) Very tight B, folds in May Downs gneiss, looking south, with a dextral or easterly vergence at 16900 W 7000 N. The length of ruler is one foot (approximately 30 cms.).

(b) Open B, fold within interbedded psammite and petite in the Judenan Beds. There is also an extremely well developed slaty cleavage. (Locality 1700 W 46000 S)
(a) Profile of isoclinal B_1 fold in the May Downs gneiss at 29600 W 37800 S.

(b) A quartzite unit lying at a moderate angle to the slaty cleavage and displaced parallel to cleavage in a thick schist unit at 11090 W 10000 S. Similar beds of quartzite are often observed as isolated lenses.
wider than that occupied by the limbs.

In areas where the pelite beds are thick compared to the psammite there is generally considerable radial thickening in the hinge region. In some tight closures in the psammite beds, the psammite may project into the pelite so that there is considerable thickening of the bed locally (Figure 3.1c). In extreme cases in the very asymmetrical folds (Figure 3.1e) many hinges may become intrafolial and be left completely isolated, while the limb regions form small lenses or boudins within the slaty cleavage foliation (c.f. Plate 3.9b).

In the quartzites, especially the Mount Guide Quartzite, the folds commonly approximate to the concentric fold style (See Figure 3.1h) but where they are interbedded with equal quantities of pelitic material there is generally considerable radial thickening in the hinge region. Hinge regions may also contain small parasitic flexures such as in Figure 3.1c. Many of the folds observed in the Mount Guide Quartzites cannot be placed in any one fold generation, for there is no axial plane structure or overprinting criteria to identify the generation of the folds.

On the mesoscopic scale cleavage $B_1$ folds are almost parallel with a small fanning towards the
convex side of the hinge. The axial plane structure is only visible in the pelitic and psammitic rocks, and is not present in the quartzites.

(iv) **Transposition associated with \( B_1 \) Folds**

Associated with the development of the \( B_1 \) folding in the area is the phenomenon of transposition where bedding has been transformed into an orientation which is effectively parallel to the orientation of the axial plane cleavage \( S_1 \), and lies at quite a discordant angle to the gross orientation of the enveloping surface. (c.f. Turner and Weiss, p. 92 and p. 111). Thus the bedding surfaces observed on the mesoscopic scale may have no stratigraphic significance as pointed out by Baird (1962) and Hobbs (1965). The bedding observed at Mount Isa, both in the mine and immediately west has commonly been believed to represent the orientation of large bedded units which are dipping to the west at a fairly high angle (c.f. Bennett (1965), Carter et al. (1961), and Knight (1953)). In outcrop this bedding dips west at approximately \( 70^\circ \), but within this layering there are very tight rootless intrafolial folds (See Figure 3.1b), and quite sudden changes in the facings of many beds within very short distances can be observed, even though there are no apparent fold closures. An excellent example of this may be
seen in King Gully where the mesoscopic bedding dips west at 70°, but when the contact between two different units is viewed from a distance the units are seen to be dipping east at a fairly shallow angle. This is sketched in Figure 3.2a. The contact, then, is an enveloping surface to a series of appressed folds; fold hinges have been obliterated, so that the folded nature of the surface is no longer easy to recognise. Instead the surface probably consists of numerous micro-lithons (de Sitter, 1956); the situation is that described by Sander (See Ingerson, 1938, p. 29) to explain the completion of the process of transposition: he states -

Where there has been isoclinal folding continued movement in the original S-planes of the limbs may destroy the crests and troughs so that folding is no longer discernable. This transposition by folding ("Umfaltung") of the original S-planes produces repetition and intercalation of strata and other original layers.

Other evidence to support the existence of transposition in the area includes structures which resemble cross bedding, but are really limbs of folds in which the hinge area has been dismembered or has been obscured as depicted in Figure 3.2c. Other examples of transposition features found in the Mount Isa rocks are also shown in Figure 3.2.
FIGURE 3.2

Structures associated with transposition

(a) Sketch of exposure on the south side of King Gully (800 W 15000 N) looking south. The major lithological units are dipping east while the transposed bedding is parallel to the axial plane cleavage and dips west at approximately 70 degrees. The presence of Fault I is indicated by the presence of large quantities of white quartz.

(b) An outcrop map showing a quartz-muscovite-biotite schist unit (bedding S) lying at a high angle to a lithological layering (S1) observed in quartz-muscovite schists at 9100 W 28500 S.

(c) Layering which resembles cross bedding in psammites and pelites at 2900 W 26850 S.

(d) Lenticular layering parallel to S1 at 2000 W 15000 S.

(e) Truncated fold closures at 2850 W 26000 S.

(f) A layering in quartzite at 12500 W 3500 S which can be mistaken for a transposition layering (current bedding type); the major layers are bedding (S) with a gneissosity (S1) lying at an oblique angle to bedding. The gneissosity is muscovite-biotite or microcline layering alternating with quartzite.
Quartz along line of Fault I

Mount Isa Fault Zone

(A) 15 cm

(B) 30 cm

(C) 15 cm

(D) 10 m

(E) 10 cm

(F) 15 cm
4. Second Generation Structures

Two quite distinct patterns of second generation folding and cleavage development occur in the area and both are characterised by a strain slip cleavage, axial plane to the second generation folds. The distinction between the two is essentially a difference in scale and distribution. The first group occurs as areas of large scale open $B_2$ folds which are separated by narrow zones of intensely deformed rocks of the second group. The second generation cleavage structure throughout the area shows a consistent overprinting relationship to $S_1$ and $B_1$ structures, and is a strain slip cleavage or a metamorphic layering.

(i) Foliation $S_2$

The development of foliation $S_2$ is generally dependent on an earlier anisotropy defined by $S_1$. Hence in the pelites and psammopelites where there is a strong mica anisotropy defining $S_1$, $S_2$ has its maximum development. There is also a marked variation in the style of $S_2$ from small microfolding of the earlier slaty cleavage to the development of a prominent mica versus quartz-mica differentiated layering. The style of the cleavage also changes with different rock composition.
Pelitic Rocks

The cleavage $S_2$ attains its maximum development in these rocks and consists of numerous small microfolds or microkinks aligned to form discrete parallel domains. The limbs of the microfolds consist of mica from the pre-existing slaty cleavage, and are all similarly oriented in any one domain. The domains are generally discontinuous along their lengths, with a tendency to disappear, or become much more open and rounded, especially in quartz rich layers. Domain boundaries are generally planar, but because of their lenticular nature they may have local irregularities about the region of propagation of a new set of kinks.

Further development of this foliation occurs with rotation of micas from the limb regions into an orientation parallel to the domain boundary, which separates the individual microfolded areas. The extreme development of this foliation is the formation of a mica versus quartz-mica differentiated layering in which the mica rich layers have been exaggerated by the growth of new metamorphic micas (See Plate 3.10a). The quartz rich layers are generally remnant crestal areas of the microfolds. This new layering is generally discontinuous along its length with individual zones varying quite considerably in width.
(a) Differentiated layering in quartz muscovite schist. The quartz rich layers generally contain the only evidence of the pre-existing slaty cleavage. The orientation of the (001) cleavage trace in the mica lies at a slight angle to the mica rich domains. 
(Specimen 7772; Scale 0.25 mm.; crossed nicols 
Locality 2080 W 22380 S)

(b) The same specimen which also shows the lenticular nature of the differentiated layering $S_2$ as it crosses a quartz rich layer from a mica rich area. 
(Scale: 0.25 mm.; crossed nicols)

(c) $B_2$ microfold in foliated amphibolite. The hornblende has been retrogressed to chlorite in the hinge of these small folds. The light layer is composed of quartz and plagioclase. 
(Specimen 7625c; Scale: 1.5 mm.; crossed nicols; Locality 14700 W 22970 S)
In some B_2 folds in biotite schists interlayered with the May Downs gneiss and Mount Guide Quartzite, large sheaf-like aggregates of sillimanite have grown; the sillimanite occurs as distinct lenses visible in hand specimen, in a matrix of muscovite and minor biotite. The lenses are elongate, lie parallel to the axial plane, and are only found in the hinge region of second generation folds. The sillimanite is fibrous and rarely exhibits the epitaxial growth relationship with the biotite described by Chinner (1961).

Psammitic Rocks and Quartzites

These cannot develop a strain slip cleavage (sensu stricto) because of the almost complete lack of mica. Instead, for quartzites subjected to extreme second generation folding, a very strong metamorphic layering is developed and the rocks take on a mylonitic appearance (See Plate 5.9). Such rocks are only found in the zones of intense deformation such as the Mount Isa Fault. The layering in these rocks is defined by a series of elongate highly undulose quartz grains which are surrounded and contain numerous very fine 1 to 5 micron diameter new grains. The larger elongate grains in these rocks are probably old grains which have suffered extreme deformation resulting in the production of numerous
new grains. A complete gradation can be observed from slightly deformed rocks, with strong undulose extinction and numerous deformation lamellae, to a rock which consists almost entirely of new grains.

These mylonites are generally restricted to zones of intense deformation such as the Mount Isa Fault. The majority of the quartzites subjected to second generation folding contain strong undulose extinction with occasional deformation lamellae.

The Amphibolites

The presence of a $S_2$ foliation is rare in the amphibolites although it has been observed in some folded foliated amphibolites and in areas with no obvious $B_2$ folds. The foliation is a retrogressive foliation and is generally restricted to narrow planar zones in the hinge region of the microfolds. The foliation is characterised by the presence of chlorite, which pseudomorphs and grows at the expense of the higher grade amphibole. The chlorite occurs as small aggregates of aligned blades having the trace of the (001) cleavage parallel to the walls of the microfold domains.

Zones composed almost entirely of chlorite with a strong preferred orientation, post-date the layering $S_1$ in the Eastern Creek Volcanics, but no obvious folds
may be found in these zones. These zones are considered to represent narrow zones of retrograde metamorphism associated with $B_2$ folding and the slaty cleavage found in such zones appears to be equivalent to the strain slip cleavage found in the adjacent pelitic rocks.

(ii) $B_2$ Folds

$B_2$ folds are ubiquitous and occur as two distinct patterns of folding. The first occurs as areas of comparatively open folding containing broad regions of minor strain slip cleavage. Separating these areas is the second pattern, which is composed of a series of narrow intensely deformed zones. These zones include the Mount Isa Fault and a number of other almost parallel north-south trending zones which overprint first generation fold structures. There is a strong overlap between the styles of $B_2$ folds between the two occurrences, but the intensely deformed zones appear to have a higher proportion of isoclinal folds.

Pelitic and Psammitic Rocks

The regions of open $B_2$ folds are developed in all rock types. Profiles of these $B_2$ folds tend to have rounded closures closely resembling concentric folds, of low amplitude to wavelength ratio. It is also quite common to find small parasitic $B_2$ folds or microfolds in earlier $B_1$ fold closures.
In the pelites there is usually a strong development of the axial plane structure $S_2$, as a series of microfolds. These are not, however, always recognisable in outcrop since they are often microscopic. The shape of these $B_2$ microfolds is greatly effected by composition. Normally in the pelites the folds appear to be a series of microkinks which die out upon entering quartz rich layers. The folds in the quartz rich layers are generally more open as illustrated in Plate 3.11.

Axes of most $B_2$ folds are generally non-penetrative on the scale of an outcrop. The persistence of fold hinges is also dependent on rock composition. In the psammopelitic rocks the hinges tend to be rounded and elongate structures continuous for only short distances, whereas in the pelites, where there is a well developed microkinking, axes are more persistent along their length. The non-penetrative nature of $B_2$ is probably related to the lenticular nature of the $S_2$ metamorphic layering and its subsequent influence on changes of microfold amplitude parallel to the fold axis. These non-penetrative fold axes are typically developed in outcrops where the bedding, which is commonly parallel to $S_1$, is relatively planar in the hinge regions of larger $B_2$ folds.
Profiles of $B_0$ folds in which the pelitic layers are a series of microkinks and the quartz rich layers are open concentric folds.
Scale: 5 cm.; Specimens 7747 (top) from 190 E 4000 S, and 7753 (bottom) from 000 E 6240 S.
B₂ Folds in the Fault Zones

The best examples of B₂ folds are found within the Mount Isa Fault, although occasional B₂ folds have been found in Fault I. Powell (1963) described the Mount Isa Fault in some detail and considered it to be a "shear zone". In his work, Powell only considered a number of small areas in a highly folded phyllitic unit to the west of the Mount Isa Mine (See Figure 2.2) and the boundaries of this "shear zone" were arbitrarily defined as the area in which there was a noticeable decrease in the intensity of crenulations. Powell identified four phases of folding within the Fault, but his identification of these phases was based on inadequate overprinting criteria with little consideration of axial plane structures. The folds were separated by Powell on fold shape and orientation; and most folds identified by Powell are identical to those described here as B₂ and B₃.

The style of the B₂ folds in the Mount Isa Fault Zone, which is composed predominantly of thinly laminated phyllites, is similar to the open folds with strain slip cleavage as their axial plane structure developed in the regions of broad open B₂ folding to the west. The zone is also characterised by tightly appressed folds with strain slip cleavage as their
(a - e) Profiles of B₂ folds in interbedded psammites and pelites from the Mount Isa Fault (Localities:
(a) 440 E 8000 N
(b) 250 E 4000 N
(e) 1540 W 18130 S
(d) 020 E 15930 N
(e) Specimen 7749 from 000 600 S

(f) Profiles of two B₂ folds: A very steep plunge reversal is present which results in the formation of a small dome structure. (Locality: 020 E 15930N)

(g - h) Profiles of B₂ folds in the foliated quartzite in Fault I (Locality: 1040 W 1170 S)

(i) Form surface of B₂ folds showing variation in plunge.
axial plane structure. In the pelites a metamorphic layering is commonly present. The metamorphic layering is very localized, discontinuous and appears to have been produced as a result of $S_2$ cleavage development. The layering tends to fan slightly within any one large $B_2$ fold and can be readily distinguished in the field by its overprinting relationship with respect to the slaty cleavage $S_1$.

Many of the tightly appressed folds tend to be intrafolial and can only be separated from $B_1$ structures in a great many cases by a thin section examination of the axial plane structure. There is a range in the tightness of many of these folds, especially in the more asymmetrical folds and in areas where tight $B_2$ folds with very different plunges co-exist they produce complex interference patterns in the form surface (Fig. 3.3e). Open folding (such as Fig. 3.3b) and very tight folds (Fig. 3.3a) may occur in the same outcrop (Fig. 3.3d) and in some folds such as the fold depicted in Figure 3.3e tight folds may contain small areas of fairly open folding.

Such folding is usually accompanied by numerous small axial plane faults paralleling the axial plane and the folded compositional layering $S$. The bedding and slaty cleavage, which are generally parallel from
the earlier $B_1$ deformation, are also commonly found to be parallel to the axial planes of the tightly appressed $B_2$ folds. Major areas exist where this is generally the case, and to separate the slaty cleavage from strain slip cleavage in such areas is nearly impossible except in areas containing occasional $B_2$ fold closures. In such rocks numerous examples, both on the scale of a hand specimen and outcrop exist where "current bedding" type structures have developed in which the truncated limb contains a strong strain slip cleavage parallel to a lenticular compositional layering. Many of these boundaries probably represent small faults and some have been locally infilled with irregular white quartz veins. The layering observed has probably been produced by transposition of the pre-existing compositional layering, which includes both bedding and slaty cleavage.

Therefore in the Mount Isa area a transposition layering has been produced with both the first and second generation folds. The latter example of transposition is only very localized and is contained in a number of restricted zones such as the Mount Isa Fault.
Rocks of Amphibolite Composition

The $B_2$ folding observed in the amphibolite varies from very closely spaced zones of kinking (Plate 3.12) to open concentric folds showing no evidence of kink development. The style of $B_2$ development in the amphibolites is dependent on composition, for it is only in the thinly layered rocks that kinks develop. In the more massive layers concentric folding is observed. But in an interbedded sequence of both massive and well foliated amphibolites both styles of folding may be observed in any one large fold.

The Quartzites

The variation in style in the higher grade quartzites and gneisses is very considerable. Open concentric folding in the hinges of broad regional structures is quite common. These have quite variable axial plane orientations in which the folding may be a series of box folds (Ramsay, 1967 p. 358) with the axial planes lying at high angles to one another (Plate 3.13) or with the axial planes forming

PLATE 3.12

$B_2$ folds in the foliated amphibolites

(a) A side view of a $B_2$ fold which is refolding the mineral lineation $L_M$, a hornblende elongation at 1400 W 26000 S.
(b) Profile of kink-like $B_2$ fold at 14700 W 23000 S
a convergent cleavage fan (Ramsay, 1967 p. 403) in the quartz rich layers and a divergent fan in the interbedded pelitic material in the hinges of large fold closures. Individual folds vary considerably in their persistency with amplitude and may be concentrated as zones of second generation folding between units of slightly different composition (see Plate 3.13a). Many of the $B_2$ folds observed are asymmetric, have variable plunges and are a mixture of both similar and concentric folding.

Most $B_2$ folds in the quartzites and in some of the quartz microcline gneisses have no axial plane structures and are generally indistinguishable from the $B_1$ folds. Occasionally, however, some folding may involve a micaceous layer, allowing the development of an incipient axial plane cleavage. In the lower grade quartzites this is generally readily identifiable as a strain slip cleavage. But in most higher grade quartzites, such as the Mount Guide Quartzites the micas in the pre-existing slaty cleavage are frequently coarsely crystalline, decussate and have a random preferred orientation. In such rocks the development of a strain slip cleavage is not common, although some suitably oriented micas may be kinked.
Profiles of B₃ folds in quartzites interbedded with the May Downs gneiss. There is a considerable variation in the orientation of the axial planes and a strong tendency to form box folds. Localities: (a) 22500 W 20060 W (b) 17950 W 4000 N.
FIGURE 3.4

(a - f) Profiles of B₂ folds in the Mount Guide Quartzites. There is generally no strong evidence of an axial plane structure, except for occasional micaceous layers. (Localities: (a) 259000 W 36000 S, (b) 17500 W 17000 S, (c) 12500 W 7000 S, (d) Specimen 7824 from 14850 W 1600 S, (e) 12250 W 2300 S (f) 26000 W 38000 S).

(g) Profile of B₂ fold in interbedded quartz-muscovite schist. (Locality: 1800 W 750 S)

(h) Field sketch map of large B₂ fold which contains a dome and basin structure in the core of the fold. (Locality: 12500 W 7200 S)
It is only possible to identify many of these open folds as $B_2$ folds if there is a cleavage parallel to the folded layer and if this folded surface is indistinguishable from the cleavage found parallel to the axial planes of $B_1$ folds. The quartz and microcline veins occurring parallel to the axial planes of the $B_1$ folds are also commonly folded, but in some $B_2$ folds these folded veins are also cut by another set of quartz and microcline veins which parallel the axial plane of the second generation folds.

In many places smaller parasitic second generation folds occur in the quartzites and gneisses; these are frequently coaxial and similar in shape to larger $B_1$ fold closures and have almost identical axial plane orientations. The different generations are readily recognisable in the gneisses and interbedded schists where there is a refolding of the slaty cleavage layering, but in the massive quartzites many of these folds cannot be assigned to any generation. In such areas the recognition of fold generations and interpretation of the geometry becomes rather difficult and many large dome and basin structures have been recognised (See Plate 3.14). Many of these complex structures could have
Dome and basin structures at approximately 7000 S 12500 W, where it is not possible to establish whether they have formed as a result of interference of different folds or are a result of heterogeneous strain. The scale is 30 cm. in (a) and 40 cm. in (b).
resulted from either the interference of the two fold generations or a single three dimensional heterogeneous strain (Turner and Weiss, 1963, p. 508). It is possible in some areas to identify first generation folds being refolded by second generation, resulting in the curvature of fold axes, but in many areas the variation in plunge appears to result from heterogeneous strain.

(iii) **Mineral Lineation** $L_Q$

$L_Q$ is a quartz rodding that is associated with the second generation folds. Two distinct types of lineations are found, the first occurs in the mylonites of the Mount Isa zone, the second occurs on the bedding surfaces of the Mount Guide Quartzites.

The lineation $L_Q$ in the mylonites is an extremely fine quartz ribbing which occurs on the surface $S_2$ in the quartz mylonites. In thin section it is defined by the elongation of the old quartz grains which occur in the plane of the foliation as elongate lenticules. It is these lenses that define the lineation together with the small trains of new quartz grains surrounding many of these old grains.

In the higher grade quartzites the lineation is defined by an alignment of coarse elongate quartz grains which only occur on certain surfaces. The
lineation is non-penetrative, it is generally curved and in a great many cases appears to be a quartz streaking formed as a result of simple slip taking place on the surfaces between concentrically folded layers.

5. $B_3$ Folds

These folds are confined mainly to the zones of intense deformation such as the Mount Isa Fault and occur as small scale folds and lineations overprinting pre-existing $B_1$ and $B_2$ structures (See Plate 3.15b). If an axial plane structure is developed it is a poor strain slip cleavage, but in many rocks it is absent.

The $B_3$ folds are usually open, varying from kink like folds in the pelites, to fairly rounded folds in the competent units (See Fig. 3.5). In the quartzites and psammopelites these folds are open and extremely rare, but they may be responsible for the gentle flexing observed in the axial surfaces of many $B_2$ folds. In the more competent psammitic units the $B_3$ folds tend to be more open with rounded hinges (Fig. 3.5 a,b,d) and are often accompanied by microfaulting, which is not necessarily parallel to the axial planes.

The presence of these microfaults and in some cases larger faults tends to indicate that much of
Profiles of $B_2$ folds from within the Mount Isa Fault Zone, there are refolded $B_2$ folds in (a) and (d). (e) is a sketch showing the relationship between an earlier $B_2$ lineation and $B_2$ fold axes. (g) is a sketch of the en echelon arrangement of the $B_3$ fold axes.

Localities of $B_3$ folds:

(a) 020W 15900 N

(b) 1000 W 15600 S

(c) 270 E 1300 S

(d) 300 E 000 S

(f) Specimen 7750 from 010 W 600 S
(a) \( B_2 \) folds confined to one particular horizon. There is a small variation in plunge and the axial plane trace \( S_3 \) varies in orientation within this outcrop (Locality: 300 W 15000 S)

(b) Refolded \( B_2 \) fold by a kink like \( B_3 \) fold which does not possess an axial plane structure (Locality: 400 W 8900 S)
this faulting accompanied late stage folding which included some brittle deformation, rather than purely plastic deformation as observed in the folding associated with the $B_2$ folds.

$B_3$ fold axes have a marked tendency to be gently curved, do not persist for very far along their hinges, and may not be strictly parallel to one another. These features generally give rise to an "en echelon" arrangement of fold axes (Fig. 3.5) or they may appear to be overprinting one another. The discontinuous nature and small variations in plunge of the fold axes appears to be due to a combination of both a lenticular and polyclinal axial plane cleavage, or axial surface $S_3$. Axial planes of these folds vary considerably in orientation and occur as polyclinal or conjugate sets.

There is usually only a poor development of an axial plane structure for the majority of the folds appear to have formed by a mechanism similar to that proposed for kink formation (Weiss, 1968). This appears to have been bending of the foliation by a type of flexural slip folding with accompanying slip on mechanically weak planar interfaces. It is only in a small proportion of the pelites that a strain slip cleavage developed.
Many B3 folds occur in groups and as distinct zones between essentially planar layers, which generally differ only slightly in composition. The change from folded to planar beds may be gradual (Fig. 3.5a) but commonly the two are quite sudden (Plate 3.15a) and in some cases are separated by visible faults or a zone of brecciation.

6. **Foliation in Sybella Granite**

The foliated phase of the Sybella Granite has previously been described in Chapter 2. It was noted that the foliation within the granite is sympathetic with the trend of the bedding and the foliations, which is generally a first generation foliation. In many places this foliation has been extensively folded, either locally or as broad warps such as in the vicinity of 14000 W 2900 S. The foliation in the Granite in such areas appears to be gently curved, in sympathy with the country rocks, and shows no signs of post-tectonic deformation. However, there is good evidence that in such areas the microstructure of the granite is typical of a metamorphic rock, which suggests that the bulk of the Sybella Granite was intruded during the second generation folding of the area and is therefore equivalent to a syntectonic granite. The
lineation in the granite is either a biotite streaking on the foliation plane or a strong elongation of xenoliths. The orientation of the linear fabric is parallel to $L_1$ within the country rocks. This appears to be the general direction of elongation for all the rocks in the area and the foliation possesses many of the features attributed to a primary flow foliation by Balk (1937, pp. 151-152).
CHAPTER 4
MACROSCOPIC STRUCTURE

1. Introduction

Previous interpretations of the macroscopic structure at Mount Isa have been based on stratigraphic premises, and have tended to emphasise stratigraphic rather than structural complexities. In the present interpretation the different structural events recognised in the Mount Isa area have been related to the distribution of rocks. In so doing it became obvious that many of the boundaries previously recognised are in fact faults and the rock distribution is governed by these; the majority of the units occur as north-south belts bounded by strike faults.

A clear example of the early structural interpretations based on stratigraphic arguments is exemplified in the work of Carter et al. (1961), and many of the stratigraphic relationships established by these authors have served as the basis for subsequent interpretations. A recent example of such an interpretation exists in the review paper of Bennett (1965) who interprets (on stratigraphic arguments) the area west of the Mount Isa Fault as overturned,
and appears to be in a complete dilemma regarding the spatial and stratigraphic position of the Western Volcanics in the Mount Isa area. The outcrop pattern apparently suggests to Bennett (1965) that the Western Volcanics overlie the Mount Isa Group, but then again he believes that they are stratigraphically equivalent to the older Eastern Creek Volcanics. As an outcome of the present work and the work of Williams (1969) the Western Volcanics are believed to be equivalent to the Eastern Creek Volcanics and that their upper boundary is a refolded fault (See Appendix B).

In this chapter then a broad interpretation of the structure of the area is made possible by consideration of such evidence as the existence of faults, orientation data, sense of overturning, sense of vergence, facing and local stratigraphic successions.

2. Orientation Data for the Area West of the Mount Isa Fault

The area west of the Mount Isa Fault has been subdivided into 17 subareas (Figure 4.1) which are characterised by lithology or structure. Subareas 5 and 6 are restricted to the May Downs gneiss.
Figure 4.1

LOCATION OF SUBAREAS IN THE AREA WEST OF THE MOUNT ISA FAULT
Subareas 11 to 16 are confined to the north-south trending Mount Guide Quartzites. Subarea 17 is restricted to portion of the western boundary of the Sybella Granite; the boundary dips west at 60° parallel to the margin of the granite. All other subareas are located between the Mount Guide Quartzites and the Mount Isa Fault, and the shape of these subareas is often elongate parallel to the strike of the major strike faults. The orientation data is represented in Figure 4.2.

The bedding S in the May Downs gneiss varies in orientation from almost vertical to horizontal and is gently folded about first and second generation fold axes. Many of the B1 fold axes have been redistributed by the B2 deformation and are subhorizontal plunging to 070°. The B2 fold axes are subhorizontal with steep westerly dipping axial planes. The slaty cleavage in these rocks has also been folded about the B2 folds, the majority of which occur as broad open flexures.

In the Mount Guide Quartzites (subareas 11 to 16), only rare folds can be positively identified as either first or second generation; because of an absence of overprinting criteria. In many outcrops different plunges occur side by side and folds commonly interfere
with one another. On the basis of one outcrop it is impossible to divide these folds into smaller groups using orientation, overprinting criteria, or any other property, therefore the groups of folds illustrated in Figure 4.2 as \( B_2 \) folds in subareas 11 to 16, may comprise more than one generation. The axial planes are commonly vertical or steeply dipping with an approximate north-south trend in subareas 11, 12, 13 and have a north-east trend in subareas 15 and 16. However, when these mesoscopic folds are related to larger structures such as the fold hinges in subarea 12 and the dome structure in subarea 16, it is believed that the majority of readings record the orientation of second generation folds. In subarea 11 there has only been minor refolding of bedding which dips steeply to the east. In subarea 12 a number of large macroscopic fold closures whose axial traces are gently curved. It is believed that these are first generation structures refolded by the prominent second generation folds. Only a few first generation folds have been recognised in this subarea and these plunge 80° to 250°. The bedding in subarea 12 is refolded about the second generation folds which plunge approximately 60° to 200°. In subareas 13 and 15 the bedding and axial plane traces
Orientation data for foliations, folds and mineral lineations west of the Mount Isa Fault. Plotted on schmidt stereographic nets, lower hemisphere. Contoured according to Schmidt method. Contours are $\geq 1\%$ per 1% area, $\geq 3\%$ per 1% area, $6\%$ per 1% area, $\geq 9\%$ per 1% area and $\geq 12\%$ per 1% area. The number below each diagram is the number of points plotted.

The mineral lineation in subarea 14 is $L_Q$, in the other subareas it is $L_M$.

In subarea 17 the mineral lineations and foliations were recorded from the Sybella Granite.
ORIENTATION DATA
WEST OF THE MOUNT ISA FAULT

Subareas
- 1
- 2
- 3
- 4
- 5
- 6

Poles to S
- 110° 120° 230° 187° 207° 240°

Poles to S_1
- 195° 175° 225° 189° 72° 66°

B_1
- 6
- 18
- 67
- 20

L_m
- 48

Poles to S_2
- 12
- 80° 12
- 6
- 12

B_2
- 17
- 4
- 11
- 20
are almost vertical with a marked variation of plunge resulting in small dome and basin structures. In subarea 14 bedding is refolded about macroscopic $B_2$ folds as is the mineral lineation $L_Q$. Subarea 16 is a very large dome structure containing many mesoscopic dome and basin structures, which accounts for the marked variation in the plunge of these folds. Isolated $B_1$ folds have been recognized in all subareas, except 11; these are generally locally refolded by the $B_2$ deformation, and the predominant plunge of most of these folds is approximately $80^\circ$ to $250^\circ$.

The readings in subarea 17 have been recorded in the Sybella Granite and the orientation of the foliation in the granite appears to be parallel to both bedding and cleavage in the adjacent subarea 15.

In subareas 2, 3 and 4 the slaty cleavage dips steeply due west, but has been refolded by prominent $B_2$ folds in subarea 2 and the southern end of subarea 4. Second generation folds are generally not readily observable as mesoscopic folds but occur as broad flexures (cf. Plate 4.1a) which plunge at approximately $30^\circ$ to $330^\circ$ in subareas 2 and 3 and at $20^\circ$ to either the north or south in subarea 4. In subarea 2 however, there are many narrow localized regions which contain an abundance of $B_2$ folds. The first generation folds
(a) $B_0$ flexure in interbedded quartzite, quartz-muscovite schist and chlorite schist (black) in cutting at 3130 W 3200 S (looking north with number 5E Tailings dam in the background). The chlorite schist is a retrogressed amphibolite intrusion and contains small lenses of unretrogressed amphibolite.

(b) The $B_0$ fold closure in the Foliated Quartzites at 656 W 41300 S (photo looks north). The quartzite at the top of the hill dips to the north east at approximately 50° and is truncated in the east by the Mount Isa Fault. The fault zone in this area is unfortunately covered by quartzite debris. The valley west of this hill is occupied by Fault I. Underlying the quartzites and continuing to the south between the Mount Isa Fault and Fault I are a group of black slates.
LOCATION OF THE MAJOR FAULTS IN THE AREA WEST OF MOUNT ISA

THE MOUNT ISA AND FAULTS 1 to 3 ARE STRIKE FAULTS

FAULTS a to h ARE PROMINENT CROSS FAULTS

Figure 2-3

This figure is repeated for the readers convenience.
and mineral lineation $I_M$ in subareas 3 and 4 plunge either to the north or south at approximately $40^\circ$, in subarea 2 the $B_1$ fold axes plunge $30^\circ$ to $330^\circ$.

Bedding between Lena and Mica Creeks, subareas 7, 8 and 9, is only identifiable in subarea 9 where it dips at a shallow angle in a range of directions between the north-west and the south-west. In subareas 7 and 8, which are west of Fault III, the dominant layering is the foliation $S_1$ and bedding where it is recognisable is parallel to this foliation. Very few $B_2$ folds have been recognised in subareas 7 and 8 but those present in 7 account for the redistribution of $I_M$. $B_2$ folds are common in subarea 9 as vertical or as shallow northerly plunging folds, which are responsible for the gentle folding of the bedding and slaty cleavage in subarea 9. The folds $B_1$ are parallel to $I_M$ in subarea 9 and plunge at approximately $50^\circ$ to $295^\circ$. In subareas 7 and 8 $I_M$ is a very conspicuous fabric element and generally plunges $60^\circ$ to $290^\circ$ except where it has been refolded.

Both subareas 1 and 10 (equivalent to subareas I and X in section 4.3.1) will be described later in the discussion of the rocks adjacent to the Mount Isa Fault. Subarea 10 is bounded by Sybella Granite in the west and the Mount Isa Fault in the east and
contains the continuation of strike Faults I, II and III. West of Fault I in subarea 10 the bedding and slaty cleavage dip at 80° to 270°, but between Fault I and the Mount Isa Fault bedding is either vertical or dips east at a fairly steep angle. \( B_\bot \) and \( L_M \) are parallel and plunge at approximately 35° to 350°, whereas the \( B_2 \) folds plunge 15° to 355°. Subarea 1 is restricted to the zone of rocks between the Mount Isa Fault and Fault I north of Mica Creek. In this zone transposed bedding and slaty cleavage are parallel, dipping steeply to the west. However, bedding measurements on major lithological contacts dip steeply to the east (cf. Figure 3.2). \( B_\bot \) folds and \( L_M \) are parallel to one another and plunge at 80° to 190°. Very few second generation folds have been observed, but those which have, have a variable plunge in a vertical axial plane.

Total diagrams for the orientation data are not presented here as the author believes that they would have no significance in understanding the large scale structure since the data represented in Figure 4.2 has been obtained from different structural blocks separated by major faults. These will be discussed in the following section.
3. **Faulting in the Mount Isa Area**

The position of the major faults in the area west of the Mount Isa Mine are shown on Figure 2.3 and Map 1. All but the Mount Isa Fault and some of the east-west cross faults are new faults and it is only the Mount Isa Fault which has been considered significant in previous structural interpretations of the area. Evidence for these faults will be discussed separately below.

(i) **The Mount Isa Fault**

**Previous Views**

Early workers in the area (Blanchard and Hall, 1942) had assumed that there was a conformable boundary between the dolomitic Mount Isa Group and the higher grade quartzites in the west; the Judenan Beds. An increase in metamorphic grade to the west and the truncation of the higher units of the Mount Isa Group led Knight (1953) and Cordwell et al. (1961) to define the existence of a fault, which became known as the Mount Isa Fault.

The fault has a strike length of 70 kilometres (Carter et al., 1961), with its position being inferred in many areas where it is adjacent to a major scarp separating gently undulating country, consisting of dolomitic siltstone, from hard resistant quartzite
hills. The limits and nature of the Mount Isa Fault had not been defined by any author until work on small selected areas within the fault was carried out by Powell (1963). Powell and subsequent authors have considered that the Mount Isa Fault in the vicinity of the mine is represented by a zone characterised by a distinct phyllitic lithology and intense small scale folding (See section 3.4.2).

Mesoscopic and Macroscopic Features Associated with the Mount Isa Fault

As pointed out in section 3.4.2, within the Mount Isa area narrow zones of intense $B_2$ folding exist and the Mount Isa Fault is considered to be one such zone. From north-west of the mine south to the White Blow the fault is confined, in part, to a unit of black carbonaceous phyllites and schists, but the exact boundary is generally hard to establish because of the paucity of outcrop in the area occupied by the fault. Immediately to the north and south of the White Blow area the fault zone appears to be represented by a number of fairly closely spaced parallel zones of intense $B_2$ deformation. These occur in a unit shown on Map 2 (Battey, 1962) as the Native Bee Formation. In the zone of intense deformation immediately to the west of the mine there are also
large lenses of quartzite many of which are now mylonites. These are possibly small portions of the quartzite from the Judenan Beds which have been deformed in the zone of intense $B_2$ deformation. This zone of deformation is therefore not a distinct lithological unit as envisaged by Powell (1963).

The width of this $B_2$ deformation zone appears to be quite variable, from a few metres to 50 metres, but generally it is impossible to define the exact boundaries. The surface expression of the Mount Isa Fault as a Fault, in the sense of a narrow zone of brittle failure, is not obvious, but there are a number of small breccia zones which are lenticular, being approximately 30 centimetres by 10 metres; these consist of phyllite and quartzite fragments set in a matrix of recrystallized quartz and muscovite mica. These breccias appear to be the only major evidence of brittle failure in the fault zone. Other minor evidence of brittle deformation are the localized faults which accompany the $B_2$ and $B_3$ deformation (cf. Figs. 3.3 and 3.5).

Strong evidence of ductile deformation is found in many of the mylonitized quartzites, which consist of undulose quartz grains, flat and elongate in the superimposed foliation and containing numerous new
subgrains. Other evidence for ductile deformation is the extensive development of second generation folds (See Chapter 3).

Within the $B_2$ deformation zone are many massive white quartz veins, one of the largest being known as the White Blow. The White Blow consists of two elongate irregular shaped, closely related masses some 20 metres in width, which roughly parallel the foliation $S_2$, and dip west at approximately $65^\circ$. There are also many smaller lenses, approximately 1 metre by 10 metres, and many fine irregular veins of approximately 1 cm. in width. These quartz veins are often seen cross-cutting $B_2$ folds (Plate 4.2) and are themselves sometimes folded by $B_3$ folds. It is believed that this quartz probably originated during second generation folding as a result of metamorphic differentiation associated with the formation of the $S_2$ cleavages in the zone of intense deformation. The quartz would be derived from mica rich differentiated layers and deposited close to the area where differentiation is occurring. Small zones of intense deformation are evident in the sediments surrounding many of these quartz deposits.

The strongest evidence for the presence of the Mount Isa Fault is on stratigraphic grounds. The
One of the many B₂ folds from the Mount Isa Fault intersected by irregular veins of quartz. (Locality 2900 W 22500 S). The quartz is believed to have originated as a byproduct from the strong differentiated layering formed during the S₂ cleavage development.
higher formations of the Mount Isa Group are terminated along a line, which is the eastern boundary of the zone of intense B2 deformation. These sediments are located on the eastern limb of a large synformal structure, the Kennedy Syncline (Murray, 1961) and abut at a high angle against the Eastern boundary of the B2 deformation zone and the "Western Volcanics". The Western Volcanics (Bennett, 1965) are a number of discontinuous lenticular outcrops along the eastern margin of the fault northwards from 5900 S. The eastern boundary of these volcanics always appears to be irregular and Williams et al. (See Appendix B) believe that this is a folded boundary whereas the western boundary is planar. The planar nature of the boundary has been confirmed by diamond drilling from west of the Mount Isa Fault, both to the west of the Mount Isa Mine and at the Hilton Mine some 20 kilometres north of Mount Isa (R.L. Hewett, pers. comm.).

A similar closure exists west of the Mount Isa Fault, but south of Mica Creek. Here the foliated quartzites are truncated by the fault (See Plate 4.1b) and the two overlying schist units lens out along the line of the fault between the White Blow and this closure.
Other evidence for the presence of a fault believed to be the Mount Isa Fault has been recorded by Lister (1969). Lister mapped a large refolded first generation fold in the Mount Novit area, 7 kilometres south of the southern boundary of Map 1. This structure has been truncated by a fault which is also a zone of intense $B_2$ deformation, and 450 metres to the east of this is another parallel zone of intense deformation. The intervening rocks, between these two zones, are black carbonaceous slates which contain abundant biotite mica. These slates first appear to the south of the truncated $B_2$ fold closure at 650 W 41300 S and underlie the foliated quartzite unit mapped further to the north. It is believed by the present author that the fault identified as the Mount Isa Fault by Lister is in fact a continuation of Fault I and that the eastern zone of intense deformation is equivalent to what is regarded as the Mount Isa Fault further north.

$B_2$ folds in the Mount Isa Fault zone are generally asymmetrical about the axial plane cleavage $S_2$, with a sense of asymmetry nearly always to the east suggesting a sinistral or westerly vergence.*

*The term vergence as discussed by Wood (1963) is a "sense of rotation" implied by the shape of the fold, it is also useful in defining the enveloping surface or "Faltenspiegel" of a major fold closure.
so material appears to be displaced relatively from west to east indicating that the folds lie on the western limb of a synformal structure. In the White Blow area and along the eastern edge of the zone of intense deformation a few of these $B_2$ folds have the opposite sense of vergence, dextral, with very low amplitudes. This suggests, then, that in the White Blow area the orientation of the folded surface may be flat or slightly westerly dipping.

Powell (1963) and Bennett (1965) use these vergences to define the movement along the fault and also postulate horizontal movements from the plunge of such folds. They consider movement has been "west block up and to the north", with a possible large horizontal displacement. They also equate the movement on the Mount Isa Fault with the small scale faulting visible in the mine area (Darlington, 1961; Herget, 1968; Murray, 1961), much of which also has the same sense of displacement.

While Carter et al. (1961) imply that movement on the Mount Isa Fault has been considerable and the strike slip component is in the order of kilometres, from the present work it is impossible to give an estimate of the actual movement, either dip slip or strike slip until further work is done to establish
the structure of the sediments outside the present area of study, especially east of the Mount Isa Mine.

Orientation Data for Folds Associated with the Mount Isa Fault

The area surrounding and including the B₂ deformation zone has been subdivided into ten subareas homogeneous with respect to the B₂ folds (see Figure 4.3). Subareas II to IV are restricted to the zone of B₂ deformation. Subarea IX also includes a number of very small narrow parallel zones of B₂ deformation. Both subareas I and X are west of the B₂ deformation zone, whereas VII and VIII occur to the east.

In subarea I the bedding, where it can be separated from cleavage, appears to vary little in orientation. The readings on bedding in this subarea are not considered to be truly representative of their macroscopic orientation, because of the extreme transposition (See Section 3.3.4). In subarea X bedding appears to be folded cylindrically about both B₁ and B₂ fold axes. The foliation S₁ is generally parallel to bedding in both subareas I and X, that is dipping to the west at approximately 80°, but in subarea X it has been refolded about the B₂ fold axes. The Kennedy Syncline outcrops in subarea VII, here bedding and slaty cleavage readings are
Orientation data for foliations, folds and mineral lineations in the Mount Isa Fault zone and immediately adjacent to the fault. Plotted on schmidt stereographic nets, lower hemisphere. Contoured according to Schmidt method, contours are $\geq 1\%$ per $1\%$ area, $\geq 5\%$ per $1\%$ area, $\geq 10\%$ per $1\%$ area and $\geq 15\%$ per $1\%$ area, in all diagrams except in subareas with fabric elements I and XS; II, III, IV, VII, S parallel to S; IV, VIII; III, V, VIS, IV, VIII, IXB; III, IV, VI where the contour intervals represent values of $\geq 1\%$ per $1\%$ area, $\geq 3\%$ per $1\%$ area, $\geq 6\%$ per $1\%$ area, $\geq 9\%$ per $1\%$ area and $\geq 12\%$ per $1\%$ area. The number below each diagram is the number of points plotted. The mineral lineations are $L$ in subareas I, VIII, IX and X, and both $L$ and $Q$ in subareas III, IV, V and VI.
cylindrically folded about the B_2 fold axes. In subareas V, VIII and IX bedding-slaty cleavage vary little in orientation. Whereas in subareas II, III and IV the bedding-slaty cleavage orientations vary considerably, but there is always a concentration of westerly dips.

B_1 fold axes have only been recorded in subareas VIII, IX and X, but no B_1 fold has been positively identified from within the zone of B_2 deformation. The sense of vergence of the gentle northerly plunging B_1 folds in subarea X is easterly indicating that they lie on the western limb of a synformal structure. The refolded B_1 recorded in subareas VIII and IX appear to be almost symmetrical and plunge steeply in a northerly direction. In subareas VIII and IX the B_1 folds appear to plunge to the south.

The mineral lineation in subarea I is a fine quartz rodding which corresponds to L_M. It plunges steeply to the south and occurs in many of the schists and foliated quartzites. In subareas III to VI the mineral lineation is both a mica-streaking L_M and a fine quartz rodding L_Q which plunges to the south at fairly steep angles. These two lineations were not separated in the field when the readings were recorded. In many outcrops in subarea 5 the lineation
L₂ is gently curved about B₂ fold axes, which may account for its irregular distribution. The mineral lineation in subareas VIII and IX are mica streakings which plunge south parallel to B₁ folds axes, whereas the mineral lineation in subarea X plunges gently to the north parallel to the B₁ fold axes.

The B₂ folds have a large spread of plunge orientations in their axial plane cleavage S₂. The orientation of this cleavage is generally fairly constant throughout the area dipping at approximately 80° to 265°, except in subarea IV where it has been redistributed about later vertical B₃ folds. In subarea II the plunges of B₂ folds are concentrated at approximately 40° to 340°, but in the southern subareas there appears to be a greater variation in plunge, varying from 30° N to vertical to 30° S. The number of B₂ folds observed also appears to be related to the shape of the folds, for in subareas II, V and VI most B₂ folds are very tightly appressed structures and few in number, whereas in subarea IV they are more open and a common occurrence. In subareas VIII and X the spread of bedding-slaty cleavage orientations can be attributed to the B₂ folding and not the co-axial B₁ folds.
The $B_2$ folds are largely responsible for the local redistribution of the bedding and slaty cleavage, but for most of the fault zone the surfaces $S$, $S_1$, and $S_2$ appear to be parallel because of the tight nature of the $B_2$ folds.

The $B_3$ folds vary considerably in plunge and are generally fairly steep as a result of the intersection of two steeply dipping surfaces the axial plane $S_3$ and the bedding-slaty cleavage. In subarea II, $S_3$ dips at approximately $80^\circ$ to $280^\circ$ whereas in the other subareas the dip of $S_3$ is approximately $80^\circ$ to $315^\circ$. The effect of $B_3$ folding on the orientation of the surfaces $S$ and $S_1$ is to destroy any cylindrical pattern developed by the $B_2$ folding and cause a greater spread in orientation. But the effect of the $B_3$ folding appears to be localized, and does not significantly alter the patterns resulting from the $B_2$ folding.

An examination of the total diagrams (Figure 4.4) of the fabrics elements in the Mount Isa Fault zone indicate that the majority of fold axes and mineral lineations recorded in this zone plunge $80^\circ$ to the south, and the bedding and slaty cleavage have been transposed parallel to $S_2$; which itself is parallel to the trend of the Mount Isa Fault zone. The $S_3$ axial planes on the otherhand are vertical and strike
FIGURE 4.4

(1) Total diagrams for the subareas II to VI within the Mount Isa Fault zone
(a) S parallel to $S_1$
(b) strain slip cleavage $S_2$
(c) cleavage $S_3$
(d) mineral lineation
(e) fold axes $B^2$
(f) fold axes $B_2^2$.

(2) Some representative readings recorded in the MicaF area of the Mount Isa Mine by G. Herget. (Unpublished data made available by Mount Isa Mines Ltd.) Most of the readings are from 9 Level between 6830 N to 6625 N, 11 Level along hanging wall of V72N drive from 7500 N to 7680 N, and 12 Level along T65 north drive hanging wall from 6550 N to 6630 N;
(a) readings on S and $S_1$
(b) axial planes of the prominent folds believed to be $B_2$
(c) $B_2$ fold axes.

All diagrams are plotted on schmidt stereographic nets on the lower hemisphere. Contoured according to Schmidt method; contour intervals are $\geq 1\%$ per 1% area, $\geq 3\%$ per 1% area, $\geq 6\%$ per 1% area and $\geq 9\%$ per 1% area, in all diagrams except figures 1d, 2a, 2b and 2c which have the contour intervals $\geq 1\%$ per 1% area, $\geq 5\%$ per 1% area, $\geq 10\%$ per 1% area and $\geq 15\%$ per 1% area. The number below each diagram is the number of points plotted.
to the north west, but as these folds postdate the major deformation in this zone they are not considered to represent a significant contribution to the strain undergone by the majority of the rocks in the Mount Isa Fault zone.

(ii) Fault I

This fault extends the whole length of the area and separates rocks of the biotite zone from the biotite-cordierite zone. Unlike the Mount Isa Fault it is not associated with a prominent zone of intense deformation, except for very localized areas. The fault plane is often difficult to recognise but in the vicinity of 1040 W 1170 S (See Plate 4.3a) two units of quite different orientations have been juxtaposed along what appears to be a very narrow joint plane. Along strike this widens to become a series of very closely spaced faults separating a zone, no more than 1 metre wide, of very tight B₂ folding (See Fig. 3.3g and h). Other surface evidence of this fault are a number of breccias which occur as small irregular augen shaped quartzite fragments set in a strongly foliated quartz-mica schist matrix. These are found intermitently along Fault I at 10000 N, 1000 S and 8500 S. In the vicinity of
(a) Fault I at 1040 W 1170 S (looking north). To the east of the fault are the westerly dipping foliated quartzites and to the west is the shallower dipping sequence of massive quartzites, which are closely jointed in the vicinity of the fault plane (Scale 2 metres)

(b) Strain slip cleavage in the muscovite-chlorite schists found in Fault I at 3500 N (Specimen 8075; Scale 0.1 mm; crossed nicols)
3500 N, there is an outcrop extremely soft, fine grained muscovite and chlorite rich rocks with a very fine $B_2$ cleavage structure (See Plate 4.3b). These occur in a zone some 5 metres in width and approximately 100 metres in length.

Except for these isolated pieces of evidence the boundary between the foliated quartzites in the east and the other westerly units is not distinct. There may however be a few small irregular quartz blows or veins paralleling the established position of this fault, as found immediately to the south of King Gully (see Figure 3.2).

On the macroscopic scale there appear to be quite significant differences in the geometry on either side of Fault I. In subarea 1 the slaty cleavage dips consistently to the west at approximately $80^\circ$, whereas the bedding dips east. In subareas 2, 3 and 9 both $S$ and $S_1$ have been redistributed by $B_2$ folding and portions of major fold closures appear to terminate along the line of the fault. This is especially noticeable in the Tailings Dam area where a quartzite and quartz-chlorite schist unit are terminated, and between Mica and Lena Creeks where second generation folds close onto a planar contact with the foliated quartzites.
The pronounced lineament created by the outcropping fault, in an area which varies by as much as 120 metres in relief suggests that the fault plane is almost vertical. This is further confirmed from the few limited observations that were made of the actual boundary. Along the western boundary from 9000 S to 000 many large open drag folds with an easterly or dextral sense of vergence outcrop. These appear to be second generation folds and indicate a downthrowing of the eastern block. It is therefore believed that this fault accompanied or immediately post-dated the second generation folding.

(iii) Fault II

Strong evidence for this fault exists immediately to the west of No. 3 Tailings Dam, where two massive quartzites and an interbedded quartz-muscovite schist are truncated. Also between cross faults f and e first generation fold closures have been truncated. Here the cleavage S₁ and the bedding S in the quartz-muscovite schist east of the fault abut against S and S₁ foliations of the adjacent muscovite schists at a very high angle. South of cross fault f there is no structural evidence for the existence of fault II. The fault has a curved
outcrop pattern paralleling the strike of a lithological boundary for a major portion of its length. Unlike the Mount Isa Fault, Fault I and Fault III (to be discussed later), this fault does not separate areas of quite different metamorphic assemblages nor does it truncate second generation folds.

North of No. 3 Tailings Dam, Fault II outcrops, along the mine meridian 1500 W and has previously been called the King Gully Fault (Heidecker, 1961). The actual fault plane in this area appears to be curved, it is not visible in outcrop except as a distinct lithological break between quartz-chlorite schist in the east and quartz-muscovite schists to the west. Adjacent to the eastern boundary of the quartz-muscovite schist units are a number of open first generation folds, in which the axial plane slaty cleavage has also been refolded. The eastern limbs of these folds can be traced into the lithological boundary, thought to be Fault II, at a high angle.

Although outcrop is very poor, in the vicinity of 2000 W 12000 N the boundary of the quartz-muscovite schist appears to trend almost east-west and dips in this unit are easterly, whereas the quartzites and quartz-chlorite schists to the south
of the boundary dip steeply to the west and appear to be overlain by this schist, suggesting the contact may be low dipping, and could represent either an unconformity or fault plane. The possibility of an unconformity seems unlikely for there is no evidence of erosion or any possible detritus along the boundary.

Also in this northern area the number of deformation phases recognised across the boundary appear to be the same with a prominent slaty cleavage on both sides which has been gently folded by second generation folds. It is believed that this boundary which from field mapping appears to be gently curved may be a fault which post-dates the $B_1$ folding in the area, but pre-dates the $B_2$. Therefore any such fault would have participated in the $B_2$ deformation resulting in a gentle curvature in the Fault plane. Such a folded fault has been described by Bailey (1910, p. 593) as a "slide" and it has subsequently been defined by Fleuty (1964) as a fault formed in close connection with folding. Tobisch (1967, p. 555) used the term "tectonic discontinuity" in preference to slide for such a feature as the latter may be misleading. The present author agrees with Tobisch and intends using the term tectonic discontinuity as
it does not have a genetic connection, and therefore would not be confused with gravitational gliding tectonics (de Sitter, 1956).

It is conceivable that many such faults or tectonic discontinuities exist in the Mount Isa area, but cannot be identified except where fortuitous outcrop patterns reveal their presence. Another tectonic discontinuity has been recognised in the Mount Isa Mine area (See Appendix B). This is the junction between the Western Volcanics and the Mount Isa Group, and this tectonic discontinuity probably outcrops over a fairly large portion of the area east of the Mount Isa Fault.

(iv) **Fault III**

This fault separates the Eastern Creek Volcanics and the most westerly outcropping Judenan Beds; it also represents the boundary between the biotite-cordierite zone rocks and the higher grade sillimanite zone rocks. The fault has a recognizable strike length of at least 19 kilometres, and has been offset in many places by a number of cross faults. Along the northern extension of the fault, a thin slither of quartz-muscovite schist separates the amphibolites from the quartzites, from 8000 S to 1600 N. This appears to be the only material evidence for the
existence of Fault III. From 23000 S southwards the boundary contains many lens shaped vein quartz deposits no more than 10 metres in length, but there is no demonstratable evidence of either brittle or ductile deformation having taken place along this boundary.

The truncation of a second generation fold closure in the vicinity of 23000 S and the western quartzite unit of the Judenan Beds just north of Lena Creek both suggest the presence of this fault. There is also an apparent divergence between the strike of the foliation in the Eastern Creek Volcanics and the bedding and slaty cleavage in the quartzite north from 6000 N. The trend of the foliation in the Eastern Creek Volcanics can be related to a large second generation fold structure. Therefore, it is concluded that this fault is related to the second generation folding.

Evidence for the continuation of Fault III between the cross faults f and h is almost negligible because of the paucity of outcrop in this area. The continuation seems highly probable and the appearance and truncation of the amphibolite and schist unit to the west of the fault indicates that the fault is dipping at a very high angle. This
angle would then have to be greater than the westerly
dip of the quartzites and must be almost vertical.

(v) **Cross Faults**

These have either an east-west strike or a west-
north-west strike and are generally accompanied by
very minor displacements. Fault a has previously
been named the No. 6 Tailings Dam Fault (Heidecker,
1961) and can be traced from 400 W 12000 N to 300 W
14200 N. The displacement of the foliated quartzites
appears to be the only evidence for the existence
of this fault.

Faults b, c and d are small faults displacing
Fault III and the Eastern Creek Volcanics.

Fault e is a very prominent fault on the aerial
photographs, but outcrop along the eastern extension
of this fault is very poor. The fault displaces the
$B_2$ deformation zone associated with the Mount Isa
Fault, and becomes the southern boundary for the
lenticular outcrop of Western Volcanics. Movement
appears to be north block down.

Fault f outcrops as two separate portions. In
the east it displaces all the rocks and strike faults
between the Mount Isa Fault and Fault II and appears
to be a continuation of one of the cross faults
mapped by Cordwell et al. (1961) to the east of the
Mount Isa Fault. The apparent movement on this portion of the fault appears to be north block down. The western extension of this fault is in an area of very poor outcrop, with Lena Creek following the strike of the fault for the greater part of its length. It displaces the Mount Guide Quartzites, Eastern Creek Volcanics and Fault III with an upthrow of the north block. The apparent movement of this fault could be described in terms of a pivotal fault (Hills, 1953).

Faults g and h are very minor cross faults and the amount of displacement cannot be accurately determined. All these cross faults are planar fractures and post-date all the folding in the area.

4. **Macroscopic Structure**

The area west of Mount Isa is divided into four narrow elongate and separate zones by prominent strike faults. This complicates any interpretation of the area as no formation is common to any block. However, a broad interpretation of the area is possible by consideration of the structure within each individual zone; this has been illustrated in the cross-sections (Map 3). For a complete interpretation of the area, and especially the rocks west of Fault III, it would be necessary to complete the mapping to the north-west, south-west, and east of the present area.
The rocks west of Fault III form the largest proportion of the area shown in Map 1. The Eastern Creek Volcanics and the Mount Guide Quartzites occur as two separate strips along the eastern and western edges of the Sybella Granite in the southern portion of the area, but join north of the last granite outcrop to become a continuous zone northwards, with the Mount Guide Quartzites flanked to the west by the May Downs gneiss. The general trend of this zone and the accompanying slaty cleavage is north-south, but swings west in its northern extremity into a large second generation fold closure. At a point five kilometres north of the area shown on Map 1 the outcrop pattern in the Eastern Creek Volcanics becomes rather complex (cf. Mount Isa Sheet F54-1 (1:253,440) which accompanies Carter et al., 1961). Here the \( S_1 \) foliation is still folded by the large second generation fold, and trends to the north west, whereas the macroscopic bedding lies at a high angle to this cleavage and strikes east in a refolded first generation fold. To the south-west of Map 1 the boundary between the Mount Guide Quartzites and the May Downs gneiss strikes west while the \( S_1 \) cleavage meets the boundary at a high angle.
Vergence zones can be recognized in $B_1$ folds in subareas 5 and 6 (cf. cross-sections, Map 3). Although many of these $B_1$ folds have been refolded it appears that the overall vergence in this group of rocks is sinistral or westerly and they lie on the western limb of a large $B_1$ synform which plunges to the south and closure exists to the north of the area shown on Map 1.

$B_2$ folds are developed on what was already a complex form surface at the end of $B_1$ deformation. The overturned appearance of many $B_1$ folds is due to $B_2$ deformation and therefore the dip of the enveloping surface is a $B_2$ feature. The large refolded fold in subarea 12 is interpreted as the hinge region of a $B_1$ syncline which has been subsequently modified by $B_2$ folding. The limbs of this fold are planar in subarea 13 but are refolded in subarea 14.

Thus, considering the area west of Fault II as a whole (See Figure 4.4), it appears that it is the western limb of a large $B_1$ syncline which is believed to plunge south at approximately $80^\circ$. The outcrop pattern of this structure has been modified by subsequent $B_2$ deformation and the eastern limb has been faulted out of the area by Fault III. This fault is believed to accompany the second generation
folding, but lacks the zone of intense $B_2$ deformation so characteristic of other second generation faults, such as the Mount Isa Fault. The Sybella Granite was emplaced during this $B_2$ deformation within the core of this large $B_1$ structure and the pegmatites post-date both the emplacement of the granite and the formation of Fault III.

The two zones between Faults II and II, and Faults I and II appear to be similar in many respects and include subareas 2, 3, 9 and 10. In contrast to the easterly dipping rocks west of Fault III the rocks in these two zones dip gently to the west. $B_1$ fold axes and the mineral lineation $L_M$ plunge in a northerly direction at approximately 30° and $B_1$ vergences are always dextral or easterly. The only significant macroscopic $B_1$ folds mapped in these two zones are those outcropping between Faults I and II in subarea 3. The asymmetry of many of these folds confirms the mesoscopic evidence that these units lie on the western limb of an anticlinal structure which has been modified by subsequent $B_2$ folding. The Fault II which subdivides the two zones (See Section 4.3) post-dates the $B_1$ folding, but has itself been refolded during the $B_2$ deformation, in a manner similar to the fault that exists at the base of the
Mount Isa Group (See Appendix B). No large or significant orientation differences exist in units on either side of this fault, except for local differences north of No. 3 Tailings Dam. It is therefore concluded that the rocks on either side of Fault II lie on the western limb of a shallow plunging $B_1$ anticline which has been modified slightly by the $B_2$ deformation.

The units between Fault I and the Mount Isa Fault are located on the western limb of a steeply, southerly, plunging first generation syncline. Little evidence exists for second generation folding, except for the area south of Mica Creek where these units have been folded into a north plunging syncline and truncated by the Mount Isa Fault.

The Mount Isa Fault is a major zone of intense second generation folding in which there has been a complete transposition of the pre-existing first generation structures. The zone is axial plane to a major northerly plunging second generation syncline, the Kennedy Syncline, and across this zone there appears to have been a large downward displacement of the eastern group of rocks relative to those in the west.
Directly east of the Mount Isa Fault outcrops the westerly dipping Mount Isa Group which rests on metamorphosed Eastern Creek Volcanics (Western Volcanics) the junction being an irregularly folded tectonic discontinuity (See Appendix B). A large proportion of the folds in the Mount Isa Group are second generation, have extremely variable plunges and form many dome and basin structures, but in the vicinity of the mine appear to plunge at approximately 30° to the north (See Figures 4.3 and 4.4). The grade of metamorphism across this tectonic discontinuity appears to be the same, but no structural information has as yet been obtained from beneath the Mount Isa Group.

The correlation of stratigraphy between these different fault blocks is only possible between the area west of the Fault III where the May Downs gneiss, Mount Guide Quartzite and Eastern Creek Volcanics are probably equivalent to the Leichhardt Metamorphics, Mount Guide Quartzite and Eastern Creek Volcanics respectively east of the Mount Isa Fault. As nothing is known of the structure in their eastern equivalents little may be said about their relationships to larger first generation structures nor can an intelligent guess be made of the relative movement that has occurred along the faults in the intervening area.
Detailed correlation both within the Judenan Beds, and of these to the Myally Beds is impossible for the Judenan Beds are internally faulted, as are both boundaries. A correlation may have been possible if there was more information about the Myally Beds but unfortunately little is known of their structure or internal stratigraphy and much of their outcrop underlies the tectonic discontinuity which marks the base of the Mount Isa Group. A correlation of the two tectonic discontinuities recognized in the area, Fault II and the base of the Mount Isa Group, is not possible at this stage of the structural investigation of the Mount Isa district.

5. **Regional Structure**

The area west of Mount Isa therefore consists of three fault blocks containing refolded limbs of a $B_1$, antiformal structure whose relationship to any larger $B_1$ fold structure is at present uncertain. Direct correlation across the Mount Isa Fault with the antiformal structure east and north-east of the Mine is only possible on broad stratigraphic grounds. The existence of tectonic discontinuities in the area, the extent of their development and numbers are still questionable. This factor further complicates
the regional picture. One such example is the existence of a tectonic discontinuity at the base of the Mount Isa Group (See Appendix B) which complicates the previously established stratigraphic sequence. Thus, the Mount Isa group may be equivalent to any one of the other pre-cambrian dolomitic units, which include the Gunpowder Creek Formation, Paradise Creek Formation, Surprise Creek Formation, Ploughed Mount Beds and Mingera Beds (Carter et al., 1961).

Carter et al. (1961) have interpreted sedimentation during Pre-Cambrian times, for the whole of the outcropping belt of Pre-Cambrian rocks in north-west Queensland, as two distinct north-south trending geosynclinal basins separated by a Geanticlinal high or "Tectonic Welt". The eastern and western margins of this possible geosynclinal belt are covered with gently folded or flat lying Upper Proterozoic and Mesozoic sediments, and also large tracts of alluvium. Therefore the width of this belt of geosynclinal sediments cannot be determined from present surface outcrops. Carter et al. (1961) have also related the sedimentation to distinct tectonic events, these being interpreted from the shape and modifications that occur in the present outcrop patterns (See Fig. 5, p. 46 of Carter et al.,
1961). The area in the immediate vicinity of Mount Isa is considered to be situated on the western flank of one of these depositional basins, the western geosyncline, and in an area of younger rocks occurring in a structural low between two basement highs. Both these marginal highs are believed to be composed of older basement metamorphics and granites which are absent in the intervening areas. The eastern geanticlinal high contains the Leichhardt Metamorphics and the Kalkadoon Granite, and the western basement high is supposed to contain the Yaringa Metamorphics and Sybella Granite.

In the light of the present structural investigation it is believed that the interpretation of Carter et al. (1961) is an oversimplification since the distribution of these pre cambrian rocks can equally as well be interpreted as a post-depositional tectonic structure. An examination of the regional maps (accompanying Carter et al., 1961) reveals a rather complex picture with many tight fold closures, curved and straight faults and quite complex outcrop patterns. This is further substantiated by field examination of the small scale structures in areas away from Mount Isa and especially in the areas of higher metamorphic grades, where folding and
metamorphic layerings become quite complex. The whole complexity of the area will not become apparent until further detailed structural investigations are carried out.

6. Structural Setting of the Quartzites at Mount Isa

The Mount Isa quartzites now outcrop in areas that apparently reflect different physical conditions that prevailed during the time of their deformation. The metamorphism in these areas varies from very low greenschist to amphibolite grade and there is generally strong evidence of either brittle or plastic deformation, and sometimes both. In the very low grade rocks approximately 150 kilometres due north of Mount Isa the quartzites exhibit little evidence of plastic deformation and most of the deformation appears to be by flexural-slip folding accompanied by brittle deformation. These Myally Quartzites have complex outcrop patterns, being intersected by many faults and joints; and the sediments themselves show few post-depositional changes with no penetrative mesoscopic or microscopic fabric elements (cf. Camooweal Sheet E54-13 (1:253,440) which accompanies Carter et al., 1961).
In the chlorite zone rocks immediately to the east of Mount Isa the penetrative deformation feature, slaty cleavage, which is characteristic of the first generation folding, becomes a dominant fabric element in the pelites; in the more competent rocks such as the quartzites it is generally poorly developed. West of the Mount Isa Fault there is a rapid increase in metamorphic grade which is accompanied by a marked change in the morphological appearance of the slaty cleavage, which at higher grades has as an equivalent structure, a strong differentiated layering.

Second generation deformation recognised in the Mount Isa area is either confined to narrow zones of intense deformation, such as the Mount Isa Fault, or occurs as large open flexures of the pre-existing first generation structures. The metamorphism accompanying this deformation is retrogressive being confined to narrow zones, and quartzites in such zones may now be regarded as mylonites (See Section 3.4.1). In other large second generation faults, Faults I and III, there is a conspicuous absence of the intense B₂ deformation zones so characteristic of the Mount Isa Fault.
These three faults, the Mount Isa Fault, Fault I and Fault III, divide the Mount Isa area into four structurally separate blocks. The eastern block contains the chlorite grade metamorphics with the westerly dipping Mount Guide and Myally Quartzites occurring beneath a tectonic discontinuity which marks the base of the Mount Isa Group (See Appendix B). The structural complexities within this block are still unknown, but the major strike of these units appears to be to the NNE with a gentle swing to the east, to the north-east of Mount Isa, in a large second generation fold (See Map 2).

The sequence of biotite grade foliated quartzites and schists between the Mount Isa Fault and Fault I occur on the eastern limb of a south plunging B₁ structure, whereas the quartzites between Faults I and III lie on the western limb of a B₁ structure modified by B₂ deformation. Separating this area into two portions is the tectonic discontinuity Fault II which also dips gently to the west.

West of Fault III the prominent Mount Guide Quartzites lie on the western limb of a southerly plunging synclinal structure. These quartzites belong to the amphibolite facies and in many places are extensively refolded by the B₂ deformation.
The correlation of the structural elements recognized in the Mount Isa area across these different blocks and with increase in metamorphic grade assumes that the sequence of deformation in the low grade areas is comparable with that observed in the higher grade regions; and that differences in the morphology of the cleavages are dependent on both differences in inherited fabric and on the prevailing conditions of deformation.

As the majority of these Mount Isa quartzites are extremely pure and any marked compositional or structural variation within them is minimal (unlike the pelitic rocks which vary markedly in both composition and type of inherited fabric element), they should serve as good indicators of the deformation conditions. From the foregoing descriptions, in which most attention was paid to the schists, only limited information can be obtained as to how the deformation of the Mount Isa area was achieved. But by a study of these quartzites and by using the structural information with a knowledge of the metamorphism and strain, it is hoped in Part II of this theis to arrive at some conclusions regarding the physical processes active during the deformation.