A STUDY OF THE STRUCTURAL AND METAMORPHIC RELATIONSHIPS BETWEEN OLDER AND YOUNGER PRECAMBRIAN ROCKS IN THE MOUNT LOFTY - OLARY ARC, SOUTH AUSTRALIA.

by

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All specimens referred to in this thesis are catalogued in the museum of the Geology Department of the University of Adelaide.. The catalogued information includes the name of the rock, the catalogue number and the approximate location. The exact location is given as a national grid reference for the Paracombe-Mt. Gawler area and is shown on Plate IV for the Whey Whey Creek area.

SUMMARY.

The relationships between Upper Precambrian and "Archaean" rocks have been studied in two small areas in South Australia. In both areas the Upper Precambrian rocks consist of a sequence of metamorphic rocks showing many original sedimentary features. The underlying "Archaean" rocks consist of a complex of schists and gneisses of unknown age. In both areas the present metamorphic grade of the "Archaean" rocks is the same as that of the overlying strata. In one of the areas (near Adelaide), the grade is that of the chlorite zone of the greenschist facies, and in the other area (in the Olary Province), the rocks have been metamorphosed to the biotite zone of the greenschist facies. The "Archaean" rocks show relic minerals from a higher grade of metamorphism, and the present grade of metamorphism results from a period, or periods, of retrograde metamorphism.

The relic minerals of the "Archaean" of the Adelaide region indicate an upper amphibolite facies of metamorphism, and the original complex consisted of garnet, sillimanite and andalusite schists with limited outcrops of microcline - oligoclase-diopside gneisses. The gneisses are probably metasomatised igneous or sedimentary rocks. The majority of the rocks, however, are now blastomylonites, possessing a strongly developed foliation which has obliterated the earlier structures. The schists have been altered to aggregates of quartz and sericite, with relic higher grade minerals, and the feldspar-diopside gneisses have been

altered to microcline-albite-actinolite assemblages. Pebbles in the overlying conglomerates indicate that some of the retrograde effects are pre-Upper Precambrian in age.

The Upper Precambrian rocks consist of a sequence of phyllites and interlayered dolomites and quartzites. The effects of two phases of folding are well displayed in these rocks. An earlier series of asymmetrical structures associated with a slaty cleavage are overprinted by symmetrical folds related to a crenulation cleavage. The foliation in the "Archaean" rocks is subparallel to the slaty cleavage in the younger sequence and the deformation of both sequences appears to be related.

In the second area investigated (in the Olary Province near Weekeroo) mica schists, dolomites and quartzites overly a basement complex of mica schists, gneisses and granites. Relics of a higher grade of metamorphism in the older rocks are rare, but sillimanite and staurolite occur in the mica schists. The metamorphic grade in both sequences is now that of the biotite zone of the greenschist facies. The equilibrium fabrics in these rocks are in marked contrast to the disequilibrium fabrics in the "Archaean" in the Adelaide region. These differences are attributed to the differences in conditions of the retrograde metamorphism in the two areas.

In the Weekeroo area, the structure in the basement rocks is simple; a vertical east-west foliation and layering dominate the structure. The structure in the overlying

Upper Precambrian rocks is more varied, and a greater complexity of folding is observed in those rocks which overly mica schists of the "Archaean". The more intimate relationships between the structures in the two sequences (compared with the Adelaide region) is ascribed to the markedly anisotropic nature of the basement rocks in the Weekeroo area.

This thesis contains no material which has been accepted for the award of any other degree or diploma in any University and to the best of my knowledge and belief, the thesis contains no material previously published or written by another person, except where due reference is made in the text of the thesis.

Dec 15. 1962

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This work was undertaken at the initial suggestion of Prof. A. R. Alderman. The author has discussed various aspects of the field work in the Houghton region with Dr. R. W. Nesbitt and Mr. R. Offler of the department of Geology, the University of Adelaide, Mr. B. E. Hobbs of the University of Sydney and Dr. H.G. Wilshire formerly of the Australian National University. Sections of the petrology have been discussed with Drs. R.L. Oliver, A.W. Kleeman, R.W. Nesbitt and J. B. Jones. The structure has been discussed at meetings of the Geological Society of Australia (South Australian Branch) and at a conference on structural analysis held at Canberra in 1962.

The author was accompanied in the Whey Whey creek area at different times by Messrs. K. J. Mills and A.Kirk and Dr. R. W. Nesbitt. Many fruitful discussions of the petrology of the area have been held with Drs. Oliver and Kleeman.

Drs. Oliver and Kleeman read and criticized the petrological sections on the Houghton Complex. Dr. B. Daily read and criticized the Introduction and Chapter I. Prof. A. R. Alderman read and criticized the whole manuscript.

Mr. J. Lorenzin prepared the thin sections. Miss B. Moore prepared the photographed figures. Miss M. Swan advised and assisted with the layout of the maps and line drawings. Mrs. J. L. Talbot typed and corrected the preliminary draft.

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INTRODUCTION.

Deformed rocks of Upper Precambrian age are exposed over large areas of South Australia. These rocks lie with marked unconformity on a "crystalline basement" generally considered to be Archaean in age.

Although the stratigraphy of the Upper Precambrian is well known, very little information has been published concerning the relationships between these rocks and the underlying Archaean. A reconnaissance of the areas in central South Australia, where the contacts between the two sequences can be observed, showed that the two main regions where the Archaean crops out have different metamorphic and structural relationships.

These two regions are:

(a) The Adelaide Region (see Fig. 1).

Isolated inliers of highly retrograde schists and gneisses of the Archaean are overlain by phyllites, quartzites and dolomites of Upper Precambrian age.

(b) The Olary Province (the Archaean complexes north and east of Mannahill on Fig. 1 and extending beyond the confines of that map).

In this region coarse-grained mica-schists, gneisses and granites of the Archaean are in contact with biotite schists, quartzites and dolomite marbles of Upper Precambrian age.

A small area from each of these regions was chosen for study in order to establish the structural and metamorphic

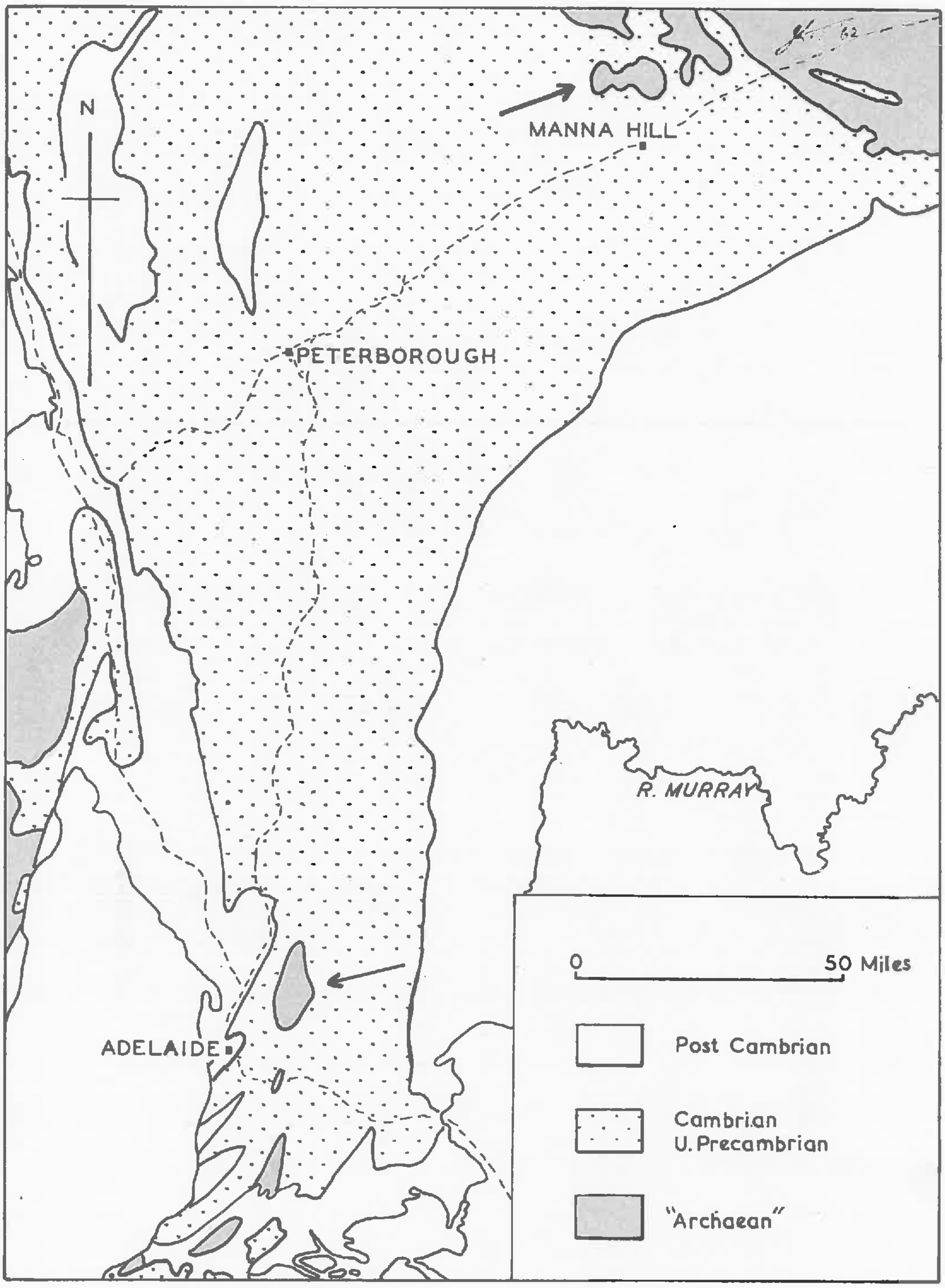


FIG 1

Locality map of Central South Australia. Arrows show approximate locations of areas studied.

history of each area, and to determine the influence of the Archaean on the deformation of the Upper Precambrian.

GEOGRAPHICAL LOCATION OF THE AREAS STUDIED

The two areas studied are; (a) an area about 10 miles northeast of Adelaide referred to as the Paracombe - Mt. Gawler area and (b) an area about 25 miles north of Mannahill referred to as the Whey Whey Creek area.

(a) Paracombe -Mt. Gawler area.

The general location of the Paracombe area is shown in Fig. 1 and in more detail in relation to Adelaide in Fig. 2 and 3. The area lies within the region of a heavily dissected Tertiary plateau. Its general level is about 1200-1300' with a maximum elevation (in the area mapped) of 1779' at Mt. Gawler and a minimum elevation of just under 400' at the mouth of the Torrens Gorge. The area is crossed by many deep creeks most of which flow only in winter although some, notably the River Torrens and its tributary at Castambul, Sixth Creek, are said to flow even in drought. The Little Para River flows in all but the driest of summers. In spite of the deep dissection, exposures are generally poor and the area is characterized by steep soil-covered slopes with disconnected outcrops of quartzites and more rarely dolomites and phyllites.

Excellent exposures are found in cuttings on many of the major roads and a study of these road cuts supplied much of the information regading small scale structures.

(b) Whey Whey Creek Area, Weekeroo Station.

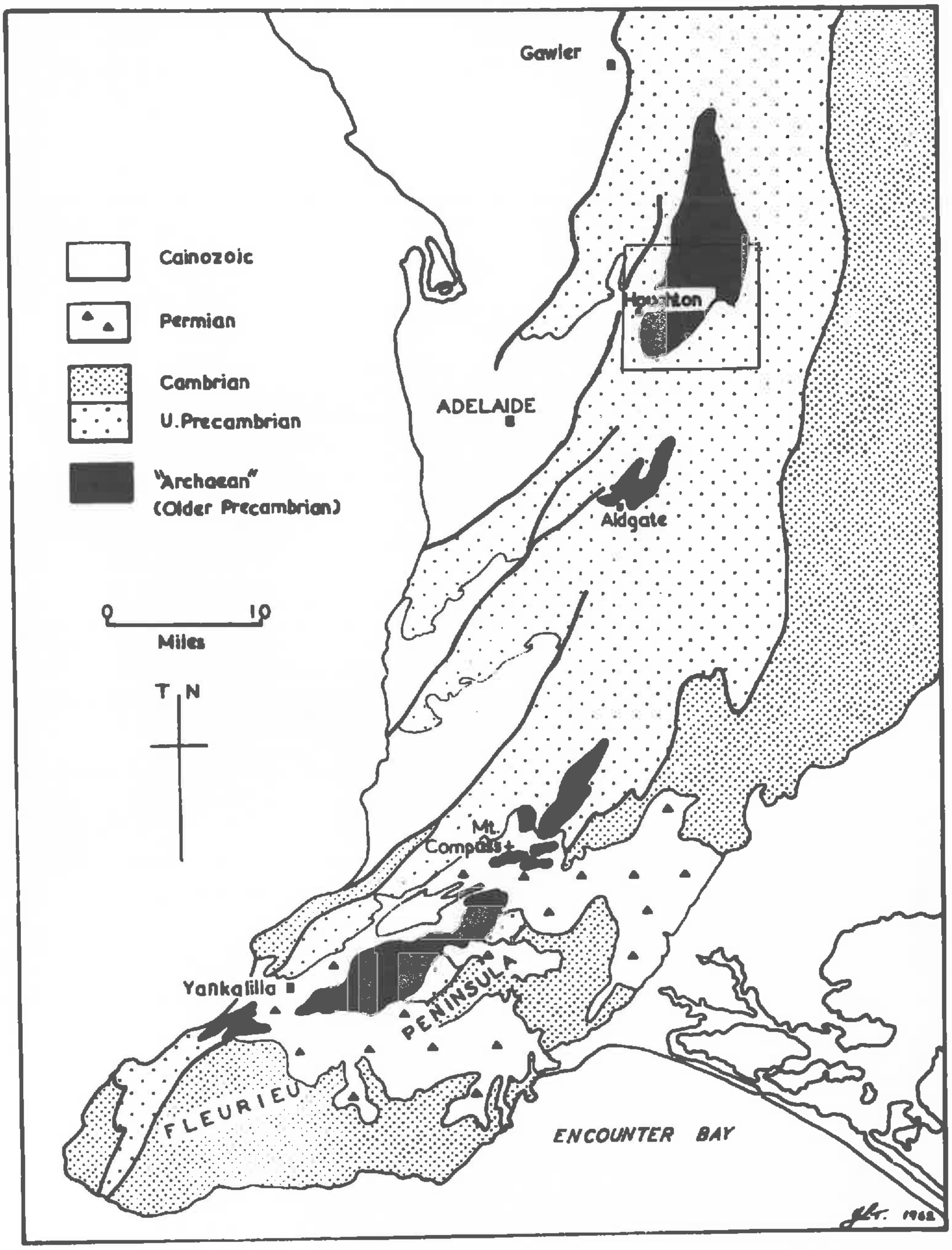


FIG.2

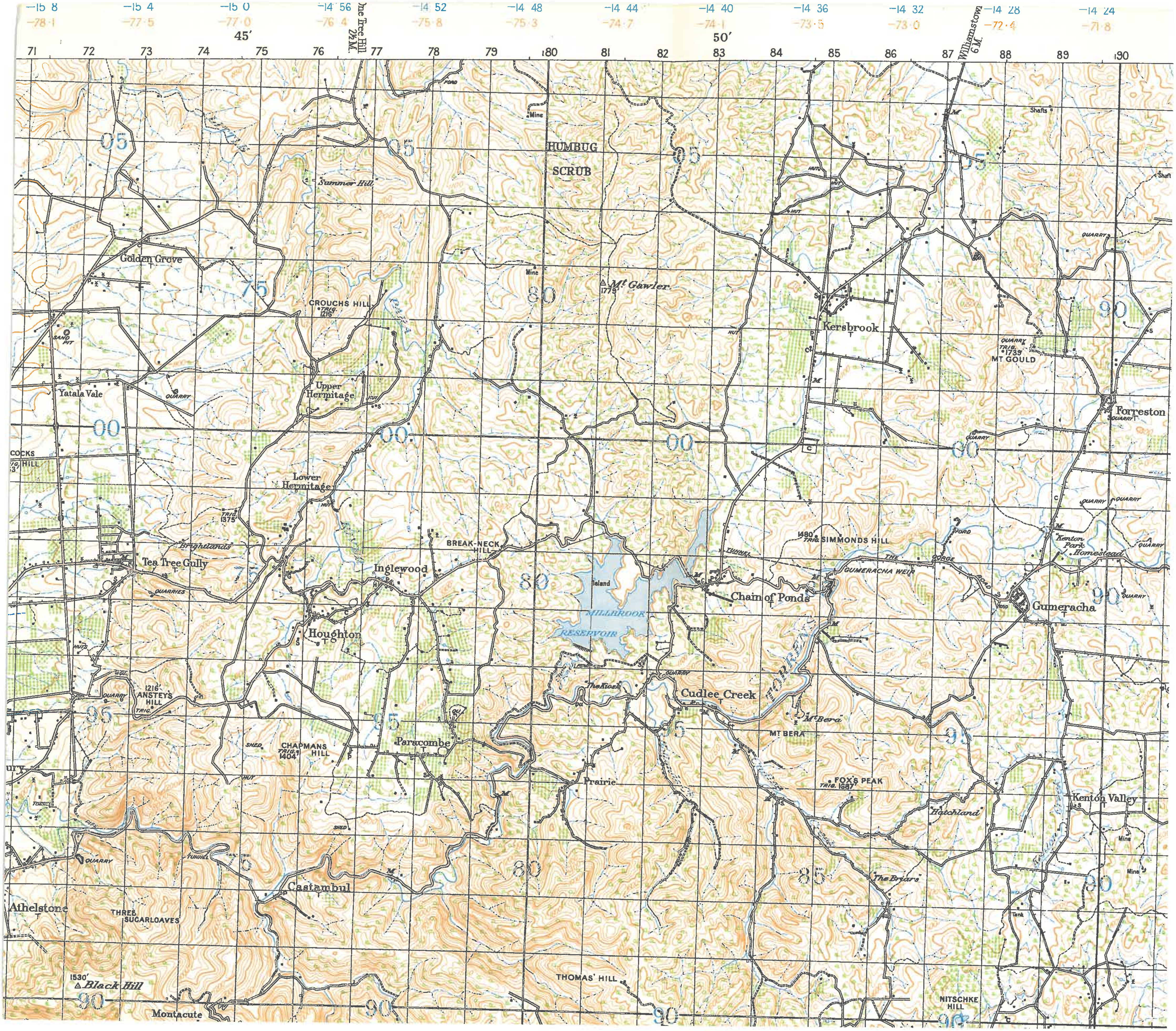
Locality map of Adelaide region.

Fig. 3

Part of Adelaide 1 mile military sheet. Grid north is $1^{\circ}17'$ east of true north.

Scale: 1 inch = 1 mile.

For grid references read east before north; e.g., the island in Millbrook Reservoir has the reference 811 976.



45′
50′
HUMBUG
SCRUB
Summer Hill
Golden Grove
CROUCHS HILL
Mt Gawler
Kersbrook
MT GOULD
Yatala Vale
Upper
Hermitage
Forreston
COCKS
Lower
Hermitage
SIMMONDS HILL
Tea Tree Gully
BREAK-NECK
HILL
Inglewood
GUMERACHA WEIR
Kenton Park
Homestead
Chain of Ponds
Gumeracha
MILLBROOK
RESERVOIR
Houghton
ANSTEYS
HILL
The Knoll
Cudlee Creek
Mt Bera
MT BERA
CHAPMANS
HILL
Paracombe
Prairie
FOX'S PEAK
Hatchland
Kenton Valley
Castambul
The Briars
Athelstone
THREE
SUGARLOAVES
Black Hill
THOMAS' HILL
NITSCHKE
HILL
Montacute

The second area investigated lies in semi-arid sheep-grazing country about 220 miles north northeast of Adelaide and 11 miles northwest of Weekeroo Head Station (Fig. 4). Access to the region is quite good, the area lying just north of Mannahill which is on the Adelaide-Broken Hill road; but access within the area is limited to a few station tracks.

The relief of the region is much more subdued than in the Paracombe region, being no more than 400' in the area mapped but exposures are generally good due to the arid climate. Differential erosion is very marked and the general outline of the structure is readily discernible on aerial photographs. Furthermore, the vegetation differences on the various rock types are very striking. Species of Casuarina are dominant on the Archaean rocks, with eucalypts on the Upper Precambrian rocks. The upper parts of the Upper Precambrian sequence support very little vegetation.

Large areas of alluvium are associated with many of the creeks (all of which are dry except after the infrequent heavy rains); these deposits and associated pediments have not been delineated accurately, but their broad distribution can be seen in Fig. 4.

SUMMARY OF PREVIOUS INVESTIGATIONS.

In view of the wide geographical separation, the two areas will be treated separately.

(a) Paracombe - Mt. Gawler Area.

Much of the earlier work in this region was of general

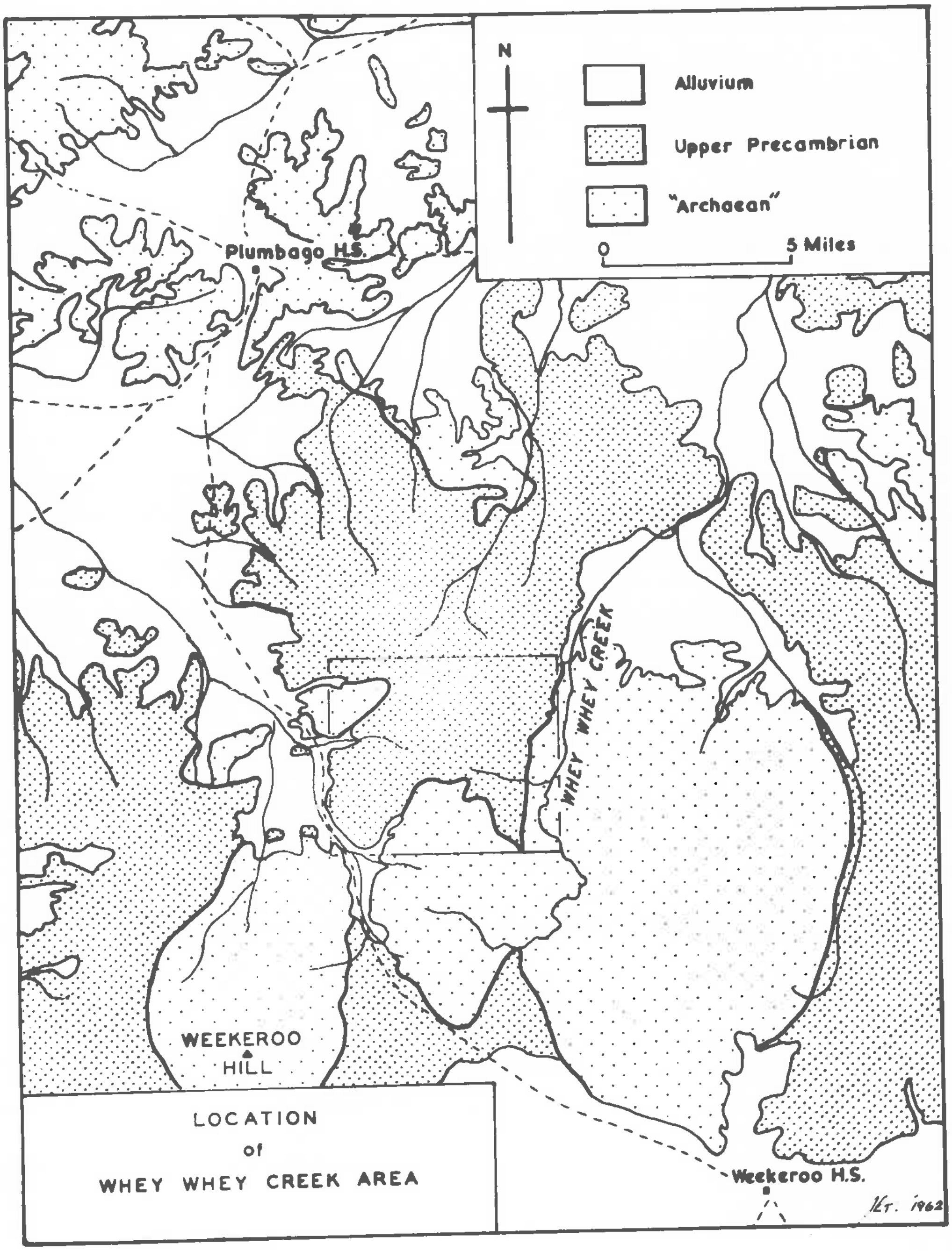

N
Alluvium
Upper Precambrian
"Archaean"
0
5 Miles
Plumbago H.S.
WHEY WHEY CREEK
WEEKEROO HILL
Weekeroo H.S.
LOCATION of WHEY WHEY CREEK AREA
1962

FIG. 4

geologic and economic interest, and dealt mainly with stratigraphic relationships. On the first geological map of South Australia (Brown, 1886) all the metamorphosed rocks of the Mt. Lofty Ranges in the Adelaide region are grouped as one sequence. However Howchin's work (1904, 1906) established that a low grade metamorphic sequence of "Lower Cambrian" age overlies a "Precambrian Complex (Archaean)".

The metamorphic rocks on the eastern side of the Mt. Lofty Ranges are of a much higher grade than those in the area studied by Howchin and on this account Woolnough (1908) considered that they were part of a basement sequence. He Proposed the name "Barossa series or Barossian" for these rocks and thought that they showed similarities to the basement rocks of Yorke and Eyre Peninsulas. Benson (1909) correlated the Barrossian Series with the older sequence of rocks in the vicinity of Houghton and Paracombe. However, Howchin (1926) demonstrated that Woolnough's "Barossian" is the metamorphosed equivalent of the "Cambrian" sequence of the western part of the Mt. Lofty Ranges and so reverted to calling the older sequence the "Archaean or Fundamental Complex (or Houghtonian)", the last name denoting the complex at Houghton. Hossfeld (1935) disagreed with Howchin's correlation, equated the Houghtonian with the Barossian, and suggested that both be called Barossian as this name had precedence. However, subsequent work has confirmed Howchin's correlations (e.g. Sprigg and others 1951; Campana, 1955; Kleeman and White, 1956), although many of the beds to the east of the ranges are younger than the beds on the west.

Much has been written concerning the ages of the two sequences. All of the rocks were originally called Lower Silurian (Brown and Woodward, 1885) on the basis of lithologic similarities to the rocks of the Victorian Goldfields. The discovery of Cambrian fossils south of Adelaide (Howchin, 1897) and the division of the rocks into two sequences (Howchin, 1904) led to a revision of these ideas. Howchin (1906) considered all of the rocks down to the basement complex as Lower Cambrian. However, David (1922) suggested that all the beds below the lowest fossiliferous beds be grouped as the Adelaide Series and a tentative Proterozoic age be assigned to them. In 1927 he concluded that the Adelaide Series might be Lower Cambrian, but later assigned them to the Proterozoic (David 1928). Howchin (1929) placed the Precambrian - Cambrian boundary of the top of the Brighton Limestone, still some distance below the lowest fossiliferous beds.

Mawson (1948) was the first to use the term Adelaide System to include all local rocks of Upper Precambrian age. Mawson and Sprigg (1950) defined the system more precisely and subdivided it into three series: the Torrension, Sturtian and Marinoan Series. The Marinoan was defined as including all the beds above the Brighton limestone up to the first fossiliferous formation of the Cambrian.

Daily (in press) has suggested that in view of the recommendations of the Australian Code of Stratigraphic

Nomenclature (Anonymous 1958) the terms Adelaide System and Torrensian (etc.) Series be not used as they have an unwarranted time significance. He suggested that the terms Torrens Group, Sturt Group and Marino group be used to replace the three "Series" of the younger precambrian sequence and if it is necessary to refer to the three groups together they should be called the Adelaide Supergroup. This suggestion has been accepted for the purposes of this thesis.

The age status of the older sequence is even more unsatisfactoy. The rocks are locally referred to as Archaean (e.g. Campana, in Glaessner and Parkin, 1958) although unsupported by age determinations. A phase of mineralization in related rocks at Normanville has been dated at approximately 1000 million years (Thomas, 1924). Hence until more definite dates are available use of terms with a time significance is not justified. It is therefore proposed to refer to the rocks of the Houghton-Barossa inlier as the Houghton Complex, using Howchin's "Houghtonian" in a manner prescribed by the Australian Code of Stratigraphic Nomenclature (op. cit.). The term Houghton Complex is chosen in preference to Barossa Complex because the term "Barossian" was originally used for rocks now believed to be Cambrian or younger in age.

Work of a specifically petrographic nature concerning the Houghton region is less abundant. Significant contributions have been made by Benson (1909), England (1935), Alderman (1938) and Spry (1951).

Studies of the structure of the ranges have been limited to considerations of the large scale distribution of rock units, the main contributions having been made by Howchin (1929), Sprigg (1945), Webb (1953) and Campana (1955.). These petrologic and structural studies will be reviewed in more detail in the later sections of the thesis.

(b) The Whey Whey Creek Area, Weekeroo Station.

Published geolgical investigations in the Weekeroo area have been more limited, probably owing to the remote nature of the area, and until recently the lack of adequate base maps. Mawson (1912) studied a very large area in the Broken Hill region and recognised two unconformable sequences of rocks; the older sequence of schists and gneisses he named the Willyama Complex (Willyama is the aboriginal name for the Broken Hill environs); the younger sequence of "sedimentary rocks" he called the Torrowangee Series, noting their similarity with the younger sequence in the Adelaide region. Although Mawson's mapping finished a few miles east of the Weekeroo region, the lithological similarities of the Weekeroo rocks to those described by Mawson is such that the terms Willyama Complex and Torrowangee Series can be extended to these rocks. Campana and King (1958), however, referred to the Willyama Complex as Archaean, and equated the younger rocks with the Torrensian and Sturtian Series of the Adelaide region.

For the purposes of this thesis it is proposed to use the rock-stratigraphic names Willyama Complex for the older sequence, and Torrens Groups and Sturt Group for the younger sequence. By analogy with the Adelaide region the younger sequence is considered to be Upper Precambrian in age. The age of the Willyama Complex at Weekeroo is unknown but the oldest phases of mineralization in the Olary and Broken Hill provinces are all approximately 1500-1600 million years; Crocker Well 1,600; Radium Hill 1,510; Broken Hill 1,500 million years (Wilson and others, 1960). These dates suggest an Archaean age for the Willyama Complex.

Other publications dealing with the Weekeroo area are Campana (1955), which considered the broad structural features of the Mount Lofty-Olary arc; Campana, 1956 (Geological Atlas of South Australia, Plumbago 1 mile sheet); and Jones, Talbot and McBriar (1962) which discussed the origin of the major amphibolites in the area.

CHAPTER I.

THE HOUGHTON COMPLEX.

Introduction.

The older complex of schists and gneisses, called the Houghton Complex, occurs as a lozenge shaped mass extending from the River Torrens in the south to the Barossa Reservoir in the north (See Figs. 2 and 3). The complex is about 17 miles long (in a north - northeast direction) and about 5½ miles wide at the widest part. Much of the area forms part of a Tertiary erosion surface, particularly north of Mt. Gawler and the basement outcrop is very sparse. Only the southern, somewhat better exposed regions have been mapped, the general rock distribution being shown on Plate I.

Until fairly recently the rocks of the Houghton Complex have received scant attention. In early papers (e.g. Brown and Woodward, 1885) they were referred to as a complex of highly metamorphosed rocks. Howchin (1906) refers to them as the Archaean complex of the "central axis of the Mt. Lofty Ranges" underlying the "Cambrian" (now considered Upper Precambrian) with marked unconformity. His comments are restricted to field observations and he concludes that the complex is a sequence of metamorphosed sedimentary rocks which have been intruded and altered by "pegmatization" along the cleavage.

Benson (1909) was the first to describe the rocks of the complex in any detail. He referred to the augen-gneisses of Barossa and Humbug Scrub and described the schists and gneisses as well as "intrusive" diorites and syenites. Much of his paper is confined to description of the "diorites" and related rocks in the immediate vicinity of Houghton. These rocks, which he considered igneous, were compared very briefly with similar rocks at Aldgate and Yankalilla. He concluded that "so widespread are the rocks... throughout South Australia that the State may be considered as a petrographic province, the characteristic feature of which is a high percentage of titanium acid, to a less degree the abundance of soda." He called the "magma" the "Houghton magma" but noted that "the majority of the rocks, however, has a marked gneissic banding" and also that some of the schists appear to be altered diorites.

Mawson (1926) noted that the rocks derived from the Houghton magma are distinctly monzonitic in character. England (1935) extended Benson's observations and studied all the "dioritic" rocks in the Mt. Lofty Ranges. He followed Benson's interpretation of their igneous nature and published 11 analyses of them, classifying them according to the C.I.P.W. classification. Unfortunately only six of his analyses are from rocks which can be correlated with the "Houghton igneous suite", analyses 9 - 11 being of a much younger suite and analyses 5 and 6 being of intrusive dolerites. However, England's study established the widespread occurrence of these unusual "dioritic" rocks which can be found from 4 miles south of Yankalilla to Houghton, a distance of about 60 miles.

Hossfeld (1935) in a regional study of the northern Mt. Lofty Ranges concluded that the "Archaean" gneisses (not the Houghton magma-type rocks, but the surrounding metamorphics) were formed by a process of lit-par-lit injection of pegmatitic material and that the common augen gneisses of this region were the result of later crushing of the pegmatitic material.

Alderman (1938) in a study of these augen gneisses noted the high content of soda feldspar and concluded that they were formed by a process of metasomatism, soda and silica having been the chief constituents added.

Spry (1951) produced a reconnaissance map of the Houghton complex and gave petrographic descriptions of the main rock types in the complex. His paper is concerned largely with the origin of the "diorite" and he concludes that it is a soda-metasomatized sediment rather than an igneous rock. Webb (1953) as a result of a more detailed mapping of the " diorite" in connexion with uranium exploration supports Spry's conclusions and divides the rocks of the complex into "strongly albitized metasediments","partially albitized metasediments" and "non albitized metasediments".

Although the petrology of the rocks of the complex has excited some considerable interest from time to time, the structure of the complex has been largely ignored. This is not surprising in view of the poor outcrop, general lack of marker horizons and the widespread later changes which have largely obliterated the original structures. Early reports dismiss the structure as "complex" or "highly folded" and the only serious attempts to unravel the structural history are those of Spry (1951) and Webb (1953), whose observations and conclusions will be discussed later.

Classification of the rock types of the complex.

The diversity of views on the origin of the various rocks of the complex has resulted in a variety of rock terms and classifications, but briefly the rocks can be divided into a highly feldspathic suite occurring around Houghton and east of Kersbrook, and a micaceous suite occurring in the remainder of the complex. The rocks of the first named suite are characterized by acid plagioclase with actinolite and/or diopside and accessories. They have a more or less well-developed layering or lineation due to the segregation and parallel arrangement of elongate actinolite prisms. They have been called syenites and diorites by Benson (1909), granulites by Spry (1951) and albitized metasediments by Webb (1953). The present author prefers the somewhat more general term gneiss to describe them.

There are further difficulties in nomenclature with regard to the micaceous suite. Superficially, the rocks resemble normal schists and gneisses but closer studies reveal that many of them are "cataclasites". In many cases the rocks are completely recrystallized, and Sander's term blastomylonite could be applied to them (Knopf, 1931; Christie, 1960). True phyllonites (Knopf, 1931) are rare and are observed only as narrow zones within the feldspar gneisses. Some of the rocks are well layered and are similar to the flaser gneisses described by Hamilton (1960).

It will be demonstrated later that these changes are manifestations of retrogressive metamorphism which has changed the rocks from the amphibolite facies to the lower green-schist facies. This phase of retrograde metamorphism has resulted in a considerable amount of neomineralization (Knopf, 1931) which has obscured many of the cataclastic textures in the rocks.

CHAPTER II.

THE PETROGRAPHY OF THE FELDSPAR GNEISSES OF THE HOUGHTON COMPLEX.

Feldspar gneisses occur in three main regions: 1) the area around Houghton and Inglewood (the type area for the "Houghton magma" rocks), 2) the area east of Kersbrook and 3) the Paracombe-Torrens Gorge region (See Plate I).

In all three areas the rocks show marked lithological similarities. The Paracombe and Houghton areas may in fact be parts of the same mass; however, they are separated by an erosion surface with deep lateritic soils. The Kersbrook mass is separated from the Houghton mass by an area of retrograde rocks the original mineralogical composition of which was quite different from that of the rocks of the two feldspar gneiss masses.

The gneisses of the Torrens Gorge are very different mineralogically from the Houghton mass but are discussed in this chapter as many of them are highly feldspathic. It is believed that they have an origin different from that of Houghton rocks.

The feldspar gneisses were studied in some detail as they represent the largest occurrence of rock types displaying pre-retrograde as well as retrograde mineralogy and textures. By contrast, in the majority of the rocks of the complex retrogressive metamorphism has largely obliterated earlier features.

Over 100 specimens of these gneisses have been collected and examined under low power magnification and thin sections have been prepared of each specimen. The mineralogy of the gneisses is relatively simple and observations have been directed towards gaining an understanding of the textural features of the rocks.

The gneisses of the Houghton-Inglewood area are described first and the other occurrences compared with them.

1. GNEISSES OF THE HOUGHTON-INGLEWOOD DISTRICT.

(a) Gross structural features.

Many of the rocks are well layered. The layers may be diffuse or sharp, (Fig. 5), and may be quite planar and regular (Fig. 6). The layering is commonly due to variations in the relative proportions of dark (actinolite and/or diopside) and light (feldspar) minerals. In many cases a gross layering is not observed and only the preferred orientation of the actinolites gives an overall foliation or in some cases lineation to the rock (Fig. 7). In other cases the layering is due to concentration of iron oxides into distinct planar layers, in which case the rocks look very much like banded quartzites (except that in thin section they are seen to be almost entirely of plagioclase with minor microcline and opaques). Cross cutting non-dilatational veins of pure feldspar may produce a secondary layering in places but is generally quite distinct from the regional layering (fig. 8).

Fig. 5. Layering in feldspar actinolite gneiss due to variations in the actinolite to plagioclase ratio.
Houghton. (Grid. ref. 772 965).
Metric scale.

Fig. 6. Regular laminations in feldspar gneiss. The darker laminations are rich in actinolite.
Inglewood. (Grid. ref. 773 977).

FIG. 5

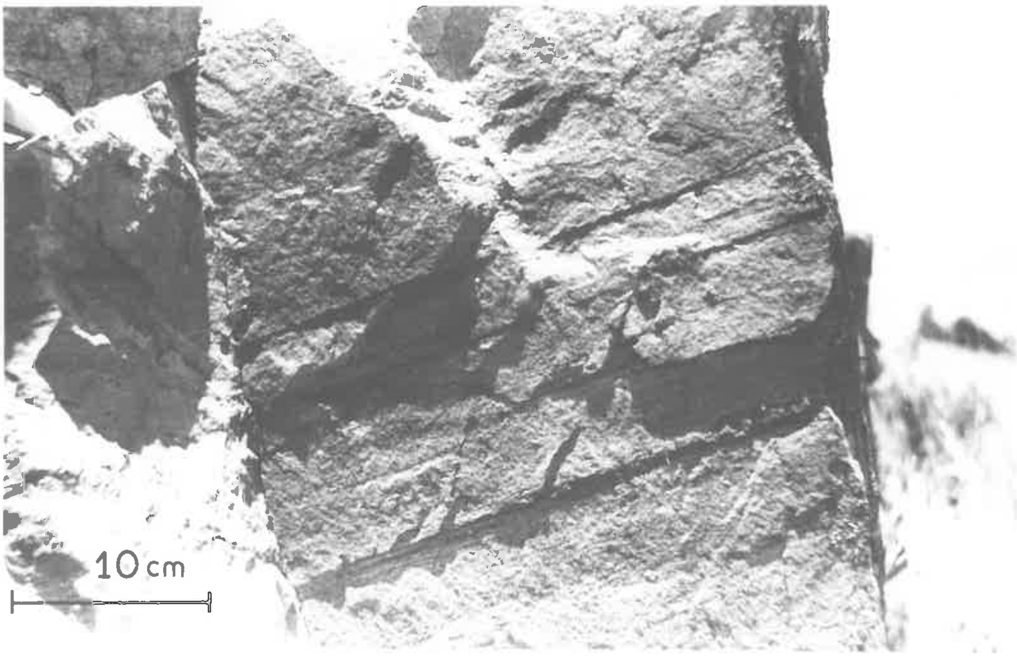


FIG 6

Fig. 7.   Foliation in feldspar gneiss due to preferred orientation of actinolite prisms.
Houghton. (Grid. ref. 772 965)

Fig. 8.   Vein of fine-grained feldspar in feldspar gneiss.
Inglewood. (Grid.ref. 773 977)

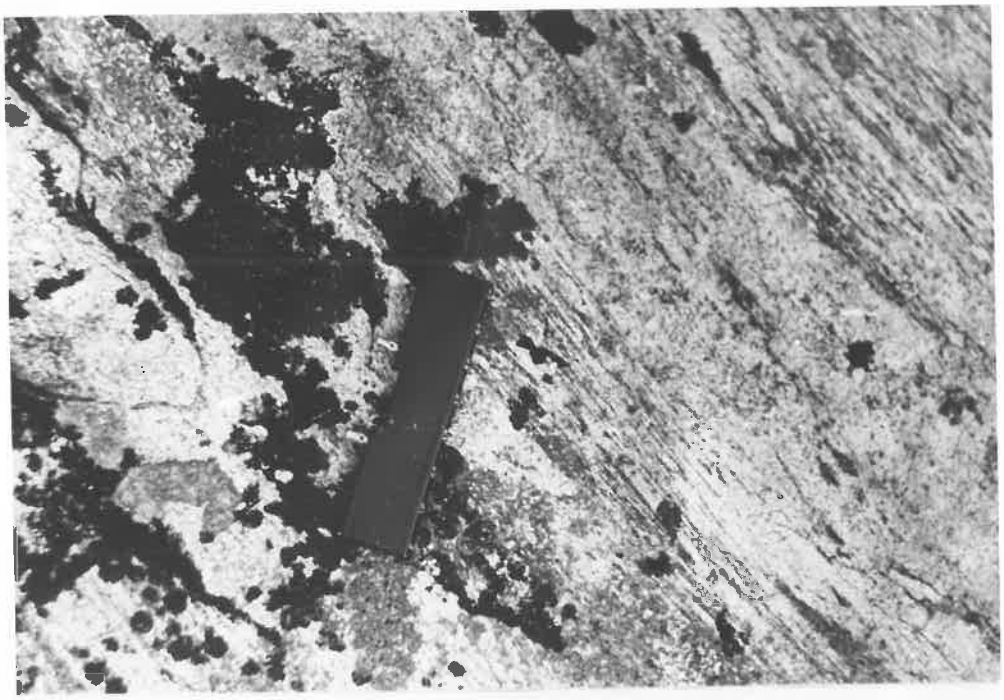


FIG. 7

FIG. 8

Layering due to differences in grain size or variations in the amount of quartz is rare in the main body of the feldspar gneiss but is important in the marginal areas where it may represent original compositional differences.

In the quartz-rich feldspar gneisses near Houghton a foliation due to flattened quartz spindles and lenticles resembles true granulitic fabric (Williams, Turner and Gilbert, 1955, p. 236); the preferred orientation of the spindles imparts a strong lineation to the rock. The feldspar gneisses in general do not show a strong lineation although as previously mentioned actinolite prisms may impart a fair lineation in rocks which are poorly layered.

The layering in these gneisses has been interpreted as mimetic after bedding (Spry, 1951: Webb, 1953) and structures resembling cross bedding have been used to support this hypothesis (Spry, 1951, Plate VII, Fig. 3). Although this interpretation may well be correct it is thought that such structures - which are rare - could be the result of transposition of an earlier layering.

(b) Petrographic features.

The dominant minerals in the gneisses are feldspars (plagioclase commonly more abundant than microcline perthite), actinolite and/or diopside, with minor quartz, epidote and opaques. The variation in the proportions of these minerals is considerable; rocks composed essentially of plagioclase are fairly common, but rocks with as much as 30% dark minerals also are common. Similarly the content of microcline and quartz is very variable, a few specimens being almost pure quartz and microcline.

For convenience of description the feldspar gneisses have been divided into three groups: 1. Feldspar gneisses with significant pyroxene or amphibole, 2. Feldspar gneisses with accessory or no dark minerals and 3. Quartz-feldspar gneisses. Such a subdivision is quite arbitrary and does not reflect any distinct geographical distribution, although the gneisses of Group I are the dominant gneisses of the area north and east of Houghton, and gneisses of Groups II and III are more common in the region east and south of Houghton. However, small scale variations in composition in a single outcrop are fairly common. (Table I).

Group I. Feldspar gneisses with significant amphibole or pyroxene.

Most of the "Houghton magma" rocks described by Benson (1909) fall into this group. In particular Benson described and analysed a specimen from a quarry in the Houghton Cemetery Reserve and this can be taken as a "type" locality for the Houghton "Diorite". He describes this rock as consisting of plagioclase (An_{30}), diopside (slightly uralitized), minor microcline, with common accessories Ti magnetite, sphene, apatite and epidote. "The rock would be best termed a diopside diorite." The analysis was carried out on this particular rock because it appeared to be the least altered of the Houghton rocks. He comments further that the plagioclase

TABLE I

Variations in the feldspar gneisses in a small outcrop in the Little Para River at Inglewood. (727 977).

Spec. No.	Representative Thickness.	Description.
157-700	2'	Somewhat altered. Av. grain size .1-.4mm. 70% oligoclase, 20% actinolite with accessory epidote, apatite, opaques. Secondary alteration of plagioclase to sericite. Some biotite.
700A	Thin concordent vein	Coarse grained phase. 70% plagioclase An_{25}. Some patch antiperthite. 15% actinolite with abundant accessory epidote.
700B	1'	Coarser phase of 700. Average grain size 1.5mm. The plagioclase shows strain shadows.
700C	3'	40% plagioclase (albite), 40% actinolite, 20% epidote most commonly as clusters in the clear plagioclase. Accessories biotite, sphene, opaques.
700D	6'	Mosaic of clear poorly twinned plagioclase (An_0) with abundant epidote. Minor actinolite. Accessories sphene, opaques, biotite.

Spec. No.	Representative Thickness.	Description.
157-700E	2'	Microcline, film perthite 60%. Plagioclase 25% in distinct layers and as patch perthite in the K feldspar. Abundant accessory actinolite (giving rise to layering), biotite and epidote.
700F	1'	Similar to E but with more biotite (5%) in good preferred orientation. Some relic pyroxene.
700G	6"	80% hornblende and epidote, 10% biotite, 10% water clear albite.
700H	4'	Well banded. 45% plagioclase, 10% microcline, 40% diopside, Accessory epidote, apatite, opaques.
700J	3'	Slightly crosscutting relationship to H. Microcline perthite and plagioclase 60%. Dippside 30%. Accessory epidote and actinolite.

seems to be more basic than any other in the mass.

A specimen collected (157-888) at what is considered to be the same quarry as that mentioned by Benson,[1] showed a far greater degree of alteration than Benson's "diorite" and serves to illustrate the variation in degree of retrogression to be expected within the "diorite" body. The hand specimen exhibits a crude layering due to varying amounts of dark minerals, but there is no lineation. In thin section the following mode has been obtained: plagioclase 39%, microcline 7%, actinolite 18%, epidote 28%, accessories (apatite, sphene and opaques) 8%.

The plagioclase is albitic(An_7, $X' \wedge$ cleavage in sections perpendicular to <u>a</u>, 13-15° neg.), and occurs in two distinct habits. In the mafic-rich bands the plagioclase is clear and is studded with granules of epidote; in the feldspar-rich bands the plagioclase is cloudy, the alteration products being a very fine grained sericite and an unidentified "dust". Microcline occurs sporadically and most commonly in patch perthitic relationship with the albite. Actinolite occurs as ragged individuals, the margins of which contain much epidote and a more fibrous actinolite. Epidote, apart from occurring in the plagioclase, forms clusters of grains in the mafic layers, and occurs in cross-cutting veinlets.

1. This author could not find a quarry in section 5637, which Benson gives as the section from which the analysed specimen was collected. The quarry from which 157-888 was collected is in section 5516 and is less than 200 yards from the cemetery.

These two rocks illustrate rather well the variation in degree of regrogression of the rocks of the mass. Rocks which show no alteration of the pyroxene to actinolite are virtually absent, and it is believed that the feldspars in most cases have altered in composition at the same stage as the alteration of the pyroxene, as low anorthite content plagioclases are found in rocks rich in fibrous actinolite.

As far as can be seen the original assemblage before retrograde metamorphism was plagioclase (An_{25-30}) and diopside with variable amounts of microcline. Plagioclase -diopside rocks are more abundant than microcline-diopside rocks. Quartz is rare as a constituent of the main part of the mass but is found in the gneisses south of Houghton. Hornblende is rare as a primary metamorphic mineral, being found only in one specimen (157-1064); opaques, apatite and possibly epidote may have been original accessories.

More commonly the present metamorphic assemblage is plagioclase of variable composition (albite-oligoclase), with actinolite and epidote as the common dark minerals. The observation of abundant disequilibrium textures indicate that these minerals are derived from the oligoclase (andesine)-diopside assemblages.

Mineralogy.

The plagioclase grains are characterised in thin section by the presence of fine albite law twin lamellae. Only rarely (157-38 a and b) are more complex twin relations observed and then in somewhat larger grains than the matrix plagioclase; the twin law in this case proved to be Manebach-albite law. In rare cases a fine lamellar twinning sub-parallel to {001} was observed.

As a consequence of the fineness of the twinning the compositions of the plagioclases were determined by the Rittman zonal method[1] (Emmons, 1943) using the data from Troger (1956). As stated before, some assemblages had plagioclase of composition An $_{25-30}$ but more commonly the composition varied from An_{7-12} (in the actinolite rich rocks), although plagioclases more albitic have been observed. In the rocks where large plagioclase crystals were observed (38a and 38b) the larger crystals were more Ca rich than the matrix feldspars. (Table II).

In many thin sections a patchy alteration of the plagioclase to albite-sericite is evident. In such cases the altered (sericite-rich) portions are always less calcic than the unaltered portions.

Microcline is present in small amounts in many of the feldspar gneisses and in some gneisses varies in amount up to about 80% of the total feldspar.

1. It was a matter of common observation that the extinction angles in sections perpendicular to a were remarkably consistent (4^{o} variation rare) in any one thin section unless the plagioclase was partially altered to sericite. This suggests the general attainment of feldspar composition equilibrium in most of the rocks even though arrested stages in the alteration of diopside to actinolite are widespread.

TABLE II.

Feldspar Compositions in 157-38a and 38b.

	157-38a		157-38b	
Method	Large grains.	Small grains.	Large grains.	Small grains.
Turner (1947)	An_{23}	An_{12}	An_{23}	An_{12}, An_{13}
van der Kaaden (1951).	(Manebach)	(Albite law)		
Rittman zonal	An_{22}(001)	An_{9}, An_{10}	An_{23}(001)	An_{11}, An_{15}
	An_{23}(010)	An_{11}, An_{12}	An_{26}(010)	An_{10}

Range of compositions: Large grains An_{22-26}

Small grains An_{9-15}

A high microcline content does not show any particular correlation with the amount of dark mineral present; amounts of up to 50% dark minerals in microcline-rich rocks have been observed, although commonly the mafic content is less than 25% of the rock. The microcline is almost invariably twinned and commonly perthitic, and is fresh in most thin sections.

Pyroxene is the most important primary mafic mineral and has optical properties suggesting diopside, $Z^\wedge c$ 40-44°, $2V_Z$ 58-60°, β ca. 1.69 (difficult to determine due to a dusty alteration). No detectable difference in 2V and $Z^\wedge c$ was found between the less altered grains and the grains in which pyroxene is relic in amphibole.

Hornblende was observed in only one slide (157-1064). It has a high negative 2V, $Z^\wedge c$ 22° and a pleochroic scheme, X light brown, Y brownish-green and Z blue-green.

Actinolite is extremely common in the feldspar gneisses, either as fibrous masses after pyroxene or as individual anhedral grains. In some cases actinolite needles are common throughout the slide or as outgrowths on pyroxene or actinolite. Where the optics can be determined $Z^\wedge c$ is 13-17° and β ca. 1.63. The pleochroic scheme is X pale yellow-green, Y yellow-green and Z light blue-green.

Epidote may occur as an essential constituent or as an accessory. In many cases it appears to be the result of alteration of plagioclase to give an albite-epidote assemblage, and in some rocks the presence of epidote veins suggests high mobility of the constituents if not introduction of material.

Common accessory minerals are epidote, opaques, apatite, zircon and sphene, with biotite occurring in only a few specimens. Sericite is also common but is thought to be due to a later alteration.

Group II. Feldspar gneisses with essentially no mafic minerals.

Only seven rocks of this type were collected and possibly this represents an oversampling as they are distinctive rocks in the field, appearing rather like white quartzites in outcrop.

Mineralogically they are very simple, consisting of albite (An_{0-7}) and microcline in inversely varying proportions, and accessory opaques and rare brown biotite (largely altered to a green-brown variety). The opaques are disposed in layers for the most part and in hand specimen the rock looks like a quartzite with sparse heavy mineral layers.

Apatite, zircon and green tourmaline are rare accessories (The last appears to be a late stage mineral).

Group III. Quartz feldspar gneisses.

With increasing amounts of quartz the leucocratic feldspar gneisses grade into quartz-feldspar gneisses, which contain quartz (up to 20%), microcline and acid plagioclase as the dominant constituents. The ratio of microcline to plagioclase is extremely variable, both quartz-microcline and quartz-plagioclase rocks being observed. Some phases appear granitic in texture and contain essential biotite (157-835). In thin section 157-835 is seen to consist of irregular shaped strained quartz,fresh anhedral microcline (.4 - .6 mm.), sericitized plagioclase and a straw-yellow to brown pleochroic biotite (up to 1 mm.). Plagioclase is more abundant than microcline and if igneous and unmetasomatized the rock would correspond to a granodiorite. Accessories are epidote, opaques, apatite and rutile. The biotite shows signs of alteration and shows the characteristic crystallographically-oriented rutile needles. The field relations of this mass (south of Houghton) are not known as it occurs as isolated low outcrops in a grassy paddock.

Another textural variant is the occurrence of quartz-feldspar gneisses with granulitic fabric in the region just south of Houghton (157-994) and will be discussed under textural features. These are the leucocratic granulites of Spry (1951) and the gneissic aplites of Benson (1909).

It is seen from the description above that the rocks of Groups II and III are closely similar but that the rocks of Group I may form a distinct compositional group. This is borne out in the geographical distribution as the mafic-rich rocks are confined to distinct regions whereas the other groups may be interlayered with schists and less feldspathic gneisses (see description of River Torrens gneisses).

Cutting across the feldspar gneisses are rare pegmatites which are commonly small veins of quartz, microcline and variable amounts of plagioclase. Graphic texture occurs in some pegmatites. In thin section microcline and plagioclase are seen to be intergrown in structural continuity, with quartz occurring as rounded grains, separated in two dimensions, but which are in optical continuity (Fig. 9).

(c) Textural features of the feldspar gneisses.

Dark minerals.

Diopside commonly occurs as subhedral grains, euhedral rarely. It is generally somewhat altered and at an early stage of alteration minute needles of actinolite project from the diopside into the surrounding minerals(Fig. 10). More advanced stages show a distinct border of actinolite enclosing the diopside, with good crystallographic continuity of pyroxene and amphibole. In other cases the amphiboles appear to cut across the pyroxene grains with no crystallographic control. Other types of alteration show as well-oriented patches of actinolite in pyroxene. These may be discrete, giving a chequerboard appearance to the crystal in extinction (Fig. 11) or the parts may be connected.

Actinolite commonly occurs as sheaves of needles, but may also occur as large discrete grains. In some instances veins contain euhedral crystals of actinolite. Some of the large actinolite grains enclose feldspar grains in a

Fig. 9. Regular intergrowth of plagioclase (light-grey), microcline (cross-hatched) and quartz (medium-grey, strained). The plagioclase and microcline are in structural continuity.
Pegmatite. Spec. 157-1025.
Crossed polars.

Fig. 10. Needles of actinolite as outgrowths on diopside.
Feldspar gneiss. Spec. 157-704 Bl.
Crossed polars.

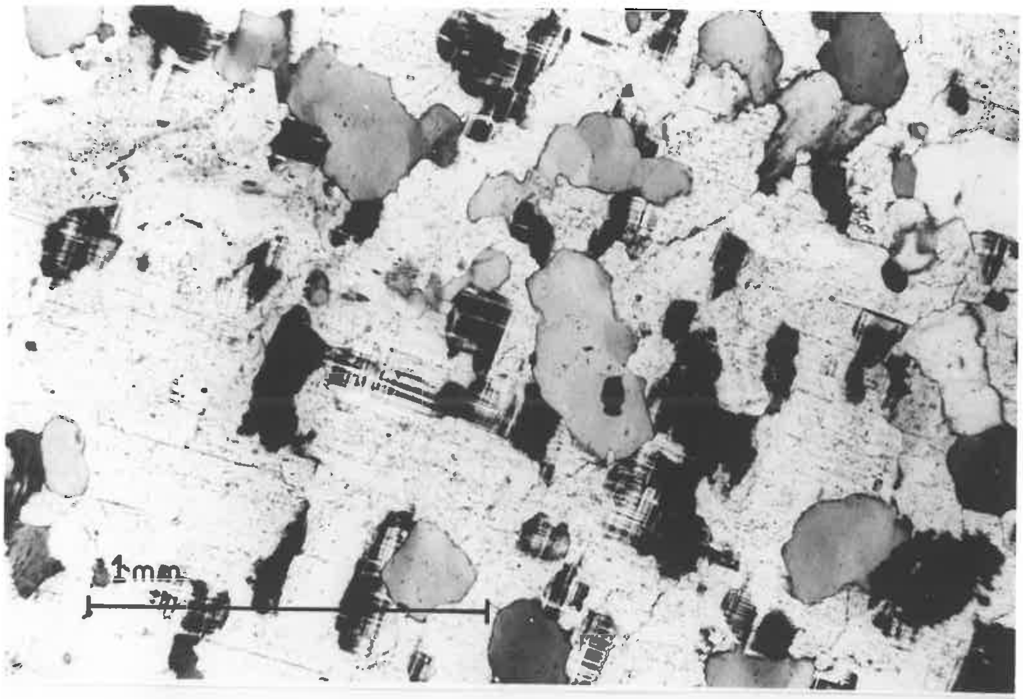


FIG. 9

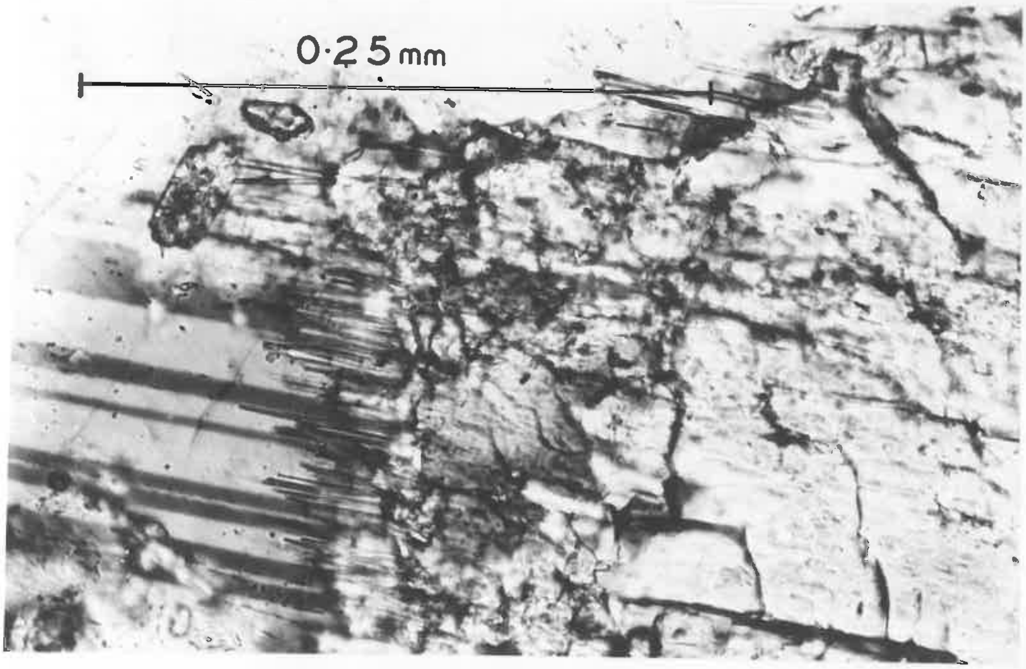


FIG. 10

poikiloblastic relationship. Fibrous actinolite is seen filling fractures in some of the specimens (Fig. 12).

Epidote occurs as granules in feldspar or as clusters of grains or in veins. In 157-704B a small vein cuts across several crystals. Where the vein is cutting across feldspar the epidote is granular but where cutting across amphibole it appears as a single crystal, having a somewhat fibrous appearance.

Biotite occurs as a minor phase in some of the rocks and commonly shows a good preferred orientation. In many cases the biotite has been altered and the new phase is an olive-green biotite with needles of ? rutile arranged in a 60° network in the basal plane (Fig. 13). More commonly, the biotite is crowded with very small grains of an opaque mineral.

The opaque minerals, titano-haematite and magnetite, commonly occur as equidimensional anhedral grains or as elongate interstitial groups of grains. Very rarely euhedral grains are observed.

Feldspars.

An anhedral granular texture with straight or slightly curved feldspar-feldspar boundaries is relatively rare in the mass. Such a texture is most common in the rocks very rich in plagioclase and may represent the rare attainment of metamorphic (or metasomatic) equilibrium. In these rocks (e.g. 157-1047, 157-916) the plagioclase forms anhedral to subhedral grains which are clear and have well-developed albite

Fig. 11. Optically continuous patches of actinolite (grey) in large diopside grain.
Feldspar gneiss. Spec. 157-1047.
Crossed polars.

Fig. 12. Layering in feldspar gneiss with actinolite prisms showing a preferred orientation parallel to the layering. Actinolite is concentrated in the fault in the centre of the picture.
Houghton. (Grid. ref. 766 966).

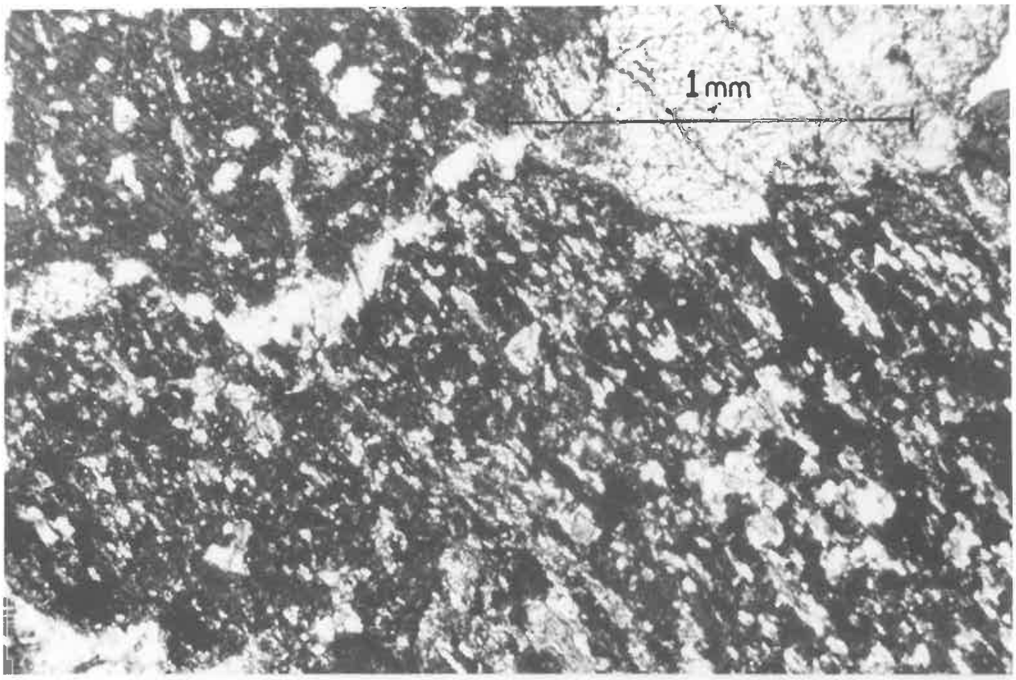


FIG. 11

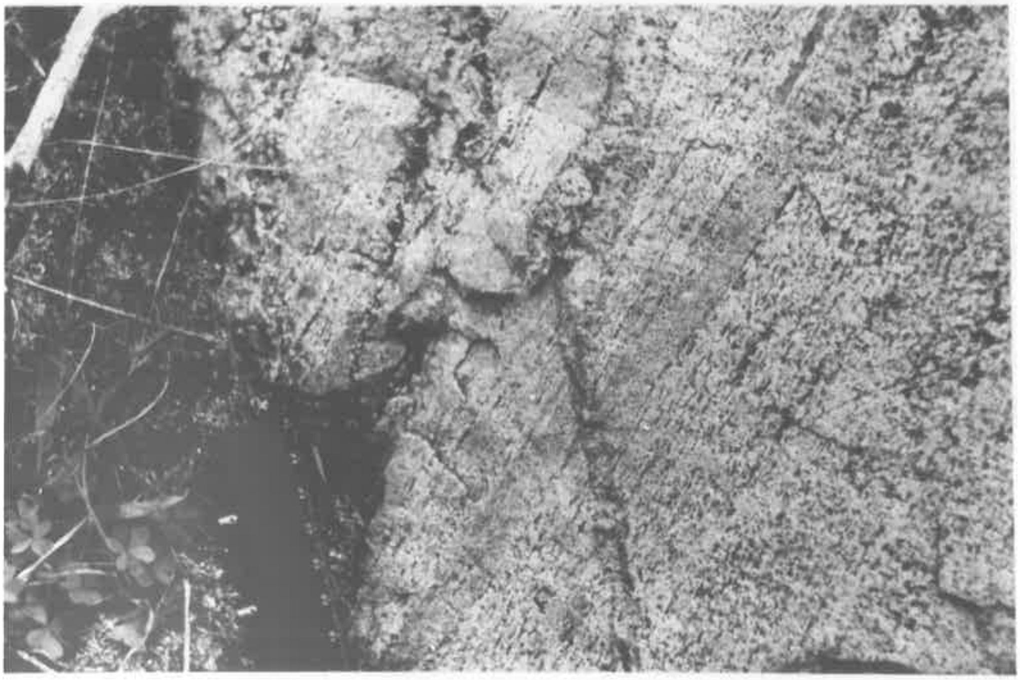


FIG. 12

twins which cross the grains without change or interruption (Fig. 14). The microcline tends to be interstitial but may form grains as large as the plagioclase. Neither of the feldspars are perthitic.

More commonly the rocks do not have an even granular texture. The boundaries between feldspars are complexly sutured and perthitic relations are common. In some rocks the microcline is simply a film perthite with the fibres of albite fairly uniformly distributed. Good braid perthite (Goldich and Kinser, 1939) has not been observed, but some film perthites show a tendency to cross cut the normal film direction ca. (100) and may be close to the (110) and $(1\bar{1}0)$ planes (Fig. 15) which Goldich and Kinser report as characteristic of braid perthite. In view of Goldich and Kinser's conclusion that much braid perthite is the result of replacement it should be noted that this texture occurs (in the Houghton area) only in perthites with a high percentage of albite.

Another type of perthite commonly observed is a mixture of patch and film perthite. In some cases veins may cut across the films or may be parallel to them. In other cases the albite occurs in patches instead of veins, the albite in the patches being in optical continuity with the albite in the films.

Fig. 13. Crystallographically oriented rutile needles in biotite.
Feldspar gneiss. (retrograde) Spec. 157 - 700.
Plane polarized light.

Fig. 14. Subhedral to anhedral granular texture in plagioclase-actinolite gneiss.
Feldspar gneiss. Spec. 157-916.
Crossed polars.

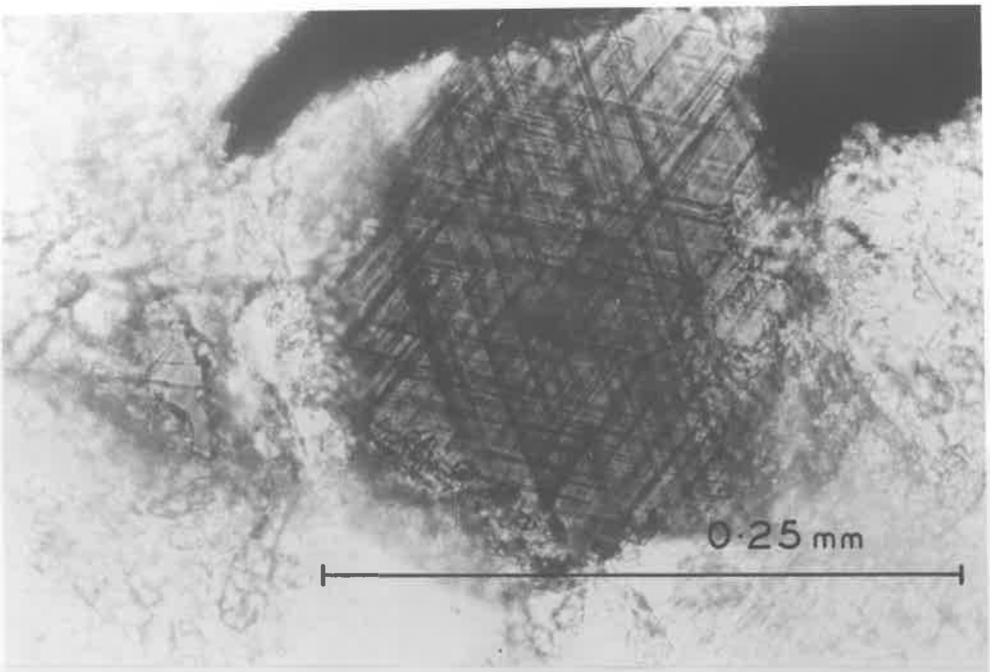


FIG. 13

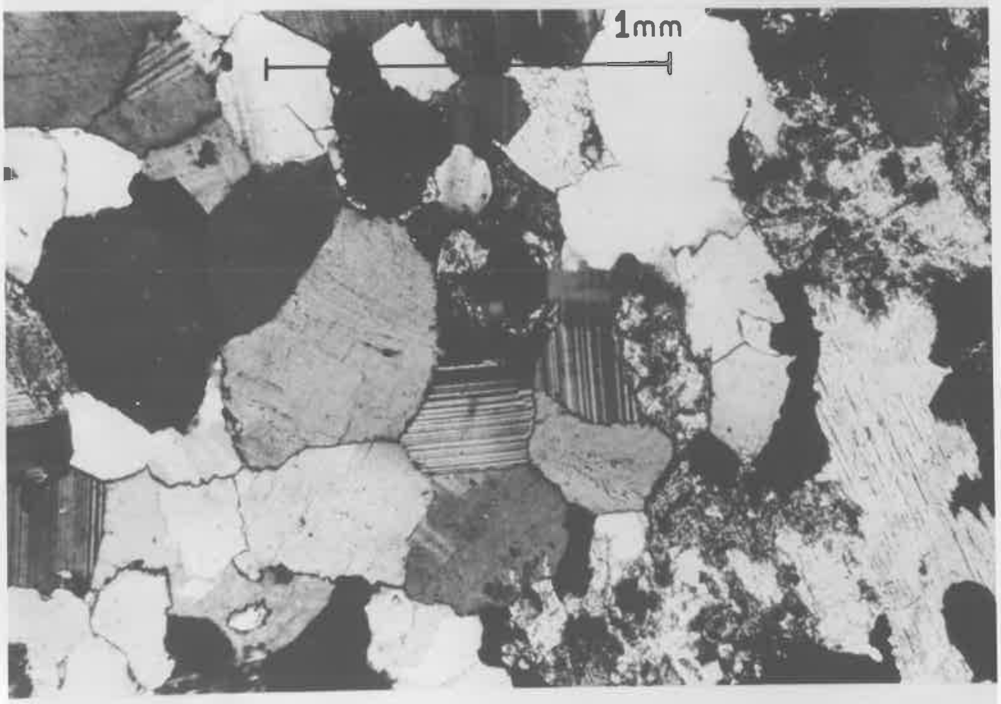


FIG. 14

The patches tend to be in areas poor in albite films, a relationship interpreted by Gates (1953) as that of the patches developing from the film perthite by more complete separation of the microcline and albite. Tuttle and Bowen (1958) invoke such a mechanism to produce completely segregated potash feldspar and albite grains in some granites.

However, the evidence at Houghton suggests that processes in addition to exsolution also have been important. Albite patches occur also in regions rich in film albite (Fig. 16), a distribution not in conformity with Gates (1953) exsolution hypothesis. Furthermore, cases have been observed where film perthite is distinctly concentrated near patches of albite (Fig. 17) and here it could be interpreted that the film perthite is an early replacement stage in the sequence rather than that the patch is a later stage in an exsolution and recrystallization.

Rims of albite around microcline grains are also very common in the gneisses. In rare instances there are rims around non-perthitic microclines (Fig. 18) but more often the rims are found in rocks where the microclines are perthitic. In a few cases the albite is not in crystallographic continuity with the albite perthite lamellae but more commonly it is. At some contacts the albite rims appear to be replacing the microcline perthite, and the albite lamellae of the perthite are unaffected (Fig. 19).

These types of perthitic-rim relationships have been

Fig. 15. Microcline film perthite grading into (?) braid perthite. Some albite is concentrated at the margins of the microcline grain and is in optical continuity with the albite lamellae in the perthite.
Feldspar gneiss. Spec. 157-846
Crossed polars.

Fig. 16. Patches and films of albite (grey) in microcline (cross hatched). The albite patches occur in areas both rich and poor in albite films.
Feldspar gneiss. Spec. 157-699B_1
Crossed polars.

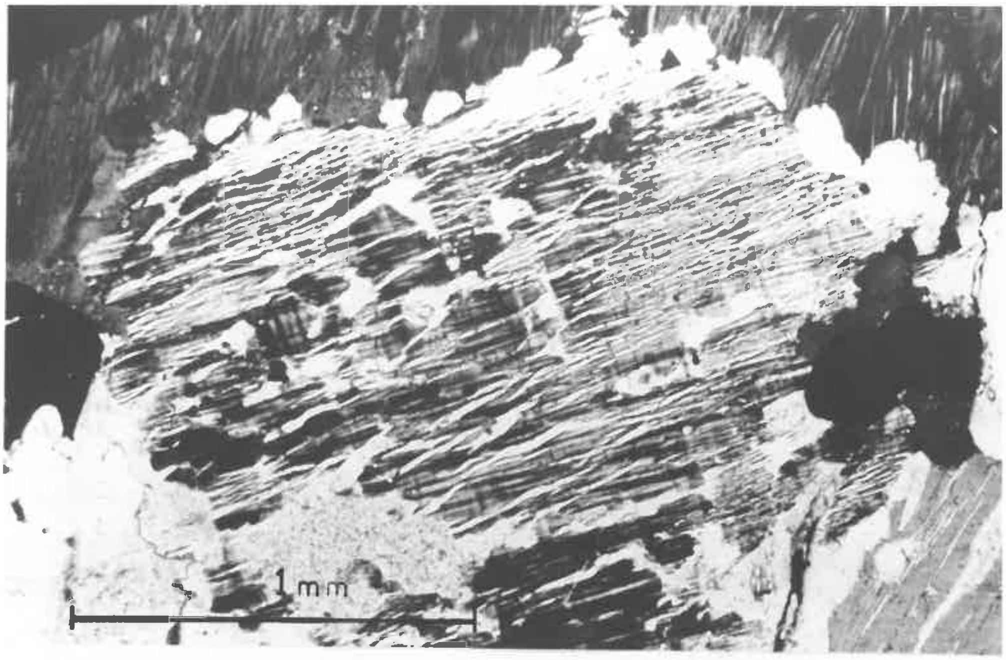


FIG. 15

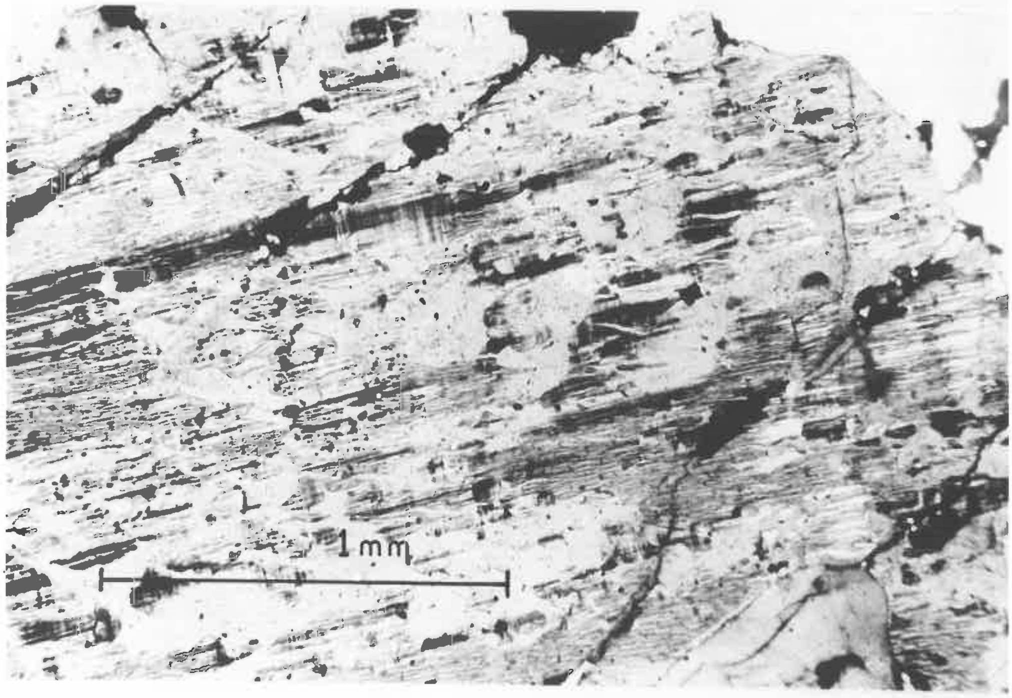


FIG. 16

Fig. 17. Marginal patch of albite in continuity with film albite in microcline (cross-hatched).

Feldspar gneiss. Spec. 157-1076

Crossed polars.

Fig. 18. Clear albite rim (grey) around microcline grain (cross hatched).

Feldspar gneiss. Spec. 157-701.

Crossed polars.

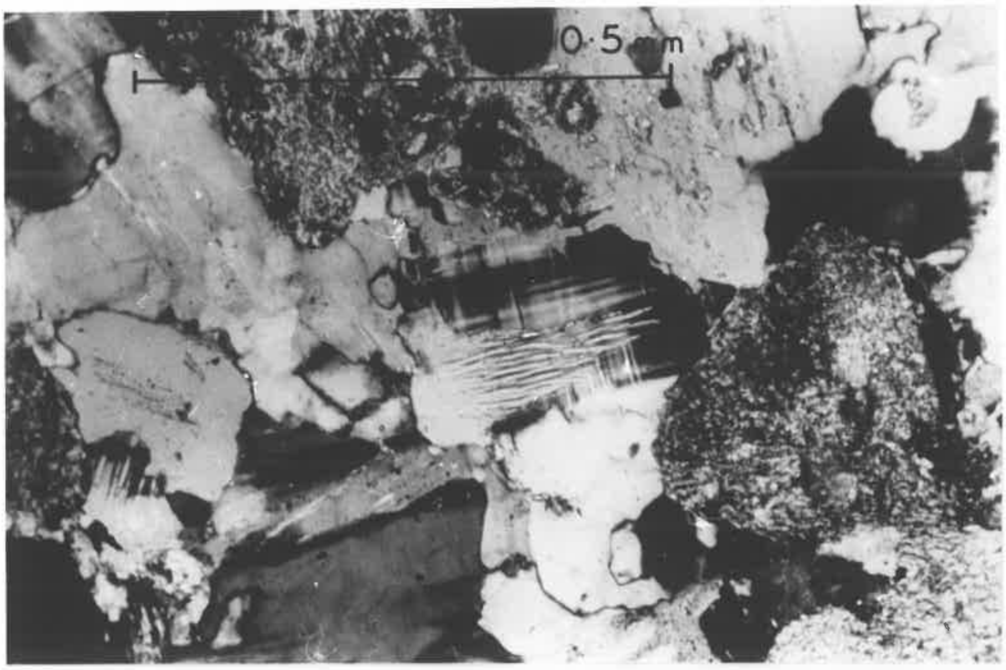


FIG. 17

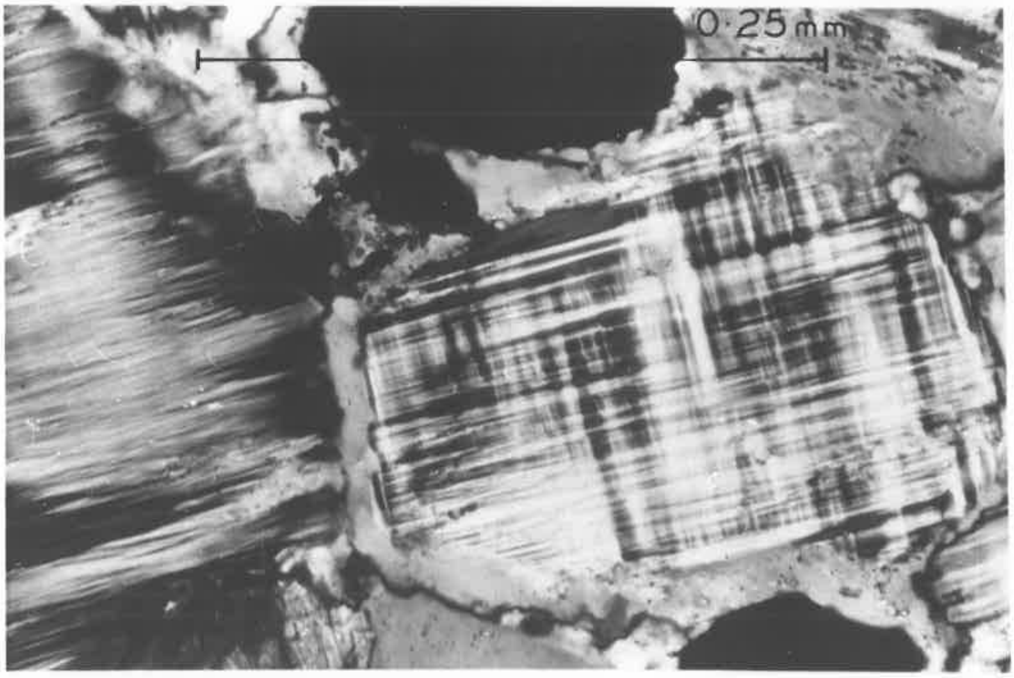


FIG. 18

Fig. 19. Microcline perthite (microcline medium grey, with white albite lamellae and patches) with a rim of albite (dark grey). The albite rim is replacing the microcline, leaving the white albite lamellae as remnants in the dark grey albite.

Feldspar gneiss. Spec. 157-846

Crossed polars.

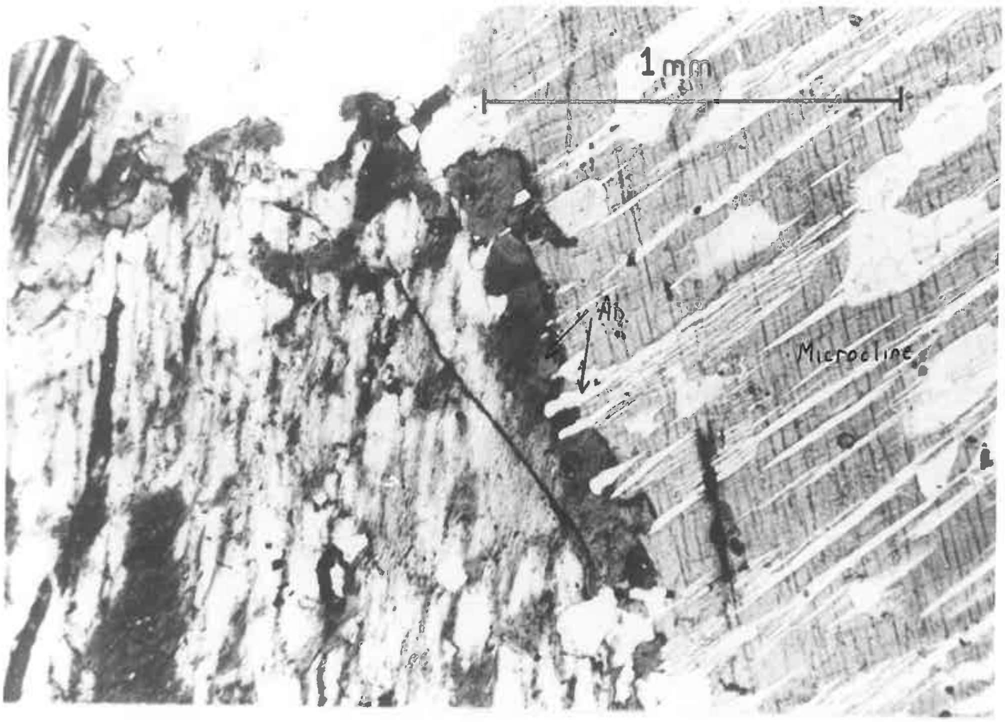


FIG. 19

interpreted by Ramberg (1962) as the result of exsolution from the microcline, followed by migration of the albite. While this seems difficult to disprove conclusively, similar relationships have been considered to be of replacement origin by other authors. (Gilluly, 1933).

With increasing concentration of albite in the perthites, they grade into antiperthites. Although film perthitic, relationships are observed in highly albitic perthites, the most common relation is patch perthitic. In these examples the plagioclase is commonly twinned and in crystallographic continuity with the microcline.

A few cases are observed where the albite in crystallographic continuity with microcline in one grain extends into another microcline grain of different orientation. (Fig. 20). Other examples are common where microcline appears as relics in plagioclase (Fig. 21).

Albite occurs more rarely at quartz-feldspar or plagioclase - plagioclase boundaries. These relationships are seen more clearly in the Torrens Gorge and a discussion of them is postponed to that section, as is also a discussion of the origin of the features mentioned above.

Quartz.

Uncommonly quartz occurs as interstitial grains between feldspars or dark minerals. More commonly it occurs as small circular grains in thin section. The rounded grains occur both at grain boundaries, in which case the feldspars are moulded onto the quartz, or as inclusions in the feldspars

Fig. 20. Patch intergrowth of microcline and plagioclase. The plagioclase is in structural continuity with one microcline grain but cuts across the principal crystallographic directions in a second grain. The plagioclase is crowded with sericite.

Feldspar gneiss. Spec. 157-988

Crossed polars.

Fig. 21 Microcline remnants (cross hatched) in irregular intergrowth with plagioclase (light-grey). The microcline remnants are in optical continuity.

Feldspar gneiss. Spec. 157 - 694B.

Cross polars.

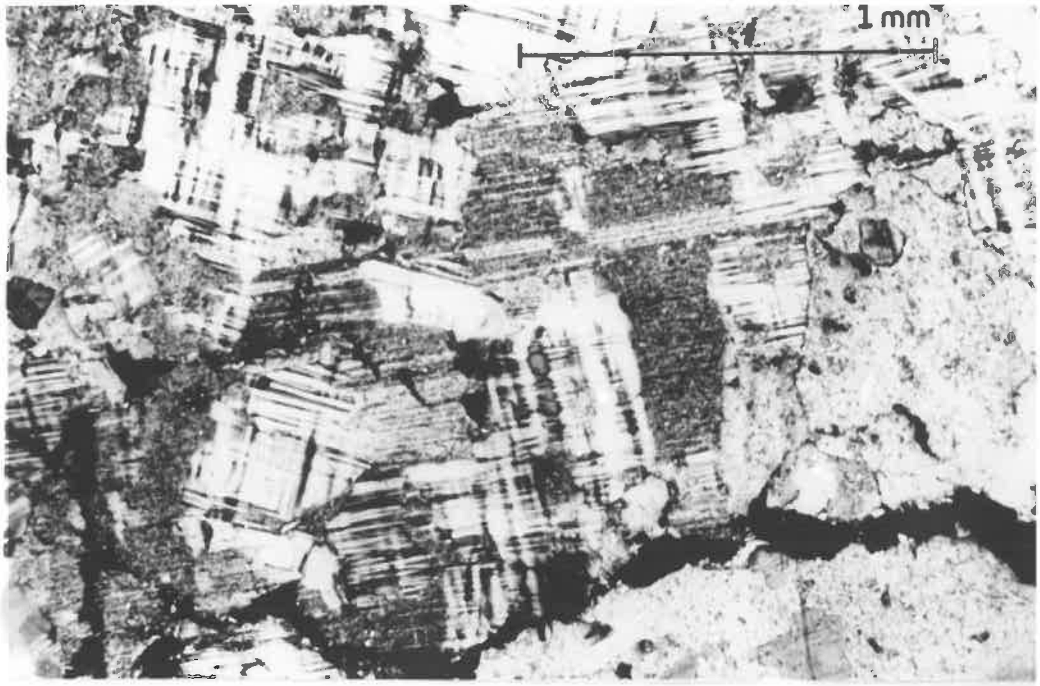


FIG. 20

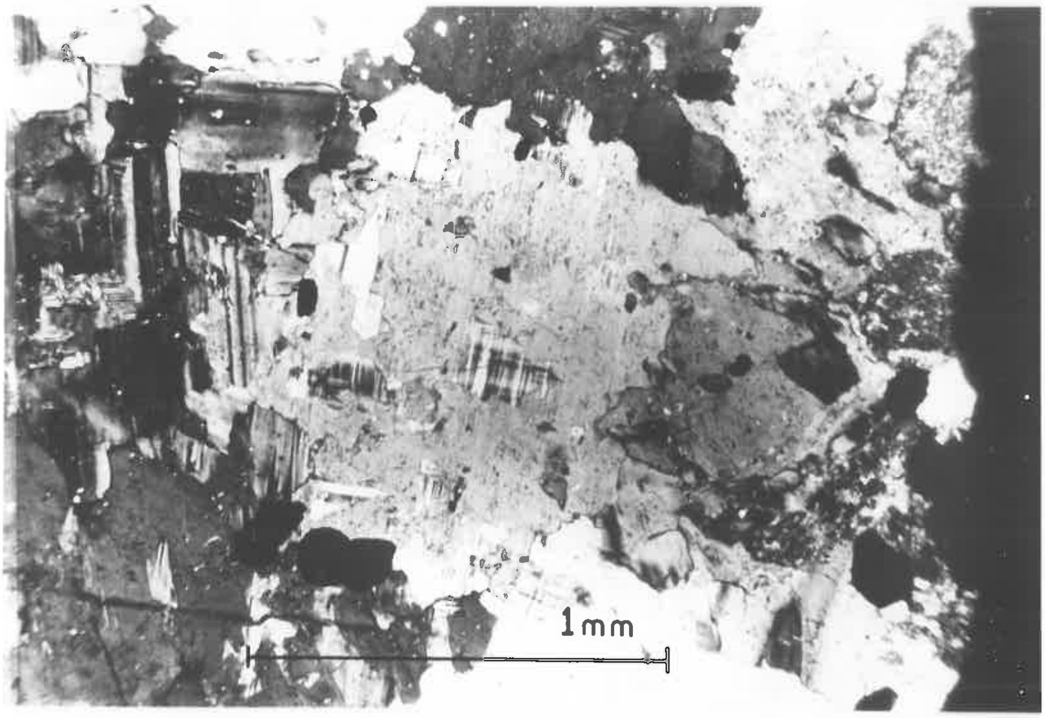


FIG. 21

(most commonly microcline).

In some cases quartz occurs as flattened spindles giving a characteristic granulitic texture. This type of occurrence is restricted to the area south of Houghton and is widespread in the Torrens Gorge area.

In certain cross-cutting pegmatites quartz occurs in typical graphic intergrowth with feldspars.

Quartz "intrudes" feldspars in some cases, along fractures and cleavages; this appears to be related to a later phase in the metamorphic history.

(d) Post-crystalline deformation and sericitization.

Quartz where present invariably shows strain shadows and in many cases is recrystallized. Plagioclase and microcline show undulose extinction in many specimens and in some cases the twin lamellae of plagioclase show distinct flexures (Fig. 22). In other, rarer cases kinks are observed in the plagioclase lamellae (Fig. 23). Recrystallization of the feldspars is rare and most re-crystallized aggregates are composed of quartz and sericite. However, some cases are observed where zones of recrystallization cutting plagioclase consist largely of plagioclase and sericite. The general absence of feldspar in the recrystallized portions may be the result of the ease of alteration of the plagioclase to sericite.

Most of the gneisses show retrogression features to a greater or lesser extent. The abundance of epidote has

Fig. 22. Bent twin lamellae in plagioclase.
Feldspar gneiss. Spec. 157-106.
Crossed polars.

Fig. 23. Kinked twin lamellae in plagioclase.
Retrograde feldspar gneiss. Spec.
157-109.
Crossed polars.

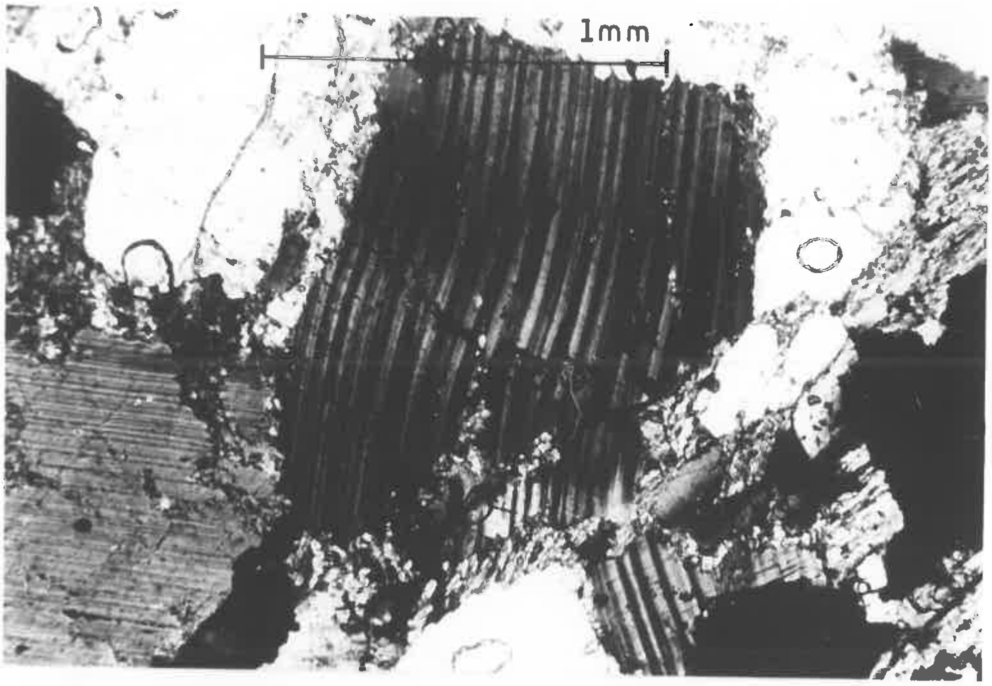


FIG. 22

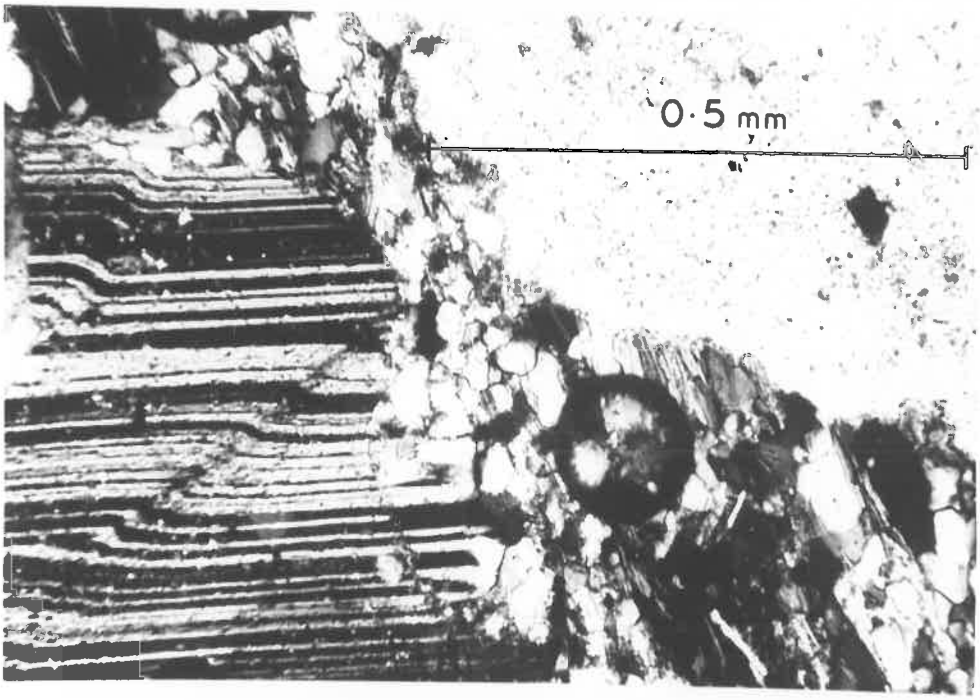


FIG. 23

already been commented on and represents a very common alteration of the plagioclase gneisses. Sericite is another common alteration product of plagioclase and is important in considering the development of the widespread retrograde rocks.

This sericitic alteration occurs in many different ways. The sericite may occur along grain boundaries or in cross-cutting fractures, the main body of the plagioclase grains remaining relatively unaltered. Such an alteration is most common in rocks which show signs of post-crystalline deformation and is probably related to the wide-spread phyllonitization which is discussed later.

More commonly the plagioclases have sericite flakes randomly distributed throughout the grains. The alteration may be patchy, the altered portion of a grain being more albitic than the unaltered parts. A more advanced stage of this type of alteration is shown in grains where the nature of the original mineral is only just discernible as vague multiple twinning.

The orientation of the sericites within the plagioclase is most commonly random but in many cases the flakes lie in the (001) cleavage plane, less commonly in the (010) cleavage and twin boundary planes. This type of occurrence is common in the less altered plagioclases and may indicate an early stage of alteration.

The end result of this alteration in plagioclase rocks is a rock in which the grains can still be seen in outline but appear as a confused aggregate of sericite with rare light green biotite and chlorite. Where the alteration is complete the nature of the original grain may be unknown.

Microcline is much less susceptible to alteration than plagioclase. Even in rocks where the plagioclase can only just be recognized, microcline is quite fresh. (Fig. 20). This feature is observed also in the various perthites, even to the finest of film perthites where the albite lamellae may be altered although the host is still fresh. This greater ease of alteration of plagioclase may lead to false conclusions concerning the relative proportions of original plagioclase and microcline in the highly retrograde rocks surrounding the feldspar gneisses.

2. THE FELDSPAR GNEISSES OF THE KERSBROOK REGION.

Rocks lithologically similar to the Houghton feldspar gneisses are found in the area south and east of Kersbrook. On the south and east sides of the mass they are in contact with younger Precambrian rocks, while their contacts with the surrounding schists of the Houghton Complex are largely obscured on the west and north by Tertiary and Recent alluvial deposits of the Kersbrook Valley.

Two distinct lithologic types are dominant in the Kersbrook mass: 1. feldspar - actinolite gneisses (similar to the mafic gneisses at Houghton) which are dominant in the northern part of the mass, 2. well-layered fine-grained leucocratic gneisses which are common in the southern part of the mass.

Of minor quantitative importance is the occurrence of a few veins of pegmatite which consist of cross-cutting masses of either diopside or actinolite commonly associated with Ti-haematite.

Petrographically the mafic gneisses consist of plagioclase, microcline-perthite, quartz and actinolite (Diopside was not observed in any of the gneiss specimens, although it occurs in pegmatite as mentioned above).

Plagioclase is the most abundant mineral (up to 90% in some rocks); microcline is rarely the dominant feldspar. The amphibole occurs as sheaves of fibrous actinolite and may be parallel to the crude layering or may be randomly oriented. There is commonly not more than 10% amphibole present, although up to 25% has been observed locally. The most common accessory is dispersed anhedral haematite. Epidote, apatite, sphene, zircon and monazite are also present (given in order of abundance -epidote most abundant).

The leucocratic gneisses differ mainly in their lack of amphibole, greater amount of microcline (with respect to plagioclase) and quartz, and ubiquitous occurrence of opaques concentrated in certain layers. Green biotite is abundant in some specimens.

All the specimens collected from the Kersbrook area show very similar textural features. In the rocks just east of the Kersbrook Cemetery (157-1128 and 1129) the quartz and feldspar occur as an anhedral granular mosaic (microcline being dominant). The quartz shows distinct strain bands and the microcline shows undulose extinction. Biotite occurs in these specimens but is not strained. It cuts across the other mineral boundaries, and has a poor preferred orientation.

In the plagioclase-rich rocks the plagioclase does not show the normal type of albite twinning. Single albite grains show an irregular development of short twin lamellae which give a chequer-board appearance to the grain (Fig. 24. Thin sections 157-1130 and 1149 also show excellent examples). Elsewhere similar plagioclases have been called chequer-albite (e.g. Gilluly, 1933). The grain boundaries of the plagioclases are commonly irregular and may be sutured in a manner similar to quartz grains.

It is believed that the undulose extinction in quartz and microcline and the development of chequer-albite are indicative of deformation. The origin of chequer-albite has been discussed in the literature and is commonly thought to be the result of metasomatic replacement of potash-feldspar by albite (Gilluly, 1933) although Battey (1955) considers that the chequer board appearance results from "porphyroblastic growth proceeding from numerous centres and approaching axial parallelism by successive recrystallizations". Starkey (1956) demonstrated that the development of chequer-albite in some metamorphosed porphyries of New Brunswick is related to

Fig. 24. Discontinuous albite-law twins in plagioclase (Chequer albite).
Feldspar gneiss. Spec. 157-1149.
Crossed polars.

Fig. 25. Folded albite lamellae in plagioclase.
Quartz-plagioclase gneiss Spec. 157-1127.
Crossed polars.

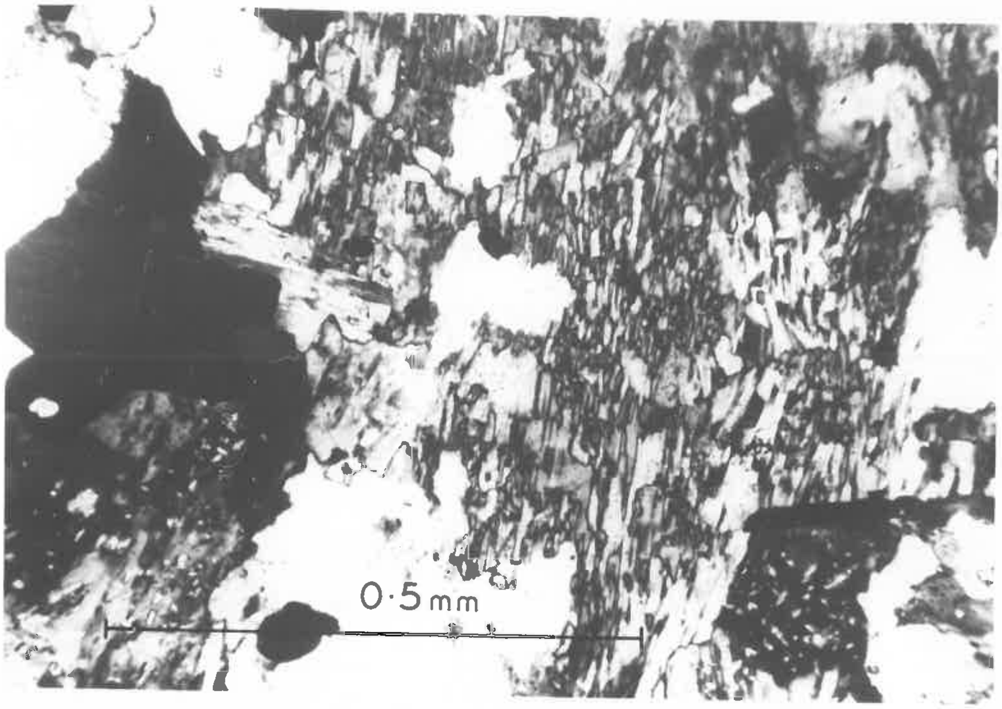


FIG. 24

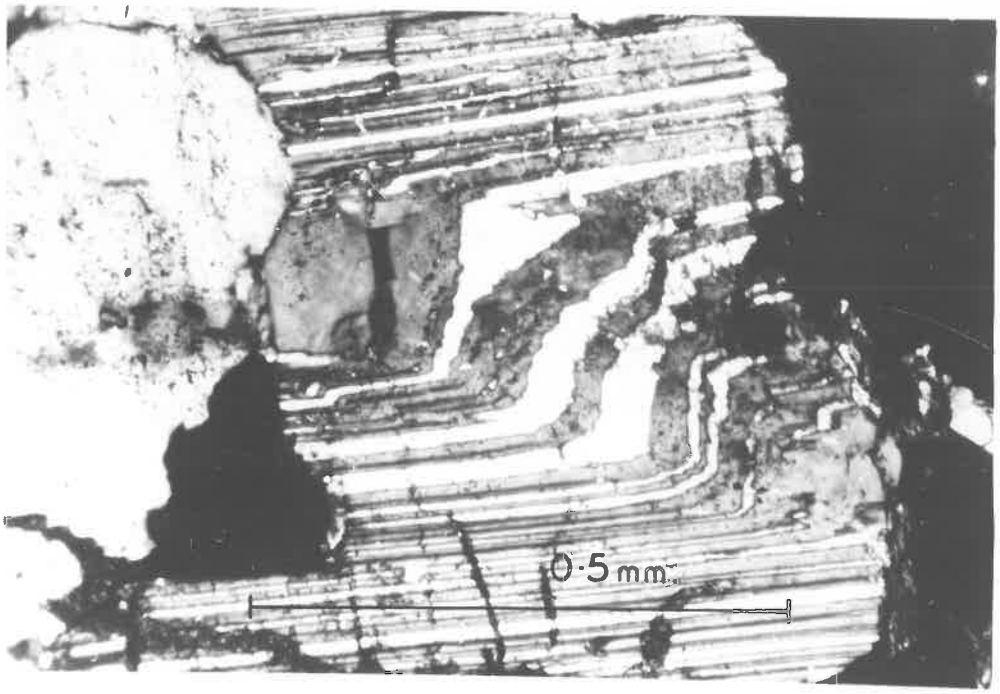


FIG. 25

increasing metamorphism (particularly increasing stress). This conclusion is in accord with the author's observations that the Kersbrook gneisses show more deformation textures than the Houghton feldspar gneisses, and that chequer-albite is exclusively developed in the Kersbrook mass.

Gneisses with textures similar to the Houghton mass are rare. An exception is an isolated outcrop of quartz-plagioclase gneiss (Grid ref. 833-988) which has an anhedral granular texture with kinked plagioclase lamellae (Fig. 25).

3. FELDSPAR GNEISSES OF THE PARACOMBE AND TORRENS GORGE REGION.

(a) Feldspar gneisses similar to the Houghton types have a very restricted distribution in the southernmost part of the Houghton Complex. A very limited series of outcrops occurring in the vicinity of Paracombe may be related to the Houghton mass, but is separated from it by a Tertiary erosion surface between Houghton and Paracombe.

Three samples of Houghton-type gneiss were collected from isolated outcrops in the erosion surface around Paracombe (157-880, 859, 941). These appear to be indistinguishable from the Houghton type in hand specimen and in thin section. They all contain predominant plagioclase (50-70%) of oligoclase composition with some rare microcline (ca. 10%). Diopsidic pyroxene was found in each, showing a variable amount of alteration to uralitic amphibole. Some larger single grains of actinolite occurred in one sample (157-880). Epidote, sphene and opaques are ubiquitous accessories.

(b) The other feldspar gneisses in the region differ markedly from the Houghton types just described and are subordinate to associated mica gneisses. Only one of the twenty-five examined microscopically showed any amphibole. All of them tend to be less massive than the Houghton type and occur interlayered with the more micaceous rocks. These feldspathic rocks appear to be of two distinct types; those with dominant plagioclase and those with dominant microcline, the latter having significant quartz in all specimens examined. No rocks with roughly equal proportions of feldspars were observed although plagioclase-rich and microcline-rich gneisses can be found in close proximity.

1. Dominant plagioclase group.

Some of the specimens examined (e.g. 157-398, 387) consist almost entirely of plagioclase (albite) in anhedral grains 1-2 mm. in diameter, the only other minerals present being accessory strained quartz, granular opaques and rare zircon and apatite. Sericite occurs as an alteration product in most of the samples. In other samples (e.g. 157-886) as much as 20% quartz may be present, and chlorite, muscovite and a light green mica are common (secondary) accessories. The quartz may occur interstially or as flattened rods or lenses cutting across the other minerals.

Microcline commonly occurs, as an accessory, as clear interstitial grains. Higher proportions (up to 30%) do occur but then the microcline occurs as an irregular intergrowth with plagioclase. The plagioclase generally appears to poikiloblastically enclose microcline. In rare cases where a grain is made up of equal proportions of the two feldspars the appearance is that of a regular intimate intergrowth.

2. Quartz-microcline gneisses.

Typically, rocks of the quartz-microcline gneiss group consist of 10-20% quartz, 70-80% feldspar and variable small amounts of mica.

The quartz occurs most commonly in cross-cutting lenses in thin section but also occurs as individual anhedral grains. Most specimens show strain shadows.

The dominant feldspar is a microcline film perthite with minor amounts of plagioclase (varying from 5-30% of the total feldspar). Both feldspars occur as anhedral grains to give an xenoblastic texture but in some cases microcline may be enclosed in or intergrown with the plagioclase.

In some cases a brown or green-brown biotite is a fairly common accessory and becomes an important constituent in the darker phases. Sericite is common as an alteration product and opaques, sphene, zircon and apatite are the most common accessories.

Textural features.

The texture of feldspar gneisses in the vicinity of Paracombe, is anhedral granular, the feldspar grain boundaries being smooth and gently curved. Pyroxene appears to have been subhedral but now exhibits overgrowths or replacements by actinolite. Actinolite also occurs as isolated ragged groups presumably representing completely altered pyroxene.

In the actinolite-deficient gneisses of the Torrens Gorge, quartz is a common constituent and exhibits a variety of textures. It may form part of an anhedral granular texture but more commonly occurs as rounded individuals onto which other minerals are moulded, or as grains enclosed in microcline. It also occurs in vein like forms (.5-1 mm. wide) which are rods or lenses flattened in the plane of the layering. Commonly each vein (in thin section) is a single grain of quartz, although polycrystalline aggregates do occur. The quartz shows abundant well developed strain shadows and in some cases small polygonal quartz grains are present within the large grain, either in clusters or chain-like forms. Occurring at right angles to these coarse veins are a subsidiary set of thinner quartz veins (generally .1mm.) (Fig. 26). They may cut, be cut or appear as offshoots of the main quartz stringers and are considered to be contemporaneous with the larger veins. They differ from the larger veins in always being polycrystalline; the veins are only one grain wide and the grain boundaries are at right angles to the vein wall. The large lenses as well as the small may appear cutting across single grains of feldspar and the boundaries with other minerals are fairly smooth and not sutured as are commonly quartz-quartz boundaries where the

Fig. 26. Veins of strained quartz crossing a microcline-plagioclase aggregate. A second system of (thinner) quartz veins appears contemporaneous with the first. (The two larger veins are connected outside the field of view of the picture).

Quartz-feldspar gneiss. Spec. 157-341. Crossed polars.

Fig. 27. Intergranular fine-grained quartz between plagioclase grains. Note also the sericite grains parallel to (001) in the top right-hand plagioclase.

Quartz-feldspar gneiss. Spec. 157-365. Crossed polars.

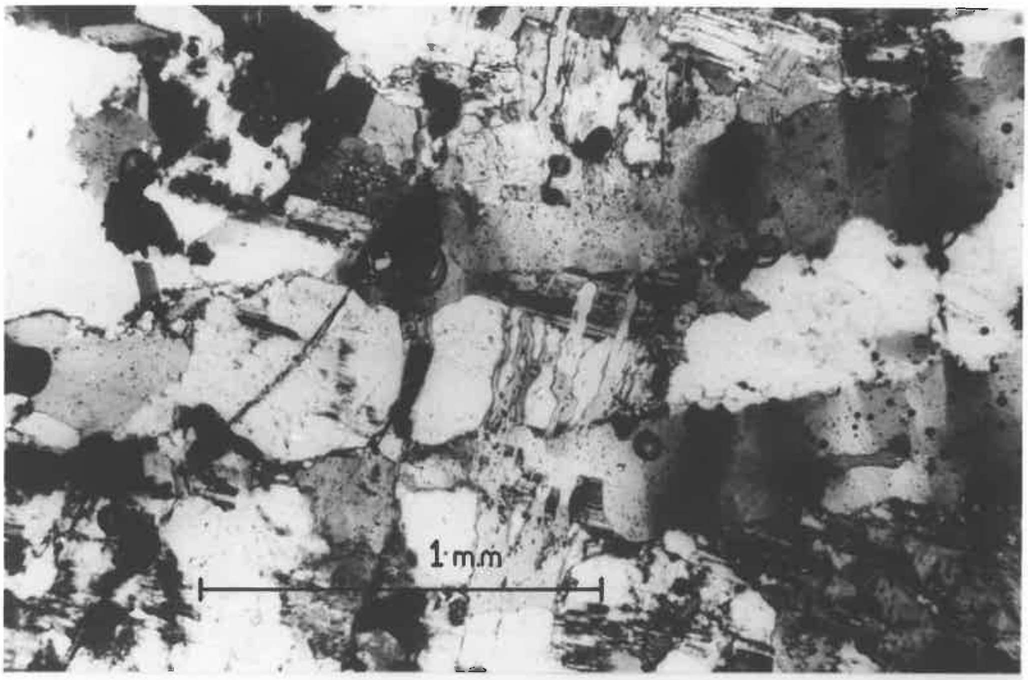


FIG. 26

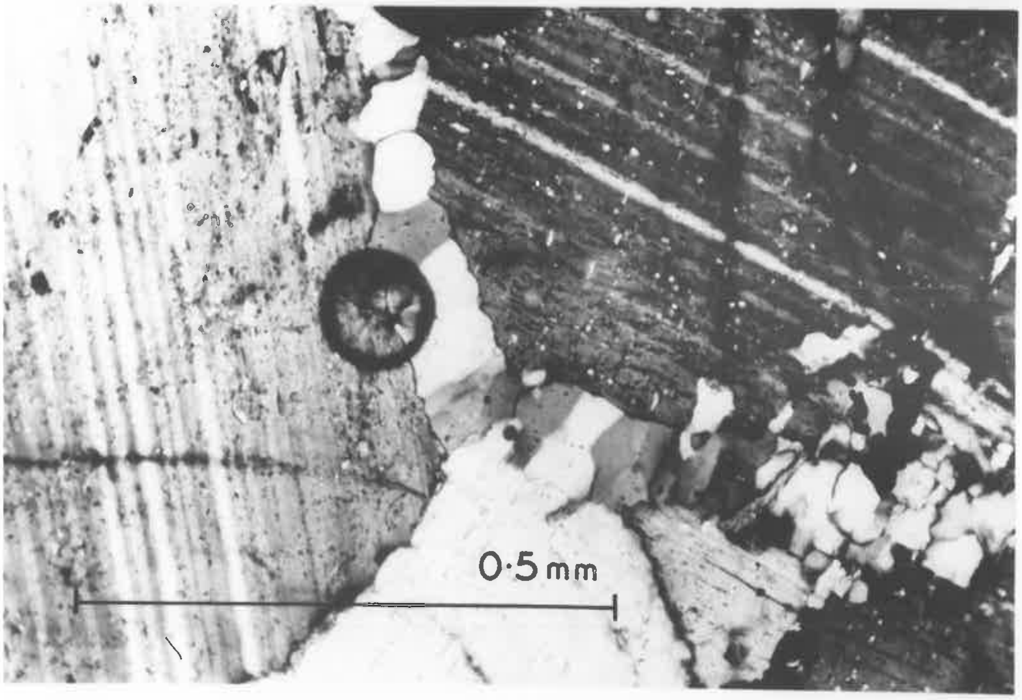


FIG. 27

quartz shows strain shadows. In some specimens the veins may be numerous and large (e.g. up to 5 mm. in 157-361a) and enclose microcline grains as well as cutting across the grains in the external aggregate. The mobility of quartz is also illustrated by the numerous polygonal aggregates which occur along the grain boundaries between feldspars (Fig. 27).

Plagioclase also shows many textures which could be interpreted as recrystallization or mobilization. Textures similar to the Kersbrook chequer-albite have been observed but are rare. Many sericitized plagioclases show water clear borders of albite. These water clear zones may also, though rarely, cut across the host grains (Fig. 28). The rims have the same composition as the sericitized plagioclase which is also albitic. Albite is not limited to microcline-plagioclase boundaries but occurs also at plagioclase-plagioclase boundaries. The boundaries between such grains are commonly controlled by the different twin lamellae, alternate lamellae "intruding" the adjacent grain more than the other (Fig. 29). In other cases the zone of clear plagioclase is not so well defined, and the boundaries tend to be highly irregular.

In many parts of the world, rims of albite around feldspars are common in both metamorphic and igneous rocks. Ramberg(1962, Plate I, Figs. 2, 4,and 6) describes textures very similar to those described here, but he does not report

Fig. 28. Rim of clear plagioclase around sericitized plagioclase. The surrounding grains are microcline. Quartz-feldspar gneiss. Spec. 157-361a. Crossed polars.

Fig. 29. Zone of clear plagioclase between sericitized plagioclase grains. Feldspar gneiss. Spec. 157-362. Crossed polars.

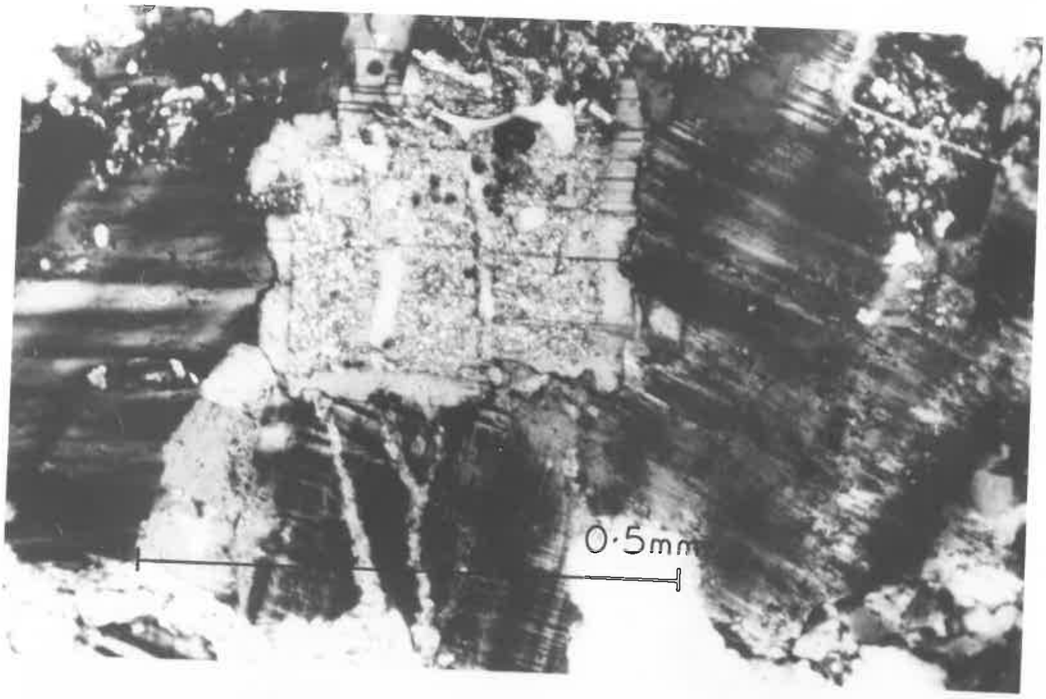


FIG. 28

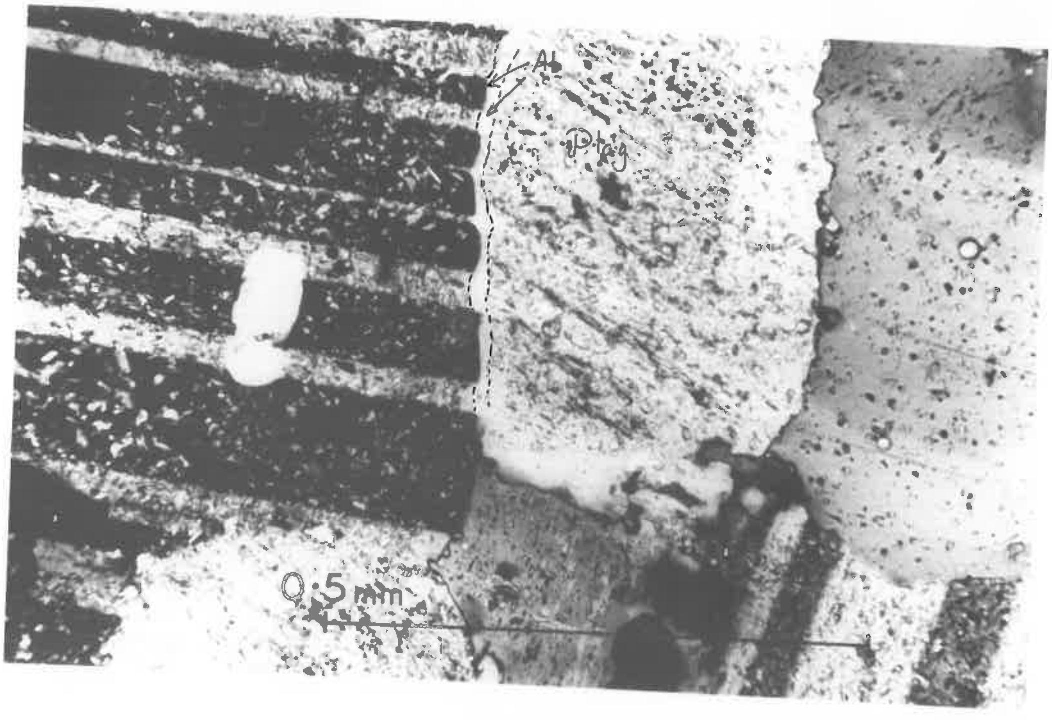


FIG. 29

albite at plagioclase-plagioclase contacts. He concluded that the plagioclase is derived from the migration of albite from microcline perthite.

Whitten (1957) has reported clear albite bordering somewhat altered plagioclase at plagioclase-plagioclase boundaries (his Fig. 10-D, p. 269) as well as at microcline-plagioclase boundaries and concluded that "late stage circulation of media along some inter-crystalline boundaries caused the alteration".

Agron (1950) concludes that albite rims may indicate two successive phases in metamorphism. Rogers (1961) considers such rims in granitic rocks to be merely a late stage direct crystallization of the albite.

In the Houghton Complex no single mechanism can be proposed which will explain all of the textural features of the albite rims. Some albite rims may result from the circulation of solutions as suggested by Whitten (1957). In many cases the solutions, migrating along grain boundaries, would simply have aided in the recrystallization of altered plagioclase or the migration of albite from perthites. In other cases where the source of the albite is not obvious (that is where albite rims non-perthitic microcline), it is necessary to assume that the solutions themselves carried albitic constituents. However these constituents could be derived from other parts of the rock and have not necessarily been introduced from outside sources.

OTHER PRIMARY METAMORPHIC ROCKS.

Apart from the feldspar gneisses described above, areas with recognizable higher grade metamorphic rocks are few.

The more important of these areas are:

1. Lower Hermitage - W. Houghton region.

These rocks are characterized by the assemblage quartz, microcline, biotite, minor plagioclase and muscovite. They are extremely well-layered (Fig. 30) and appear at first sight in hand specimen to be low grade metamorphosed sandstones and as such were included with the Torrens Group on the Adelaide 1 mile Geological sheet. However, both Spry and Webb subsequently grouped them with the "Archaean."

The rocks commonly possess a granular texture with quartz and microcline as the principal constituents. With increasing biotite the texture becomes more oriented with the quartz and feldspar fitting between the elongate biotites. Muscovite, where present, occurs as large randomly oriented poikiloblastic flakes enclosing all other grains.

The microcline is most commonly perthitic, film perthite predominating but occasionally irregular vein perthite also occurs. Many of the specimens show rims of clear plagioclase (An_{0-5}) (e.g. 157-847) surrounding the microclines. Individual grains of plagioclase of comparable size with the microcline are found in only a few rocks.

In a few rocks 10-15% plagioclase is present. This may indicate an early stage in the soda-metasomatism of these rocks

Fig. 30. Layering in mica gneisses. The layering varies from ½ cm. to 8 cm. and is separated by thin mica-rich layers.
Houghton School. (Grid. ref. 757 964).

FIG 30

as suggested by Webb (1953), but considerable amounts also of Ca would have to be introduced in order to produce the typical oligoclase-andesine-diopside gneisses.

The retrograde changes are similar to those of the feldspar gneisses. Plagioclase shows the greatest degree of alteration, namely to sericite. Biotite has altered to an olive-green variety and also to chlorite in some cases. Microcline, quartz and muscovite show little alteration. The striking feature of the alteration is that it is concentrated in zones and is responsible for the strikingly "bedded" appearance of these rocks in hand specimen. In thin section relatively unaltered rock alternates with thin layers (up to 5 mm.) consisting essentially of sericite. The sericite layers may be shear planes in the rock controlled by the mica foliation or they may represent layers which were more susceptible to alteration due to differences in original composition (e.g., higher proportion of plagioclase).

2. Torrens Gorge region.

In the region around the feldspar gneisses in the River Torrens and surrounding country, many of the rocks show pre-retrogression textural and mineralogical features. Rock 157-352 can be taken as fairly typical of this suite and occurs about ½ mile southwest of Prairie Bridge. In hand specimen the rock is moderately well-laminated, the laminations being composed of alternations of micaceous and quartzo-feldspathic material, the latter varying up to about 5 cm. in width. The layering is accentuated by the presence of sericitic partings parallel to the general layering. (In other rocks, particularly near the western margin of the complex in and around the River Torrens, the sericitic parting forms a foliation at an angle to the layering). In thin section the rock consists of an irregular granular aggregate of quartz, microcline and minor plagioclase, with about 5% colourless to olive-green pleochroic biotite flakes exhibiting a good preferred orientation parallel to the layering. Apatite, zircon, sphene and anhedral opaques are common accessories. Epidote is a rare accessory. The layering shows as prominent lensoid and irregular bands up to 2 mm. thick which are composed of a highly oriented mass of sericite grains enclosing a few relic quartz and feldspar grains. Specimen 157-352a illustrates a more advanced stage in the production of layering. The hand specimen could still be classified as a quartzo-feldspathic gneiss and differs from 157-352 in being highly laminated, the laminations (1-10 mm.) being defined by very thin sericitic partings separating quartzo-feldspathic material. In thin section the sericitic alteration is more widespread than suspected in hand specimen, irregular lenses of sericite being fairly uniformly distributed throughout the slide, with prominent zones of sericite crossing the slide and corresponding to the macroscopically visible foliation.

In many of the rocks the sericitic zones are concentrated in the more mica-rich layers of the gneisses and the sericite zones serve to accentuate the original foliation and layering.

It should be pointed out that these rocks are also interlayed with the actinolite-free feldspar gneisses of the Torrens Gorge and that they presumably have a related origin. The presence of abundant albite in some of the rocks has led Webb (1953) and Spry (1951) to the conclusion that the rocks have undergone soda-metasomatism and both these authors have correlated these rocks with the Houghton suite considering them as less metasomatized equivalents.

However, the petrographic differences and field occurrences suggest that these gneisses should not be considered to be simply related to the feldspar gneisses near Houghton. The feldspar gneisses with oligoclase-actinolite (or diopside) are distinct from the quartz-microcline-albite rocks of the Torrens Gorge. The latter are also interlayered with "normal" quartz-microcline-mica gneisses; the feldspar gneisses are not.

It is quite possible that soda metasomatism has played an important part in the origin of the quartz-albite rocks of the Torrens Gorge but such a conclusion should not necessarily be regarded as an indication of a similar origin for the feldspar-actinolite gneisses.

OTHER AREAS WITH RELIC HIGH GRADE METAMORPHIC ROCK TYPES.

Apart from the areas already discussed no large area within the Houghton Complex can be considered to have been free from large scale retrogressive changes. A few isolated outcrops show typical high grade metamorphic minerals, notably sillimanite, andalusite and garnet. Andalusite and garnet have each been found in three localities and sillimanite in six. Sillimanite shows a replacement relationship with andalusite in one specimen (157-13) but no other primary relationships have been observed. All three minerals occur as large (up to 2 mm.) generally subhedral grains, mostly showing considerable alteration to sericite (and chlorite in the garnets).

The distribution of these metamorphic minerals show no apparent zonal relationship. Spry (1951) states that (p. 123-4), "Sillimanite and garnet occur together in the Torrens Gorge, but that garnet alone occurs in the vicinity of Kersbrook. This suggests a series of sillimanite-garnet- and biotite zones towards the north, although the limits cannot be accurately delineated in the field." Sillimanite, however, has been found by the present author as far north as One Tree Hill (north of the area mapped) and its distribution appears to be random.

Other minerals, of less use in determining metamorphic grade, are common as relics in these rocks. Biotite, now largely altered, is present in many slides and perthitic microcline is present in most rocks not completely altered to quartz-sericite aggregates. The microcline is most commonly a film perthite. Plagioclase is also preserved in some rocks, commonly albitic in composition, though this composition may reflect retrogressive changes and need not be indicative of the original composition.

CHAPTER III.

STRUCTURAL FEATURES OF THE FELDSPAR GNEISSES.

Owing to the later retrograde structures present in most of the Houghton Complex exposures, information concerning the structures of the high grade gneisses is limited to the areas around Kersbrook and Houghton. Even within these areas structures which can be regarded as pre-retrograde structures are rare; mesoscopic folds can for the most part be related to later phases of deformation.

The only structure that can be definitely related to the high grade metamorphism is the layering. This layering, which is the result of differing concentrations of dark and light minerals, is commonly very regular (Figs. 5, 6 and 12) and varies in thickness from 2 mm. to greater than 2 cm. In other cases the layering is more irregular and resembles the layering of migmatitic gneisses (Fig. 31). Structures resembling cross bedding have been reported by Spry (1951), Plate VII, Fig. 3) and have been observed by the author (Fig. 32). Such structures are found only in the feldspar-actinolite gneisses and have not been observed in the more quartzitic rocks (arkoses and quartzites of Spry and Webb).

The development of a lineation depends on the degree of preferred orientation of the long axes of the actinolite grains, which commonly lie parallel to the layering. Many of the exposures which appear foliated in the field are found

Fig. 31. Irregular gneissic layering in feldspar-actinolite gneiss.
Paracombe. (Grid. ref. 789 937).

Fig. 32. Hooked cross laminations in feldspar-actinolite gneiss.
Houghton. (Grid. ref. 771 964).

FIG. 31

FIG. 32

to be lineated when cut surfaces are examined. Due to this difficulty of determining lineation in the field only a limited number of lineation measurements could be taken; however, these show a fair preferred orientation for the mass as a whole.

Mesoscopic folds in both the Houghton and Kersbrook masses are almost completely lacking. An indistinct antiformal structure can be seen in the Little Para River at Inglewood (Grid. Ref. 768 977); the structure is symmetrical and at least 10 m. high. Both limbs dip about 45° and the crestal region is not visible. Neither the style of the fold nor its axis could be observed. The presence of larger folds in the Houghton region is suspected from the curved trend of disconnected outcrops in some areas and also in the distribution of layering attitudes in π diagrams (Plate II). The correspondence of the lineations with the statistical fold axes so deduced is only approximate and so it cannot be stated with certainty whether or not the large scale folding took place contemporaneously with the formation of the lineation.

A number of folds of doubtful age have been described by Webb (1953) and are well exposed in the road cuts along the River Torrens just north of Prairie bridge. These folds have a concentric style and except for one example are larger than 10m. amplitude. The axial surfaces of these structures dip about 70° to the east; a new foliation (equivalent to the retrograde foliation) is developed in part of the area and is sub-parallel to the axial surfaces of these folds, a genetic relationship being suggested.

Folding in the Kersbrook mass was not observed and the layered attitudes show a marked planar preferred orientation (Plate II).

Structures not directly related to the layering and lineation are present in both the Houghton and Kersbrook masses. Cross-cutting veins of actinolite-rich material are fairly common and in some cases the actinolitic material is so abundant that it completely encloses patches of the feldspar gneiss and the rock has the appearance of an agmatite. Pegmatites composed of coarse-grained diopside (up to .5 cm. diameter grains) with minor plagioclase are also found and were called "yatalite" by Benson (1909). In rarer cases, plagioclase occurs in veins, some of which are cross-cutting. The gneisses have been mineralized in part, small veins of quartz and haematite being extremely common; uraniferous minerals have also been reported (Dickinson and others, 1954).

These structures tend to obscure the earlier fabric in some areas but in many parts of the feldspar gneisses layering and lineation (or foliation)are still dominant structures. Benson (1909) was apparently not influenced by the layering when he referred to the Houghton feldspar gneisses as diorites and syenites. Spry's (1951) account of the area is the first to discuss the layering in any detail and he concludes that the layering "may be treated as bedding and mapped similar to a normal sediment" (p. 125). While it is possible that these structures represent original bedding features they may well represent the combined effects of metamorphic differentiation, deformation and transposition. The mobility of the components of the feldspars and amphiboles is seen by the numerous intergrowth textures in thin section and by the cocentration of actinolite in cross-cutting structures such as joints and small faults (Fig. 12); and by the concentration of dark minerals into pods and pegmatitic veins. If such ready structural control by minor features can be seen, it seems likely that a well-developed metamorphic foliation, especially if of a phyllonitic type, could be expected to control the deposition of new minerals during metamorphic differentiation. "Cross bedding" structures could then be interpreted as transposition structures associated with the development of a foliation.

A somewhat different picture is obtained from the layered gneisses surrounding the Houghton feldspar gneisses. In the Torrens Gorge many of the gneisses consist of inter-layered quartzo-feldspathic material and mica-rich gneiss. The layering tends to be ill-defined (except where accentuated or produced by later changes) but where sharp, the rock types represent distinct changes in composition. It may be that these changes in composition represent original differences in composition and are probably metamorphosed transposed bedding.

Whatever the origin of the layering it has been affected by later deformations, especially in the more micaceous rocks. The resulting structures will be discussed later.

ORIGIN OF THE FELDSPAR GNEISSES.

As already indicated, earlier writers considered the feldspar gneisses to be igneous in origin ("diorites" of Benson 1909, England, 1935, Mawson 1926) and later, sedimentary in origin (Spry 1951; Webb 1953). The igneous origin was advanced largely on account of their massive appearance, compared with the surrounding schists, and their chemical similarity to known igneous rocks. Veins of similar composition to the "diorites" were thought to be pegmatitic phases. The rocks were recognised as metamorphic rocks but the banding was not considered in detail.

After a study of the field relationships and the small scale structures Spry (1951) concluded that the rocks were metasomatised sediments. Webb's (1953) detailed mapping of the disconnected outcrops showed that the form surfaces were folded or warped on a large scale.

A discussion of the origin of the feldspar gneisses is hampered by a lack of any objective criteria for distinguishing between an igneous and a sedimentary origin for the features observed. Field relations should be of considerable importance in such a discussion but unfortunately the Tertiary erosion surface has obscured all the contacts. In areas where the feldspar gneisses are in close proximity to the surrounding schists and gneisses (i.e. east of Lower Hermitage), the contact is seen to be parallel to the (retrograde) foliation in the schists and at a small angle to the foliation in the feldspar gneisses. Brecciated rocks are found at the northern end of the contact and prominent quartz-haematite veins occur at intervals along the contact, which is interpreted as a fault. The lack of intrusive contacts as noted by Spry may be explained if the contacts are tectonic in origin.

Spry also noted gradational contacts with the surrounding schists. Such gradational contacts appear to be due to the greater degree of retrogressive effects in the surrounding schists. His sequences of feldspar gneiss through transition gneiss to schist would be interpreted by the present author as a sequence of unaltered feldspar gneiss through a retrograde border zone in the feldspar gneisses to highly retrograde rocks derived from schists and gneisses of an original chemical and mineralogical composition entirely different from the feldspar gneisses.

The layering and cross laminations (Fig. 31) in the feldspar gneisses have also been used as evidence that the rocks were originally sedimentary in origin. However, as Dietrich (1960a) showed, layering in gneisses can be the result of (1) relic sedimentary layering, (2) relic igneous layering, (3) various processes of migmatization and permeation, (4) metamorphic differentiation. Dietrich attempted to set up criteria for distinguishing these processes but in reviewing the evidence showed a strong bias towards a sedimentary origin for banding in gneisses, a fact which he later admitted (Dietrich 1960b). Banding, similar to that seen in sediments, is not common in igneous rocks but has been described from ultramafic rocks (Jackson, 1961), from syenites (Upton, 1960) and from granites (Harry and Emeleus, 1960). Small scale sedimentary features such as cross bedding and trough banding are also seen in these igneous rocks (especially well figured in Upton, 1960).

Layering produced by metamorphic differentiation may also resemble sedimentary layering in its regularity and persistance. Evans and Leake (1960) have shown that the layering in amphibolites of Connemara is parallel to the foliation in the rocks and is not folded. They conclude that the layering is produced by intense metamorphic differentiation in rocks which were originally igneous. Walker and others (1960) in a comparative study of ortho and para-amphibolites concluded that the differences between the two types decreases with increasing grade of metamorphism especially where the rocks are affected by metasomatism.

Layering can also be produced by intense transposition and by shearing. Bowes and Jones (1958) describe pseudo-sedimentary features in greywackes and conclude that the structures are tectonic in origin. "Cross bedding" is particularly easily produced by transposition of an earlier layering (not necessarily sedimentary layering).

The layering in the feldspar gneisses is not therefore as conclusive of a sedimentary origin as was thought by Spry. If the rocks were of a lower grade of metamorphism (they have reached at least as high as the upper amphibolite facies) it might be possible to determine the origin of the layering. However the composition of the rocks suggests that if they were sediments they have been fundamentally changed by metasomatism (as Spry suggested). The metasomatism combined with the high grade of metamorphism makes it extremely unlikely that sedimentary structures could still be distinguished from igneous or metamorphic structures in these rocks.

Other evidence which has been used by Spry is that of the large scale heterogeneity of the feldspar gneisses. The present author believes that this heterogeneity was brought about by several processes. Firstly, there may have been original differences in composition in various parts of the mass. These differences may have been in either igneous or sedimentary rocks. Secondly, the redistribution of material, possibly accompanied by metasomatism, would have increased the compositional variations in the mass. Thirdly rocks suffering different degrees of retrogression would impart additional inhomogeneities to the rocks. (The diversity of rock types in the mass has been over-emphasized by Spry who included some granitoid rocks (recrystallized "arkoses") with the feldspar gneisses. These rocks are quite distinct from the feldspar gneisses and occur for the most part as "islands" in the schists).

Mineralogically and chemically the rocks do not match either common sediments or common igneous rocks. Benson (1909) and England (1935) both calculated norms from their analyses and matched the rocks with some rock types in the C.I.P.W. classification, notably syenites and diorites. However, the compositional similarities are only approximate and if the feldspar gneisses were igneous, their chemical composition has been modified by later additions or redistribution. If the rocks were derived from sediments, their composition must have been even more profoundly altered. Both the mineralogical composition and the chemical analyses (Appendix I) show that the feldspar gneisses have high contents of Na_2O, CaO and MgO. By comparison, the surrounding schists are characteristically poor in Ca and Mg-rich minerals. Similar schists and gneisses from Humbug Scrub (a few miles north in the same inlier) show low contents of CaO and MgO (Alderman 1938, pp.175-6). In particular the CaO content is very low (0.26 and 0.43% CaO in the schists and 0.86% in the augen gneisses). These values are lower than those given by Clarke (1924) for average shale (3.11% CaO) or by Pettijohn (1957) for a variety of shales (1.00 - 2.17% CaO).

If the feldspar gneisses were formed by soda metasomatism of sediments, these sediments would therefore have to be calcium-rich as suggested by Spry. This hypothesis is not unreasonable but is extremely difficult to establish with certainty. A necessary corollary to this hypothesis is that the soda metasomatism is selective and affects only the calcium-rich rocks and not the surrounding schists which are poor in both calcium and sodium. In addition soda metasomatism must have affected <u>all</u> of the calcium-rich sediments, as no soda-poor calcium-rich rocks have been discovered.

The mineralogical, chemical and structural evidence is therefore considered to be inconclusive and the present author considers that the origin of the feldspar gneisses is unknown.

CHAPTER IV.

RETROGRESSIVE CHANGES IN THE HOUGHTON COMPLEX.

In previous sections only those retrograde changes incidental to the understanding of the high grade metamorphic rocks have been considered. However, most of the rocks outside the zones of feldspar gneisses have been more or less fundamentally changed until now they represent phases stable in the greenschist facies of metamorphism.

In hand specimen these retrograde rocks vary considerably in appearance. The less altered rocks appear as quartzo-feldspathic gneisses in which there is a well-defined schistosity commonly parallel to the layering. This schistosity, as has been noted before, is due to narrow zones of sericite, possibly representing selective alteration of differing compositional layers in the gneiss. With greater degree of retrogression the rocks are well-layered and resemble the flaser gneisses described by Hamilton (1960). The layering is very variable in thickness (1-10 cm.) and reflects differences in amount of alteration and the original composition of the rocks. In some outcrops the layering appears to be uniform with individual layers persistent over several metres; in others the layering is distinctly lensoid and rather irregular.

Where the amount of sericitization is large the rock loses its well-layered appearance and even the schistosity is irregular and the rocks do not cleave parallel to the foliation. However in thin sections and polished blocks the schistosity is still easily observed.

In many of the rocks, larger quartz and feldspar grains have survived as augen. Alderman (1938) has described these augen in the Humbug Scrub region and concluded they were the result of growth of crystals by metasomatism. Hossfeld (1935) thought they represented the uncrushed remnants of interfolial pegmatitic material.

The Humbug Scrub augen have a large grain size; individual subhedral feldspar grains larger than 1 cm. are fairly common. The margins of the augen are always bounded by zones of abundant sericite and it appears that the augen have been distinctly modified in shape during the later retrograde metamorphism.

In the area examined by the author the augen are not as large or as coarse-grained as the Humbug Scrub augen. The augen consist of a granular aggregate of quartz and feldspar, in marked contrast to the subhedral textures of the larger feldspars in the Humbug Scrub examples. It is thought that the differences between the two areas reflect differences in the degree of retrogression and recrystallization, the southern area showing very few examples of original textures.

The mineralogical changes which the high grade minerals have undergone have been discussed in the previous chapter and only a discussion of the textural features of this alteration will be discussed here.

In one type of alteration the mineralogical changes are brought about without any noticeable distortion of the shapes of the original minerals nor does the sericite show any visible degree of preferred orientation. Perfect cross sections of sillimanite are preserved in sericite (Fig. 33)and presumably many of the patches of unoriented sericite in the rocks could represent this same type of replacement (Fig. 34). The end product of this type of alteration is a rock composed almost entirely of a fine-grained quartz-sericite aggregate with relics of quartz and rare microcline.

In another type of alteration sericite is concentrated into irregular sub parallel zones and within these zones the sericite tends to show strong preferred orientation (Fig. 35). Relics of quartz and microcline are not uncommon in these zones; plagioclase and biotite occur rarely. These zones tend to be lenticular but in some cases can be traced across a single hand specimen without difficulty. With an increase in the number of oriented zones the rocks become more schistose than layered (Fig. 36). In the earlier stages of alteration the relic minerals show no particular habit-preferred orientation to the foliation but in many of the rocks in a more advanced stage of alteration the relics tend to lie with their long axes in the foliation, although exceptions do occur (centre of Fig. 36).

Rocks resembling true phyllonites are uncommon and are confined to limited zones within the feldspar gneiss mass (e.g. 157-908, 117). They consist of thin lenses of finegrained quartz in a very fine-grained quartz-sericite

Fig. 33. Pseudomorphs of sericite after silliman-
ite.
Retrograde schist. Spec. 157-212.

Fig. 34 Relics of quartz and microcline in a
ground mass of non-oriented sericite
and fine-grained quartz.
Retrograde schist. Spec. 157-1031.

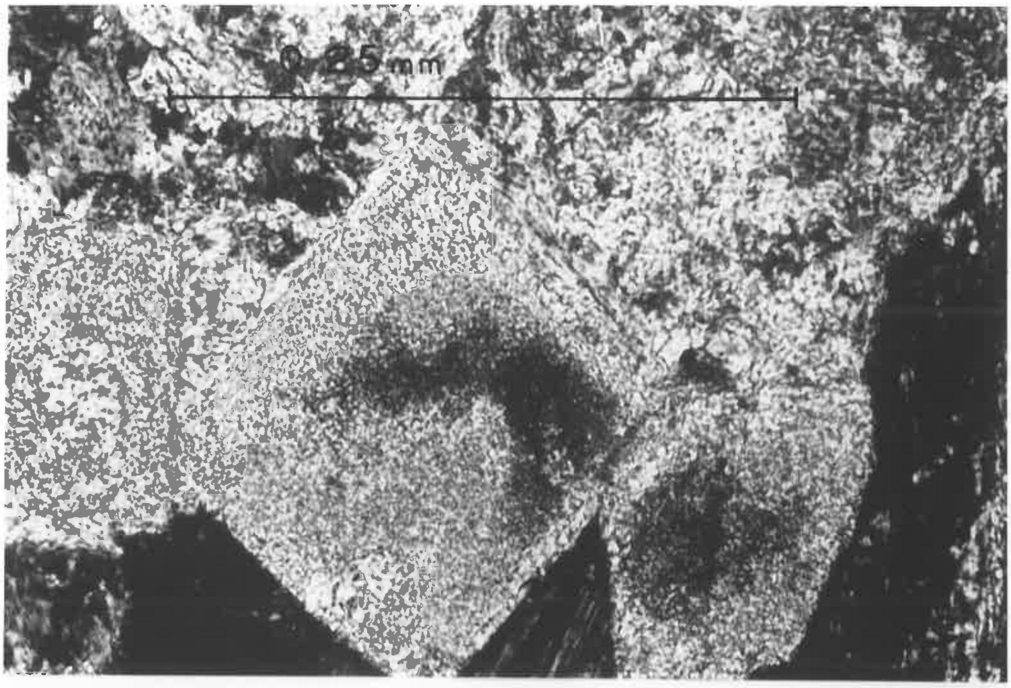


FIG. 33

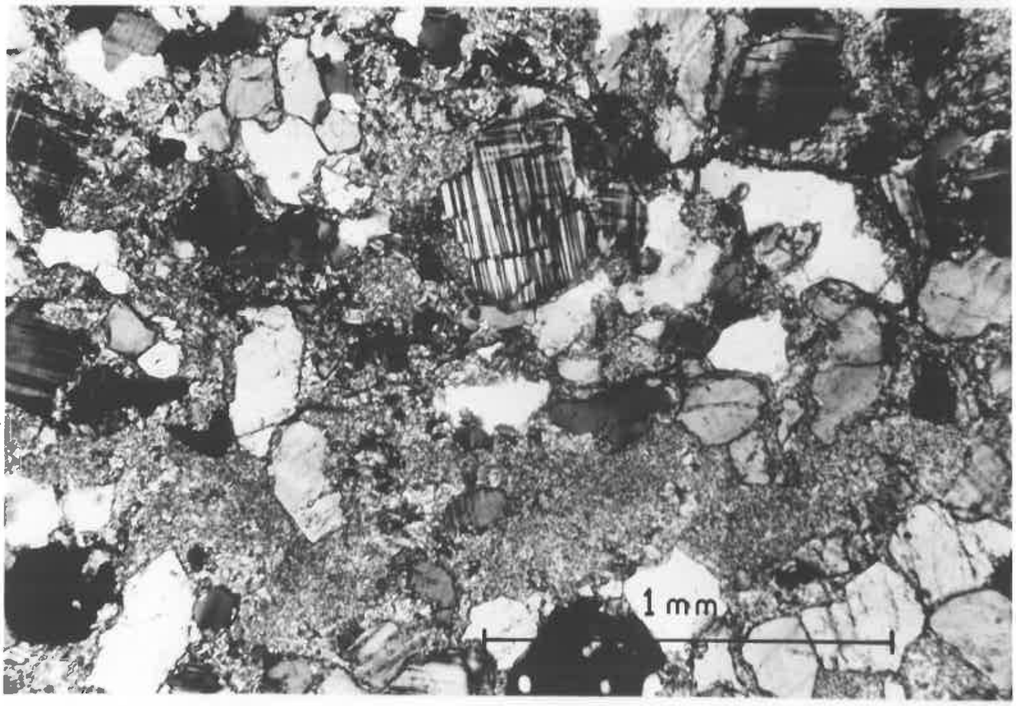


FIG. 34

Fig. 35. Zone of fine-grained quartz-sericite crossing retrograde gneiss.
Flaser gneiss. Spec. 157-907.

Fig. 36 Foliated blastomylonite with sericite concentrated in lenticular zones.
Blastomylonite. Spec. 157-102.

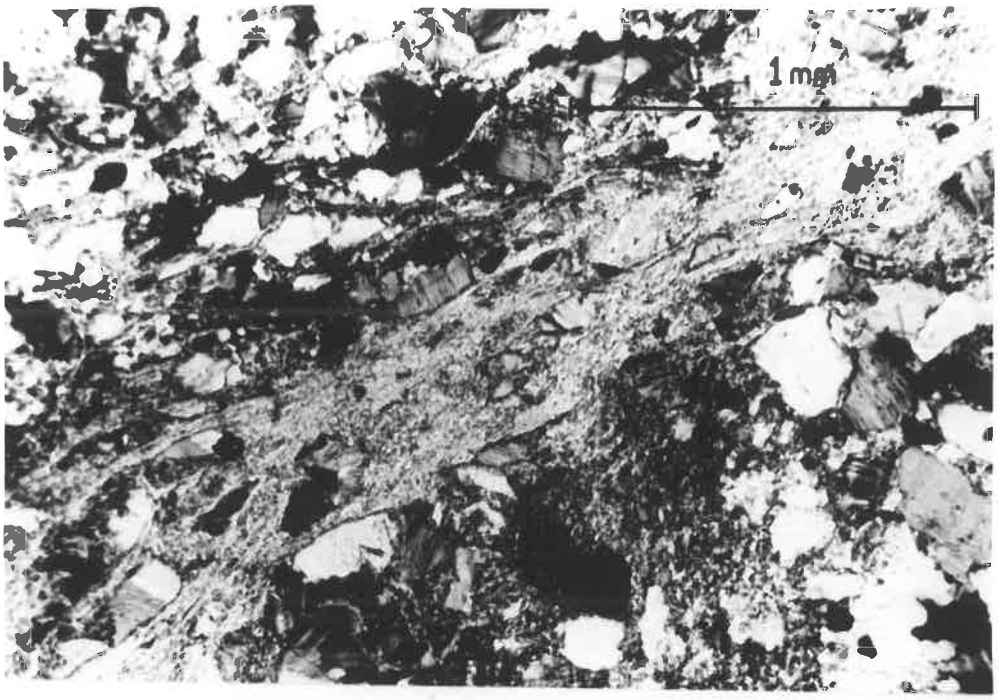


FIG. 35

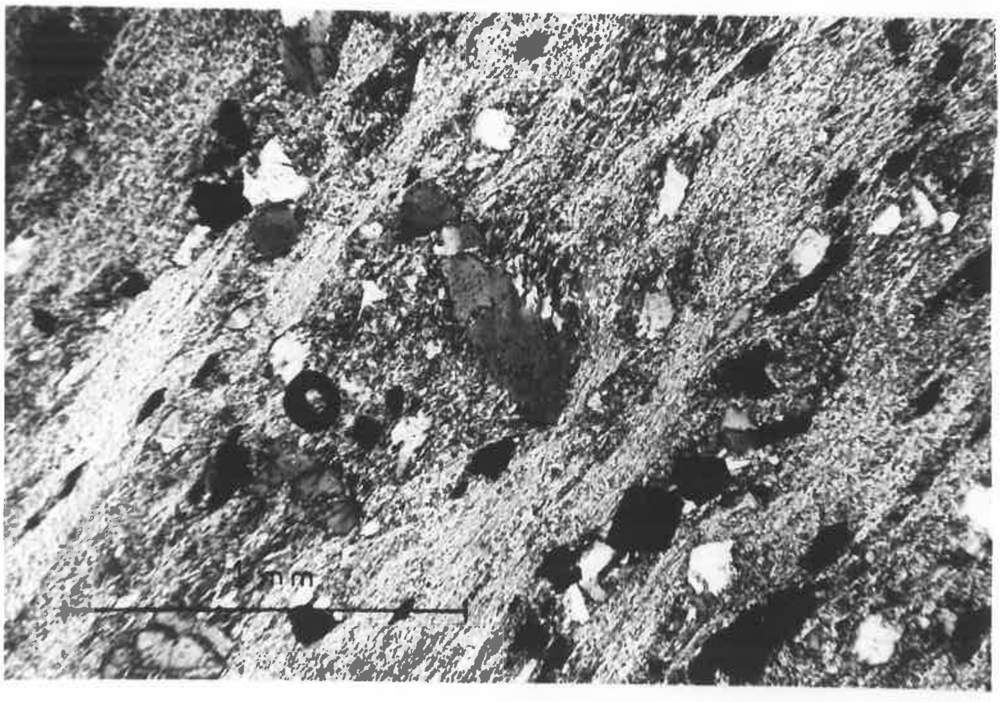


FIG. 36

matrix (Fig. 37). In the most advanced phyllonite observed (157-908) the rock has the appearance of a roofing slate and in thin section hardly any coarse-grained relics are observed, the rock being a finely foliated and extremely fine-grained ($<$ .005 mm.) quartz-mica aggregate with small lenticular masses of slightly coarser material (.01 mm.)

The common observation that these zones cut across the non-oriented portions suggests that the orientation process is later than the production of non-oriented sericites. Whether the sericitization and later orientation represent two distinct episodes of alteration or two phases of the same alteration is not known.

Associated with the retrograde changes is a considerable degree of deformation of the relic minerals. Although deformation effects are visible even in relatively unaltered rocks they are most apparent in the foliated sericite rocks. Quartz shows a greater variety of deformation features than the other minerals, and exhibits undulatory extinction, fracturing, marginal granulation and recrystallization. The undulatory extinction varies from a broad sweeping of the extinction across the grain, to fairly sharply defined bands of extinction. The extinction bands are sub-parallel to the c axes and in the course of petrofabric investigations no exceptions to this were observed. (This is the common orientation of extinction bands reported in the literature, e.g. Hietanen, 1938). No distinct correlation exists between the amount of rotation of the c axes in single quartz grains and the evidence of deformation in other minerals. In fact quartz can deform to a marked

Fig. 37. Lenses of quartz in a highly foliated quartz-sericite matrix.
Phyllonite. Spec. 157-117.

Fig. 38. Rhombic strained quartz in quartz augen in a fine-grained quartz-sericite matrix.
Blastomylonite. Spec. 157-76.

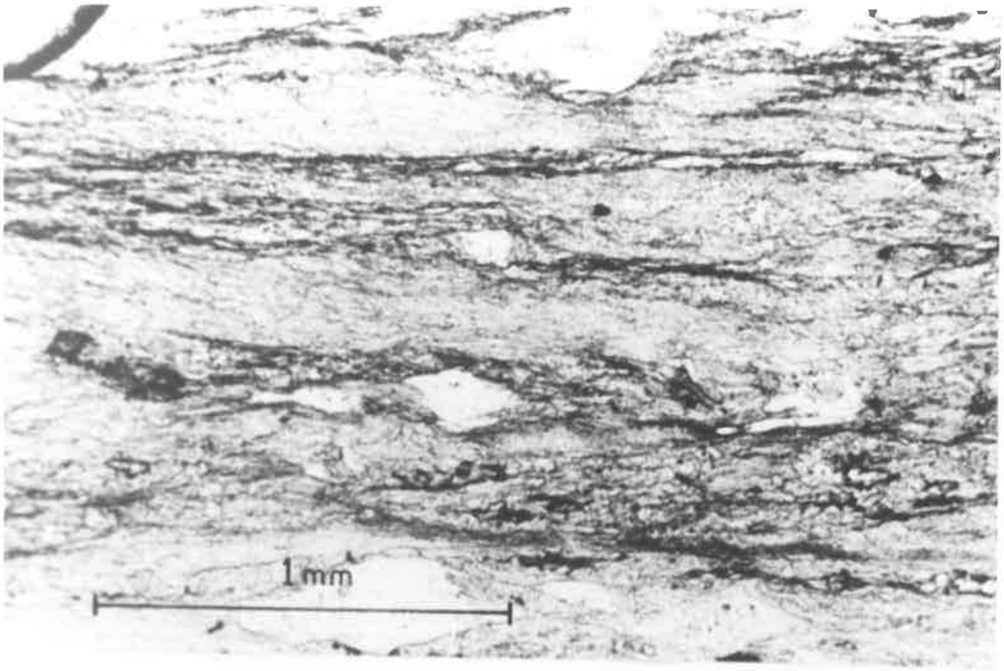


FIG. 37

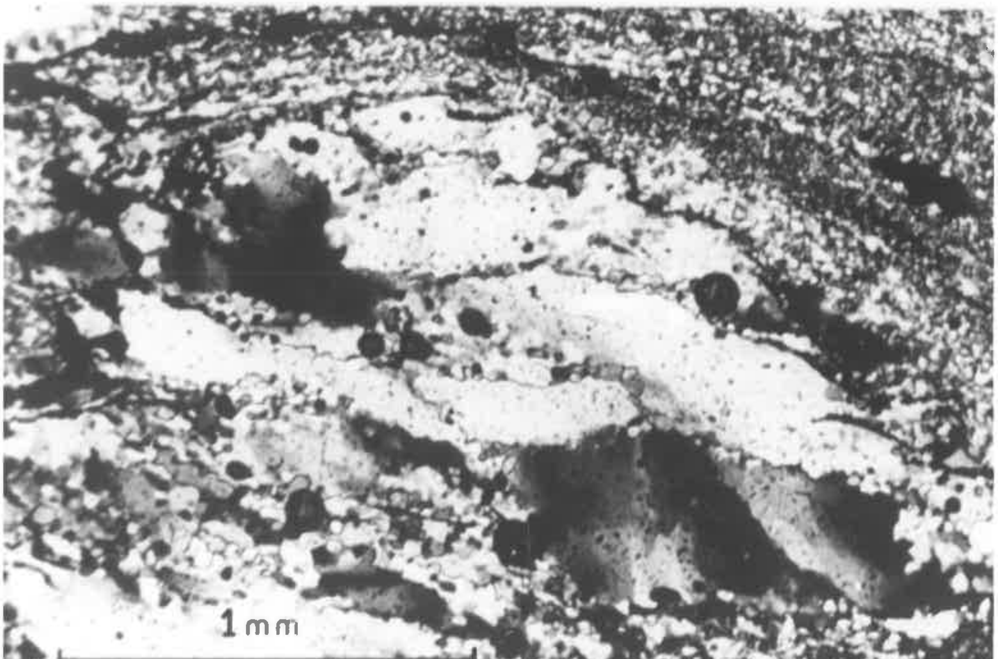


FIG. 38

extent with no evidence of deformation of feldspars (cf. the deformation of "sand" grains in dolomites, Chap. V). Fractures in the quartz grains are quite irregularly developed and show no marked tendency to occur parallel to the extinction bands as reported by Bailey and others (1958). In many cases the quartz grains are lenticular and suggest the operation of inclined shear planes which are apparent in some quartz-rich augen (Fig. 38).

Many grains of quartz have recrystallized to a mosaic of unstrained polygonal grains. The boundaries between these grains are gently curved and smooth and tend to meet in triple points, but the angles at which the faces meet commonly depart from the equilibrium angle of 120° (Voll, 1960).

The presence of abundant sericite in the polygonal aggregates distinctly modifies the quartz boundaries. The quartz-quartz interfaces tend to lie perpendicular to sericite $\{0001\}$ planes and tend to butt against the sericite grains rather than form quartz-quartz triple points. The quartz a nd mica grains are of comparable grain size even where different grain sizes for the quartz grains are seen in one thin section. (Fig. 39).

The distribution of the quartz polygons in relation to the large grains is quite variable. The large grains commonly show sutured boundaries (e.g. Fig. 38) and in many cases the sutured appearance is due to a single layer of intergranular polygonal grains whose orientations are closely related to the adjoining large grains (Fig. 40). In a few cases stringers

Fig. 39. Coarse quartz-muscovite aggregate in a finer sericitic matrix. The large grain is microcline.
Blastomylonite Spec. 157-87A.

Fig. 40. Intergranular recrystallization along sutured quartz boundary.
Retrograde gneiss. Spec. 157-1188.

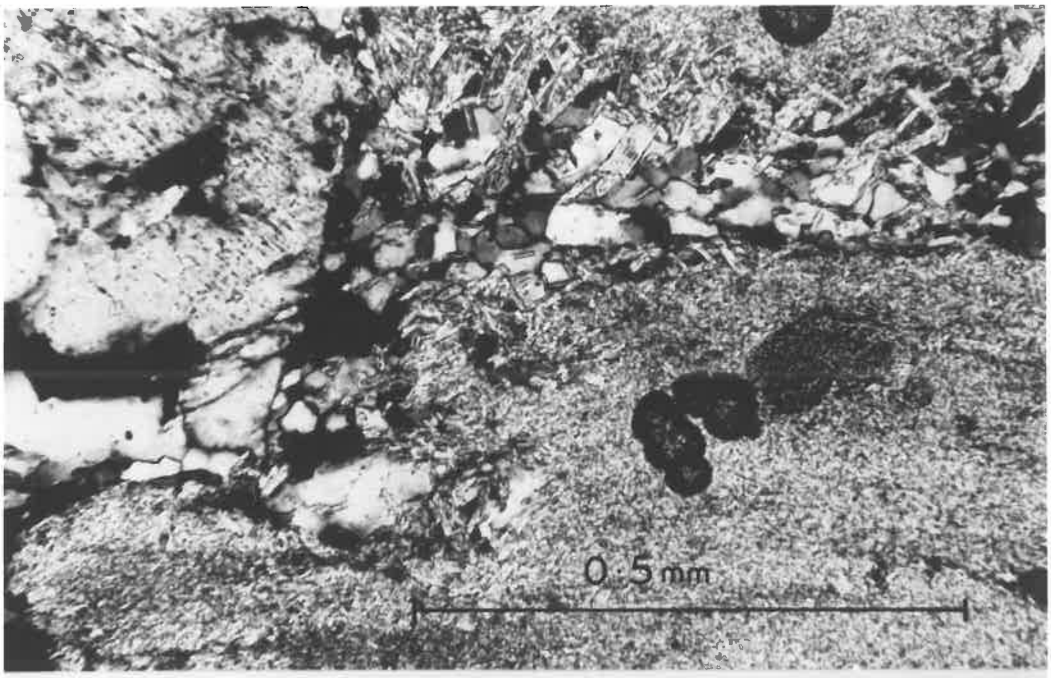


FIG. 39

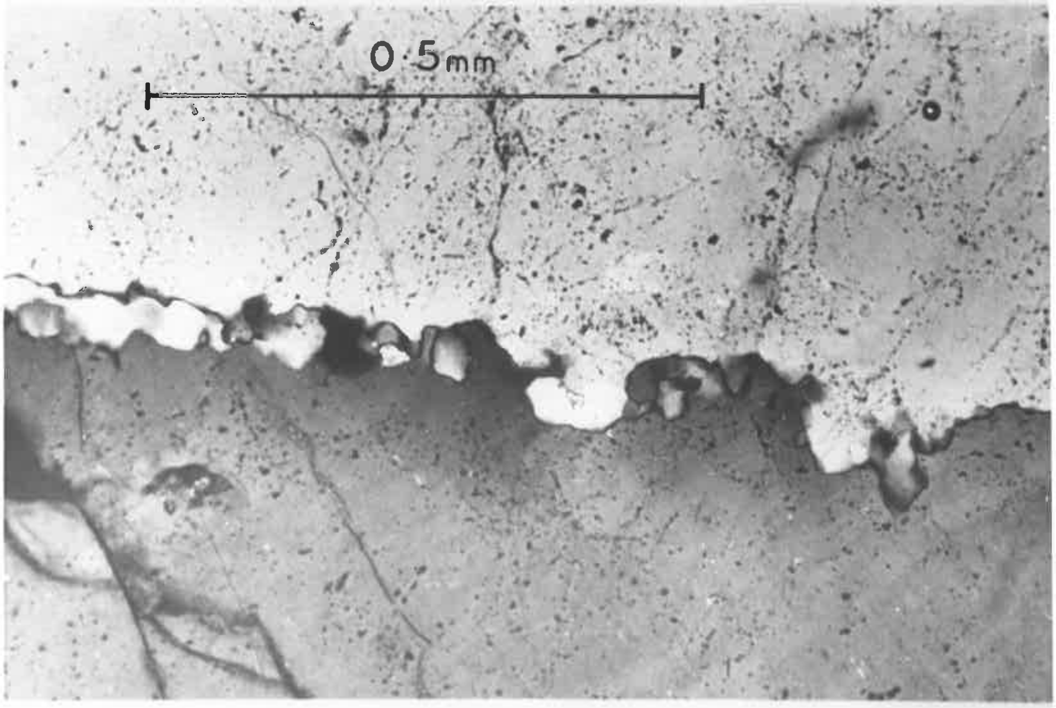


FIG. 40

of polygonal grains cross single quartz grains (Fig. 41) and the preferred orientation of the c axes of the polygonal grains is fairly close to that of the host grain (Appendix II). These features may represent incomplete stages in the recrystallization of strained grains; more commonly the process is more advanced and patches of polygonal grains are observed. In many cases only relics of the large quartz grains are observed in the polygonal aggregate. In other cases the polygonal grains occur as tails on lenticular quartz grains.

The features described above are interpreted as changes brought about by deformation. Undulatory extinction and extinction bands of quartz have been interpreted by Bailey and others (1958) as the result of plastic deformation by bend gliding, and they have also shown that such features represent polygonization of the strained grains. The wide-spread development of fine-grained polygonal quartz is interpreted as recrystallization and grain growth from a stressed aggregate, the relics of which are still visible (Voll, 1960; Griggs and Handin, 1960). In quartz-sericite aggregates the textures suggest contemporaneous recrystallization of these minerals and possibly the widespread sericitization is contemporaneous with the recrystallization. This possibility is supported by the generally greater degree of sericitization in the more deformed aggregates. There is, however, a strong possibility that the textures observed represent two phases of deformation, a retrogressive phase prior to the deposition of

Fig. 41. Trains of polygonal quartz grains crossing large strained grains. Augen schist. Spec. 157-172.

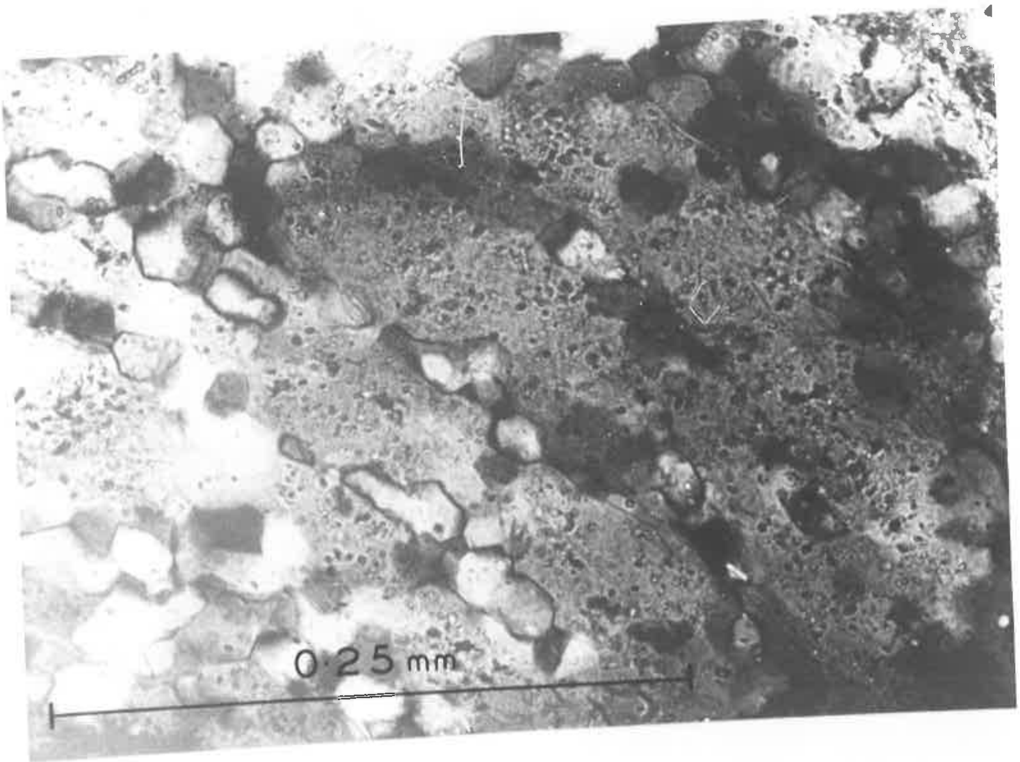


FIG. 41

the Upper Precambrian sediments and a second phase associated with the deformation of these sediments. Evidence for these two phases of deformation and retrogression is discussed later (Chaps. V and IX).

MESOSCOPIC STRUCTURES RELATED TO THE RETROGRESSIVE METAMORPHISM.

As noted previously a strongly developed foliation is the dominant structure in the cataclasites which form the bulk of the Houghton Complex. In most cases the foliation is not continuous over a whole outcrop and consists of a series of lensoid zones containing abundant sericite. These sericite zones are commonly irregularly distributed but may impart a crude layering to the rock. A lineation is also present in many outcrops, and results from the intersection of the somewhat irregular foliation zones and also coincides with the long axes of quartz and feldspar augen.

Fold structures associated with this foliation are rare except in the southern part of the complex. (Fig. 42). The style of the folds is variable, 42A, D and F showing some concentric characteristics and 42B showing similar style. In all of these cases the folds are in quartz-feldspar rich layers in sericitic schist. Folds are not preserved in very thin layers, the layering being completely transposed into the foliation. Even though many of the microscopic folds have a concentric style, the movements in the rock as a whole appear to have been dominated by the foliation.

Some of the larger folds in the Torrens Gorge (Webb, 1953)

Fig. 42 Style of folds associated with the retrograde foliation. Houghton Complex.

Castambul. (Fig. 42, A and D at Grid. ref. 762 925).

(Fig. 42 B, C, E and F at Grid. ref. 762 926).

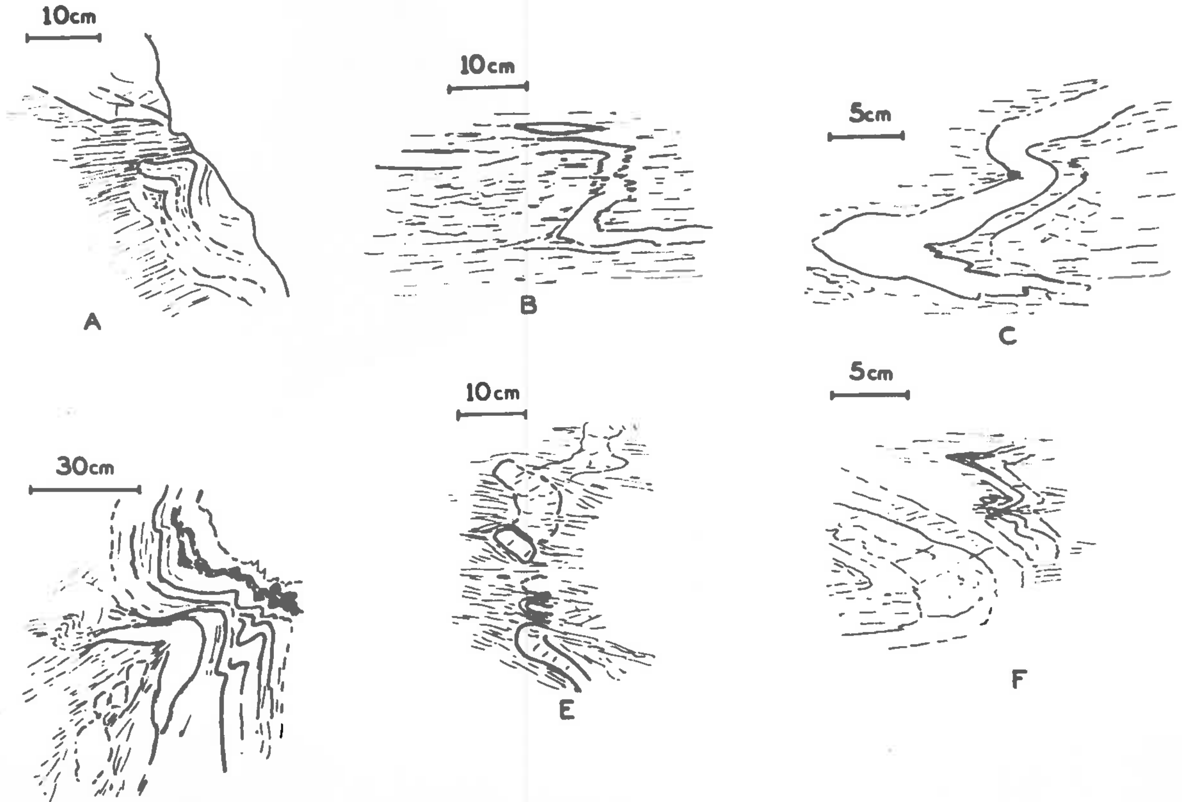


FIG 42

have a concentric style although many of them have been transposed in part. The sericitic foliation is also associated with these folds. In some of the thicker or more massive gneisses concentric folds occur with no associated foliation. (Fig. 43). The folds apparently are S-active folds and the folding has taken place by slip along planes of weakness, now sericitic partings, in the original gneiss.

Fig. 43. Small scale concentric folds in feldspar gneiss.
Paracombe (Grid. ref. 788 939).

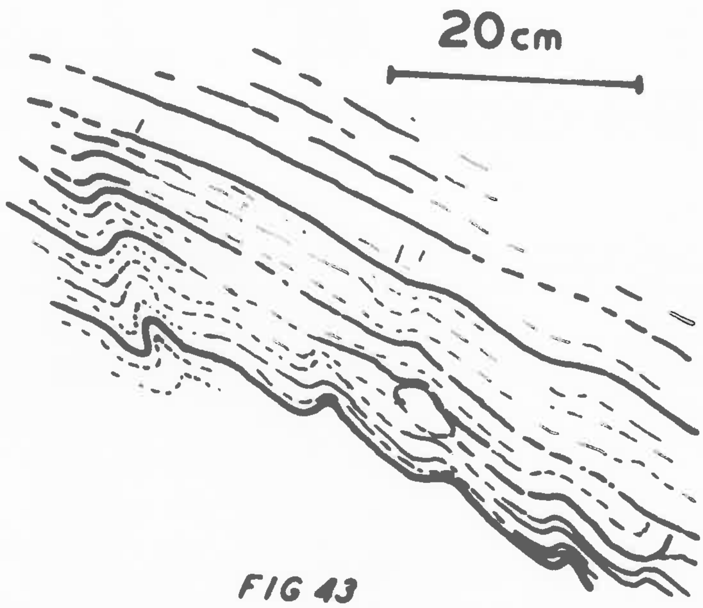


FIG 43

CHAPTER V.

THE PETROGRAPHY OF THE UPPER PRECAMBRIAN SEQUENCE.

In much of the area mapped the contact between the older and younger Precambrian sequence is faulted. On the western side of the Houghton Complex, the contact dips steeply and many of the beds in the younger series are overturned. Contacts along the eastern margin are less disturbed and although faulting may have occurred, horizons near the base of the sequence can be traced parallel to the contact for considerable distances. Some degree of shearing is observed at most of the exposed contacts, notable exceptions being the conglomerates occurring as small outliers in the "Archaean" at Houghton and near Castambul.

Faulting along the western contact is so extensive and the relation between these "basal" conglomerates and the rest of younger beds so indistinct that it is not known how much of the Torrens Group is missing. Certainly to the north, beds much higher in the sequence are seen against the "Archaean" than in the south. This makes the River Torrens area a particularly unsuitable area in which to establish the type sequence for a group.

Publications concerning these beds of the Adelaide "system" have been mentioned in the introduction. The sequence has been described formally by Mawson and Sprigg (1950) and their description of the Torrensian is reproduced here:-

"Torrensian Series Thickness 7450 - 9450 feet

(Range due to variable thickness of basal sandstone)

Glen Osmond slates with occasional thin dolomite bands	1540'
Beaumont Dolomites and interbedded slates	450
Upper slates (and phyllitic phases) with minor quartzites	1000
Stoneyfell(Mt. Lofty) quartzite, in part arkosic and argillaceous	1000
Lower slates (and phyllites) with included minor quartzites	1100
Montacute Dolomite: blue and grey dolomites, limestones and sedimentary magnesites with chert bands and minor quartzites	430
Slates (and phyllites) with minor quartzites	680
Castambul Dolomite	150
Slates(and phyllites)	100
Aldgate Sandstones, mainly argillaceous and in part ilmenitic; also with interbedded lenticular conglomerates and recurrent argillaceous bands. This formation rests with violent unconformity on the underlying Barossian Complex.	1000 - 3000 ft

In the area mapped the Stonyfell quartzite is the youngest formation encountered.

Detailed observations on the stratigraphy and correlation of the various rock types were not attempted but certain observations concerning the nature of the sediments and sequence are relevant to the present discussion.

Basal beds.

The basal beds of the "Proterozoic" are preserved in only a few places on the western side of the "Archaean". Where present, they consist of polymictic conglomerates with fragments up to 4 cm. diameter. (Good exposures are north of Houghton and east of Castambul). The rocks are well-layered and vary from grits to conglomerate. These beds have a rough cleavage subparallel to the cleavage in the underlying "Archaean" (however, no case was observed where the two sequences are in actual contact and this conclusion is based solely on comparison of measurements on adjacent outcrops).

The localities at Houghton are particularly instructive. Specimen 157-1025 is from a small outcrop just north of the main road (Grid ref. 763-975) and in hand specimen appears a schistose conglomerate. It contains one large pebble (2 x 1 x ? cm.) of medium-grained quartz-feldspar rock, with many smaller (ca 5 x ? x 2 mm.) pebbles of feldspathic material. Many of the fragments have a marked tendency for the short axis to lie in the plane of schistosity. No marked lineation of long axes of pebbles is apparent.

In thin section the first impression isthat the rock is an "Archaean" phyllonite, and with the absence of pebbles (or heavy mineral lamination in 157-1025A) the rocks may well have been mapped as "Archaean", especially as the outcrop is a small outlier in the Houghton Complex near the main contact. The "sandy" matrix consists of angular fragments of quartz, microcline and some albite in a granular matrix of quartz and sericite. Many of the grains are elongate parallel to the foliation and show tails of sericite and polygonal quartz. Thus the clastic nature of the matrix is not obvious except by analogy with the "pebbles". That the pebbles are not of tectonic origin is readily apparent from their variety of composition. The most common type of pebble is a quartz-microcline rock. A few fragments are large grains of microcline but most are polycrystalline and contain quartz and microcline. Graphic texture occurs in one pebble, and granulitic texture in another. One fragment of plagioclase rock was observed. It is of note that the plagioclase in this pebble is crowded with sericite. This alteration is probably pre-depositional and not induced at the time of the formation of the schistosity, as many grains of fresh plagioclase are observed in the rock. In the "Archaean" feldspar gneisses such small scale variations in amount of retrogression are not common.

Many of the microcline grains are film perthites and some contain rounded inclusions of quartz, a common feature in some of the feldspar gneisses. Other microclines show crinkled twins typical of the phyllonites but this could result from the deformation of the conglomerate. Other pebbles consist of quartz-microcline and poikiloblastic muscovite, typical of the rocks around Houghton, west of the feldspar gneisses. Other fragments appear to be typical phyllonites. They are elongate but have sharp contacts with the sandy matrix and are not thought to be produced during the formation of the schistosity but are derived from the Houghton Complex.

The adjacent finer grained rocks are arkosic in composition consisting of plagioclase, microcline and quartz. The rocks have a foliation marked by zones very rich in fine-grained micaceous material. The layering is well defined by heavy minerals (haematite and minor amounts of zircon).

The nearby outcrop in Houghton itself (157-1015) is much less schistose and the clastic nature of the microcline and quartz is readily recognizable. The grains are angular to sub-rounded, having been modified somewhat by the formation of sericite at the margin. Many of the microclines have outlines controlled by the cleavage faces. A heavy mineral facies of this outcrop contains, in addition to microcline and quartz, much haematite in rounded grains. Clastic tourmaline and zircon are present in accessory amounts and a new generation of fine grained euhedral tourmaline is also present as an accessory replacing the felsic minerals.

The outcrops just north of the River Torrens (157-980) show a different mineralogy reflecting the different nature of the rocks of the Houghton Complex in this region. The adjacent Houghton Complex consists of micaceous retrograde quartz-feldspar gneisses. The conglomerate contains much more quartz than the Houghton occurrences, the grains showing pronounced undulatory extinction. Feldspar is fairly abundant (both microcline and albite). Textural features which can be matched in the feldspar gneisses are rare.

In the Torrens Gorge east of Castambul is another outcrop of basal beds in contact with the older sequence (157-324). Although the contact is sheared the beds above are thought to be very low in the sequence if not the true basal beds. Rounded grains of quartz and microcline are dominant with opaques abundant in certain layers. Plagioclase is a common accessory. Sericite and quartz form the groundmass. The rock is relatively undeformed although some recrystallized quartz is present.

On the eastern side of the Houghton Complex, the basal beds are somewhat better size-sorted and have a high proportion of quartz. Specimen 157-1176 contains about 60% quartz with minor rounded microcline and opaques. The original clastic nature of the quartz is preserved as rounded grains, and optically continuous quartz outgrowths form a completely interlocking mosaic with adjacent quartz grains. The strain bands which occur in all the detrital grains are continuous into the outgrowths. Some outgrowths on microcline are observed but the interstices between microcline grains are mostly filled with quartz. Of three large pebbles sectioned from this locality two appeared very similar to the nearby feldspar gneisses, showing the confused chequer nature of the plagioclase twins characteristic of the gneisses in the Kersbrook region. The third pebble is a granular quartz-microcline rock which could not be matched with immediately local rocks.

Other quartzites close to the Houghton Complex - Torrens Group contact have not been included in this discussion as the nature of the structure suggests that they are faulted against the Houghton Complex. As such, they are discussed later with other quartzites. Spry (1951) also notes some Torrens Group outliers in the region around Paracombe in the centre of the southern part of the complex (Spry, 1951, Fig. 4, p. 126). These occurrences are not consistent with Spry's ideas of the structure and from his schematic cross section (his Fig. 3) the unconformity would be estimated to be more than a thousand feet above these localities. These are in fact conglomerates related to the Tertiary peneplanation and have no structures attributable to deformation.

Features of these basal beds of the Torrens Group have been described in some detail because it is thought that they offer some evidence concerning the relative ages of events in the Houghton Complex and Torrens Group. The evidence that many of the retrograde features of the Houghton Complex rocks are found in the pebbles of the Torrens Group conglomerates suggests that many of the textural features observed in the Houghton Complex were produced before the deposition of the Torrens Group. However, as indicated by the structures in the basal beds, fabrics similar to the phyllonites in the Houghton Complex have been produced during the deformation of the Torrens Group. Effects of these two phases of low grade metamorphism are presumably present in the Houghton Complex but it seems impossible to distinguish them on fabric evidence.

THE TORRENS GROUP SEQUENCE ABOVE THE BASAL CONGLOMERATE.

The remainder of the sequence will be discussed according to rock type rather than order of deposition. Stratigraphic relations are only briefly mentioned and have not been studied in detail.

Quartzites.

Quartzites occur as prominent horizons in various parts of the sequence; the most important are the Stonyfell Quartzite and those associated with the Montacute Dolomite.

Mineralogically and texturally all the quartzites are very similar and they also show similar deformation features. They are characterized by the assemblage quartz-microcline-albite with some interstitial sericite and accessories zircon and tourmaline. In most cases they would be classified as arkoses or feldspathic quartzites (if they were unmetamorphosed) with the exception of certain of the coarser-grained rocks which consist predominantly of quartz. Plagioclase is less common in the quartzites higher in the sequence. The quartzites close to the dolomite horizons commonly contain interstitial carbonate and with increasing carbonate grade into "sandy" dolomites.

In some quartzites outgrowths on microcline are observed (Fig. 44) and the original clastic nature of the grains is well preserved. Quartz less commonly shows the original clastic grains but this may be largely due to the difficulty

Fig. 44. Square outgrowth of microcline on a clastic microcline grain.
Quartzite. Spec. 157-584.

Fig. 45. Undulatory extinction in large quartz grains, and intergranular polygonal quartz.
Quartzite.Spec. 157-34.

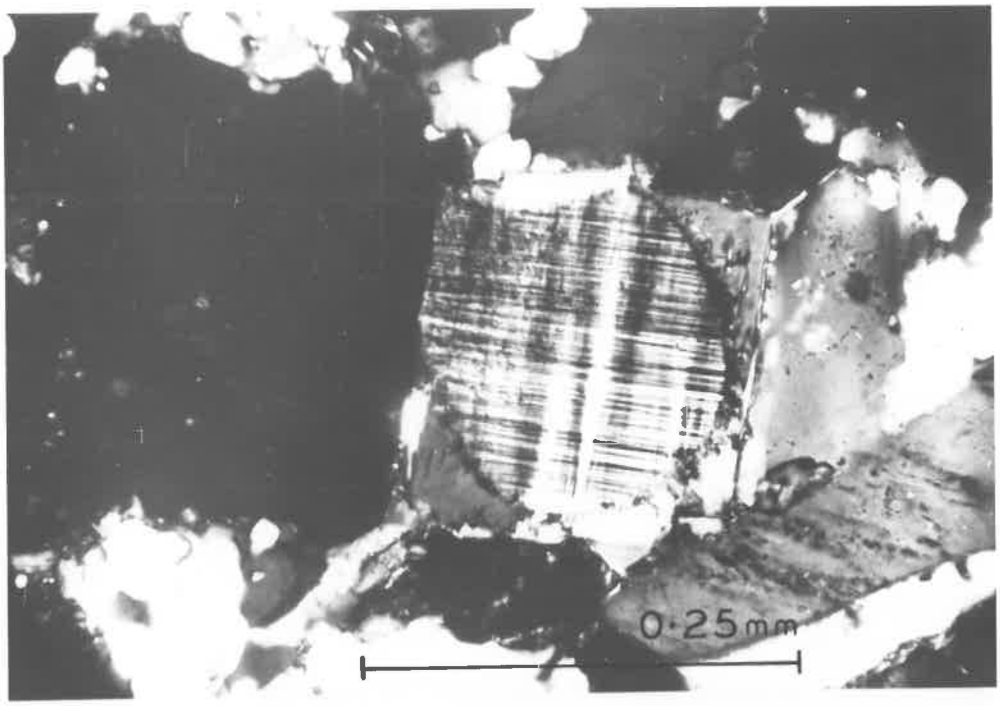


FIG. 44

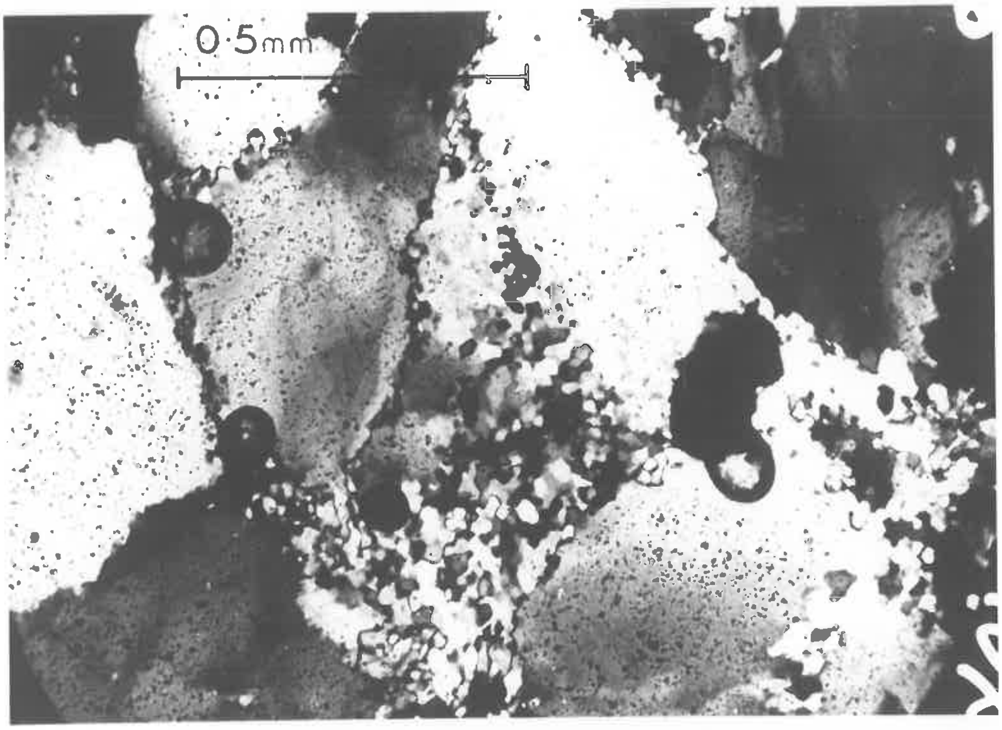


FIG. 45

of observing the boundaries between the clastic grain and the outgrowth where this boundary is not marked by inclusions.

Evidence of deformation is mostly limited to undulose extinction and extinction bands in quartz although patches of intergranular polygonal aggregates are to be found in many specimens (Fig. 45).

In the southeast and east parts of the area some of the quartzites are completely recrystallized and a quartzite near Mt. Gould appeared to be a blasto-mylonite (157-1170). The rock is finely layered, the layers being lensoid and approximately $\frac{1}{2}$ - 2 mm. thick. In thin section elongate relics of coarser quartz are observed, but most of the rock consists of a mosaic of irregularly shaped grains which show a high degree of preferred orientation when tested with a first order retardation plate. The orientation groups are in layers parallel to the foliation (Fig. 46). Unfortunately an oriented specimen could not be collected so the fabric cannot be related to the macroscopic geometry.

Dolomite marbles.

Thick layers of dolomite marble, which can be recognized as original dolomites occur in the Torrens Group near the base (Castambul Dolomite) and in the middle of the sequence (Montacute Dolomite). Dolomitic beds also occur above the Montacute but are thinly bedded and highly phyllitic.

The dolomites are fine-grained rocks, .01-.04 mm. with coarser patches and stringers with grains up to 0.5 mm.. White mica is a common accessory. Set in this matrix are

Fig. 46. Elongate relics of strained quartz in a polygonal quartz aggregate.
Blastomylonite Spec. 157-1170.

Fig. 47. Fine-grained quartz dolomite aggregate. Clastic quartz grain (medium grey in centre of picture) shows a quartz outgrowth which is embayed by carbonate grains.
Dolomite marble. Spec. 157-713A.

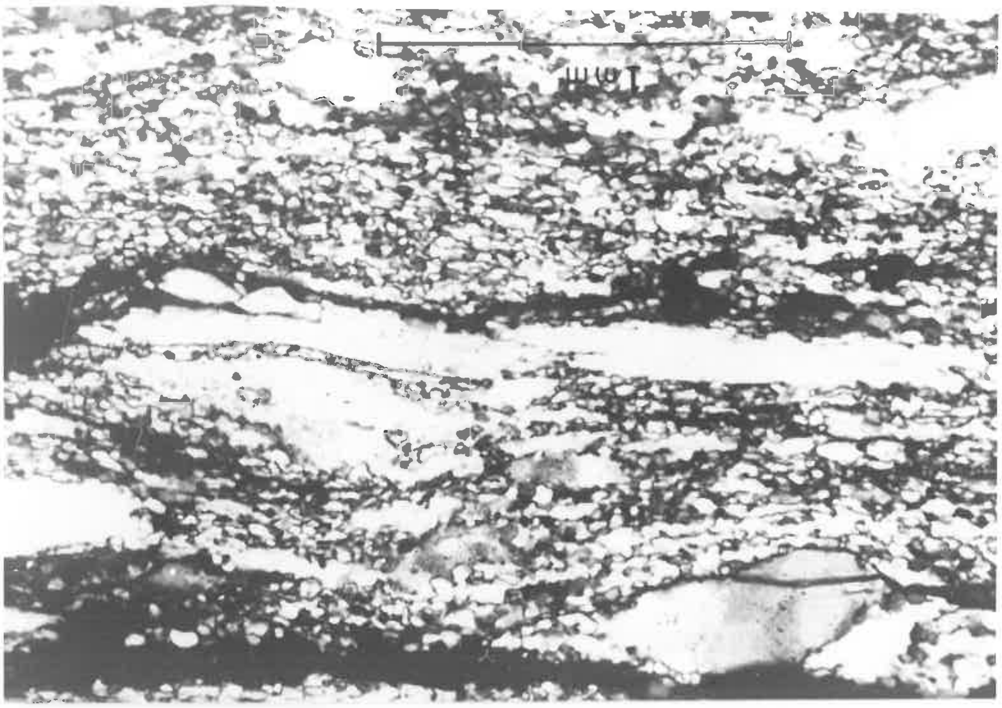


FIG. 46

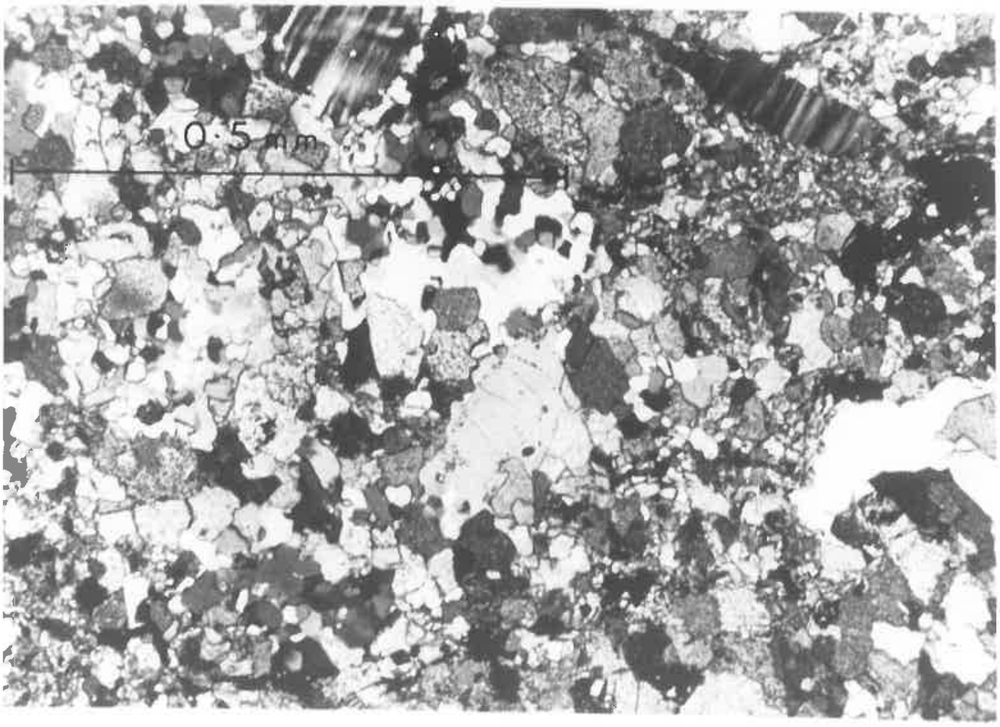


FIG. 47

sand-size grains of quartz and microcline .3-.8 mm. diameter. The content of this clastic material varies from less than 5% to layers which have over 90% clastic material. Some rocks contain rounded pebbles of magnesite.

The texture of the dolomites is granoblastic; however, the microcline grains and to a lesser extent the quartz grains still show rounded clastic outlines. Microcline commonly shows square outgrowths, as in the quartzites; such outgrowths are considered by Baskin (1956) to be authigenic. Quartz does not commonly show its clastic nature and occurs as lenses; usually each lens is a single grain, and the long axes of these lenses tend to be parallel and are oriented parallel to the local foliation. It is possible that the quartz has been recrystallized or has had its shape modified by some other metamorphic process. Certainly solution or replacement of the quartz and microcline has occurred, embayed contacts being common and in many cases microcline has been replaced by carbonate preferentially along cleavage planes. In some rocks (e.g. 157-713A) the quartz has recrystallized to a fine-grained aggregate (Fig. 47) and carbonate grains are found in the mosaic.

All of the carbonate grains are considered to be re-crystallized and show no evidence of the original nature of the grains. However, Forbes (1960) in a study of the Montacute dolomite in localities where it is unmetamorphosed, reports rounded carbonate grains, some with an angular quartz grain at their centres. The abundance of mud cracks, ripple marks and slump structures in these areas suggests that the dolomites and magnesites were deposited in shallow water and some re-working has produced the magnesite conglomerates. This mode of origin is consistent with that deduced from the Torrens Gorge occurrences.

Phyllites.

Phyllites form the bulk of the rocks in the Torrens Group. They consist of quartz, sericite, chlorite and minor feldspar but many of them have considerable carbonate. In the lower part of the sequence abundant lenses of fine-grained quartzite occur in the phyllites; in the phyllites between the Montacute Dolomite and the Stonyfell Quartzite, dolomite is abundant and most of the rocks can be described as phyllitic dolomites. In these latter rocks the mica is somewhat coarser grained and it is in these rocks that the best examples of crenulation cleavage are observed (see Chap. VII).

Distribution of the Torrens Group.

The Torrens Group rocks can be traced around the southern end of the Houghton Complex and certain lithological changes are noteworthy. It is obvious that the sequence on the east of the Houghton Complex cannot be matched with the sequence on the west. This fact has been long recognized and is embodied in the Adelaide 1 mile Geological sheet where two stratigraphic columns are given, one for each side of the "Archaean" core.

The dolomite horizons are the only marker formations which can be followed around the southwest end of the Houghton

Complex. The Castambul dolomite has a very restricted occurrence, not being found north of the Anstey Hill-Paracombe Road (possibly faulted out) and being faulted out against the Houghton complex just south of the River Torrens. No changes could be observed in it over the extent of its outcrop. The Montacute Dolomite is more widespread, extending from the Tea Tree Gully-Inglewood Road to east of Castambul. To the north, phyllites occur in the position where the dolomite should be seen and the relatiohships between these two rock types is not known, they may represent a facies change or a fault contact. To the south the dolomites can be traced south of Castambul and around the southern end of the Houghton complex. Individual bands are richer in quartz grains to the east and most of them grade into quartzites before they appear to be faulted against the Houghton Complex.

Outcrops to the east of the complex are very poor but the absence of carbonate beds is a striking feature of the River Torrens sections. However, many of the phyllites are very carbonate-rich and may represent the equivalents of the dolomites to the west.

It seems reasonable that this area now occupied by the Houghton Complex acted as an axis of sedimentation during the deposition of the Upper Precambrian sediments.

This axis of sedimentation is coincident with the axial surface of the regional anticlinorium. This may be a coincidence but suggests a fundamental structural control which operated over a long period of geological time.

CHAPTER VI.

NOMENCLATURE OF STRUCTURES IN THE TORRENS GROUP ROCKS.

The structures observed in the Torrens group are:

1. Folds on all scales with a slaty cleavage parallel to the axial surfaces of individual folds.
2. Mesoscopic folds in certain horizons with a crenulation cleavage as axial surface.
3. Rare chevron folds.

Rare small-scale structures are observed which cannot be related to the other structures. Structural analysis suggests that there may be macroscopic structures present which are not recognizable on the scale of a single sub-area.

In accordance with established practice, symbols have been assigned to the various structure surfaces and lineations observed.

<u>Surfaces.</u>

S Original bedding

S_1 Transposed bedding

S_2 Slaty cleavage

S_3 Crenulation cleavage

S_4 Axial surface of kink folds

<u>Fold axes</u> are called B_2, B_3, B_4; for example B_2 indicates that the fold has S_2 as axial surface and is a fold in S.

The scale of folds[1] is referred to as:

1. Microscopic - scale of a thin section.
2. Mesoscopic - scale of a small outcrop or hand specimen.
3. Macroscopic - scale larger than a single outcrop.

All figures of folds are transverse profiles with W on the left and the scale horizontal.

1. The scale of folds used follows that of Weiss (1957).

CHAPTER VII.

B_2 FOLDS. FOLDS WITH SLATY CLEAVAGE PARALLEL TO THE AXIAL SURFACE.

Folds with S_2 as axial surface occur in all regions where the Torrens Group has been studied. The folds occur on all scales from microscopic to over 1 km. in wavelength and ½ km. amplitude, and in all rock types from phyllite to massive quartzite. Mesoscopic folds are best studied on the west side of the Houghton Complex where outcrops are better and road cuts more numerous. Much of the ground east of the Complex is devoid of all outcrop except quartzite, and mesoscopic folds are rarely seen. However the best exposed area of macroscopic folds is in this region and the profile of this series of folds is quite consistent with the mesoscopic style in the west.

STYLE OF MESOSCOPIC B_2 FOLDS.

The style of folding varies with the gross lithology and with the small scale variations in lithology, and characteristic examples of each type are described.

a) Interlayered quartzites and phyllites.

Folds are markedly asymmetric in profile, and if any differences in the lengths of limbs is present the east limb[1] is always longer than the west limb, in some cases markedly so. Commonly the boundaries between the quartzite and phyllite are

1. Note, always referring to the antiforms.

quite sharp and the folds are outlined as smooth curves of the interfaces between the two rock types. Each individual quartzite layer may be considered separately and is seen to have the characteristics of concentric folding, namely decreasing radius of curvature downwards in the antiform and essentially the same thickness measured perpendicular to S at any position on the fold. In some cases the curvature of the top of the antiform in a quartzite is reflected in the curvature of the overlying quartzite but in other cases there is a distinct structural break between the two quartzites, the smooth top of one antiform in contact with the sharp cuspate base of the antiform in the overlying quartzite (Fig. 48a). These differences reflect differences in competence of various of the interlayered phyllites, structural breaks occurring only in places where the differences in properties between the phyllite and quartzite are large; where the differences are not as great the fold continues across the phyllite into the overlying quartzite. Small scale laminations within the quartzites commonly show smooth fold outlines parallel to the larger scale layering but in other instances a small scale puckering of these laminations is observed in the crestal regions of the folds.

In many cases the thickness of individual beds varies around the fold and a thickening of the crestal regions is observed (Fig. 48c). Ramsey (1962) has interpreted such structures as modifications of concentric folding due to

Fig. 48. Concentric folds in Torrens Group quartzites.

48 A and B Lower Hermitage (Grid ref. 764 996)

48C River Torrens (Grid ref. 748 927).

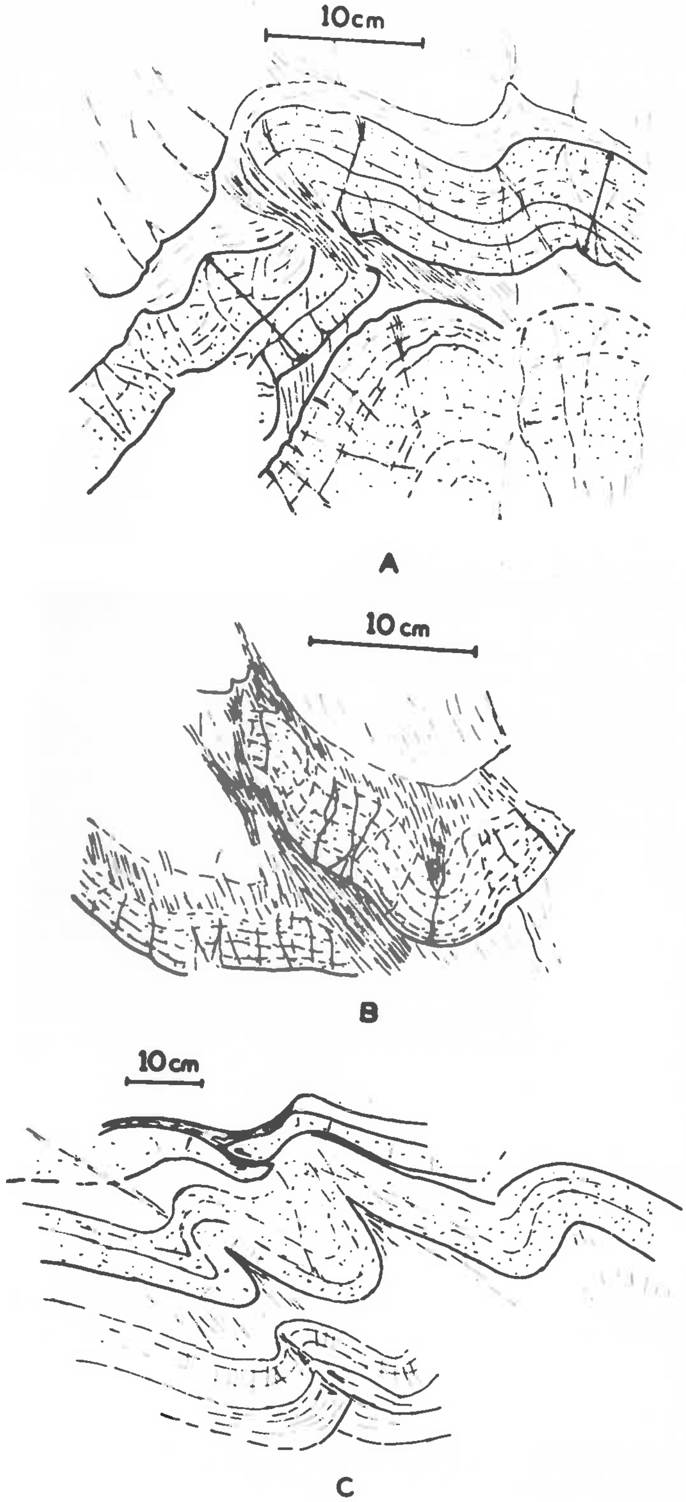


FIG. 48

flattening.

A common feature of folds in these rocks is the presence of small thrust zones subparallel to the axial surface and breaking through the west limbs. They are particularly common in the closely bedded rocks and emphasize the marked asymmetry of the folds (Fig. 48).

b) Minor quartzites in phyllitic sequences.

Whereas in the previous group the amount of appression of a particular fold is controlled by the confining nature of the adjacent quartzites, no such constraint is noticeable in the deformation of single thin quartzites in a phyllite sequence. Even in the neighbourhood of thicker quartzitic sequences, great differences in the degree of appression are observed. It is not uncommon to see a quartzite showing fairly broad concentric folding with the extension of the fold into the phyllites becoming much more appressed (Fig. 49A). Completely appressed folds of quartzite in phyllite are not uncommon (Fig. 49B) and rootless folds are extremely common (Fig. 50A) Extreme variations in the thicknesses of the limbs can be seen, the west limb being more commonly thinner than the east (Fig. 50B).

c) Homogeneous sequences.

To this category are referred poorly laminated phyllites and some fine-grained quartzites with fine laminations. In both cases the layering is of less importance than in the previous examples and appears to have played a more passive

Fig. 49. Folded quartzites in phyllites.

A. Appressed syncline core beneath open fold in quartzite.
Lower Hermitage (Grid ref. 764-996).

B. "Isoclinal" fold of quartzite in phyllite.
Little Para River (Grid ref. 774 031).

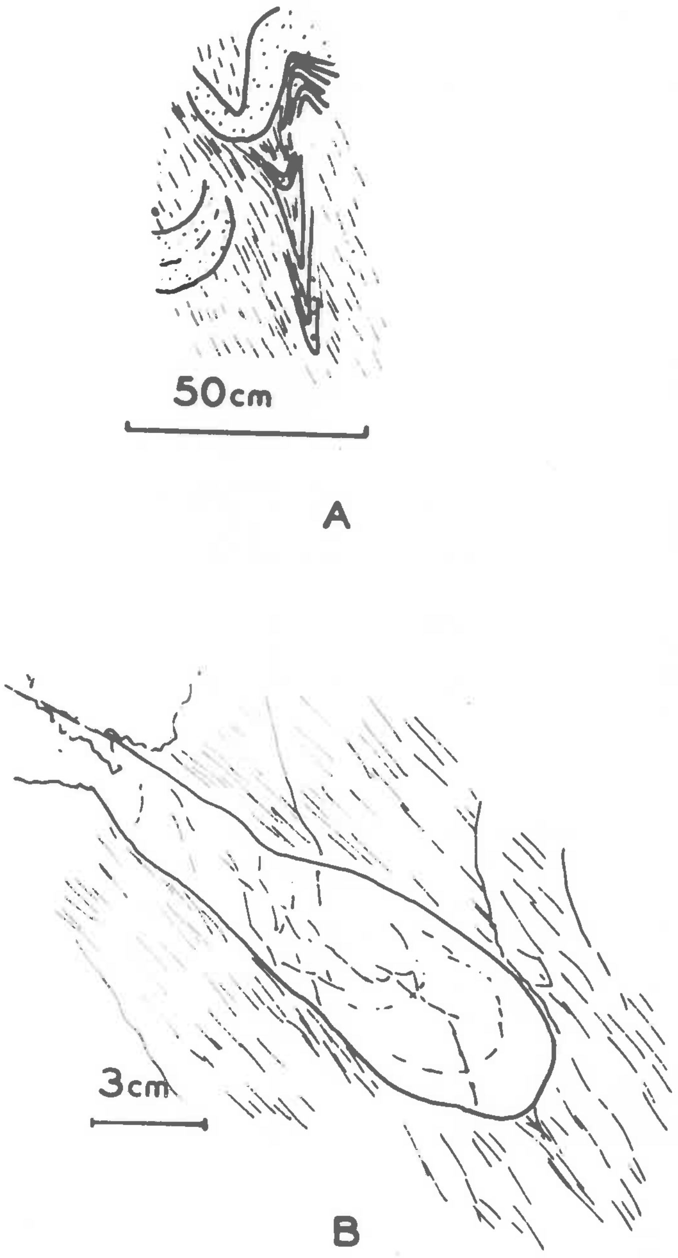


FIG 49

Fig. 50. Folds in quartzites.

A. Rootless folds of quartzites in phyllites.

Lower Hermitage (Grid ref. 781 022).

B. Fold of quartzite in phyllite. One limb is considerably thinned.

River Torrens (Grid ref. 746 926).

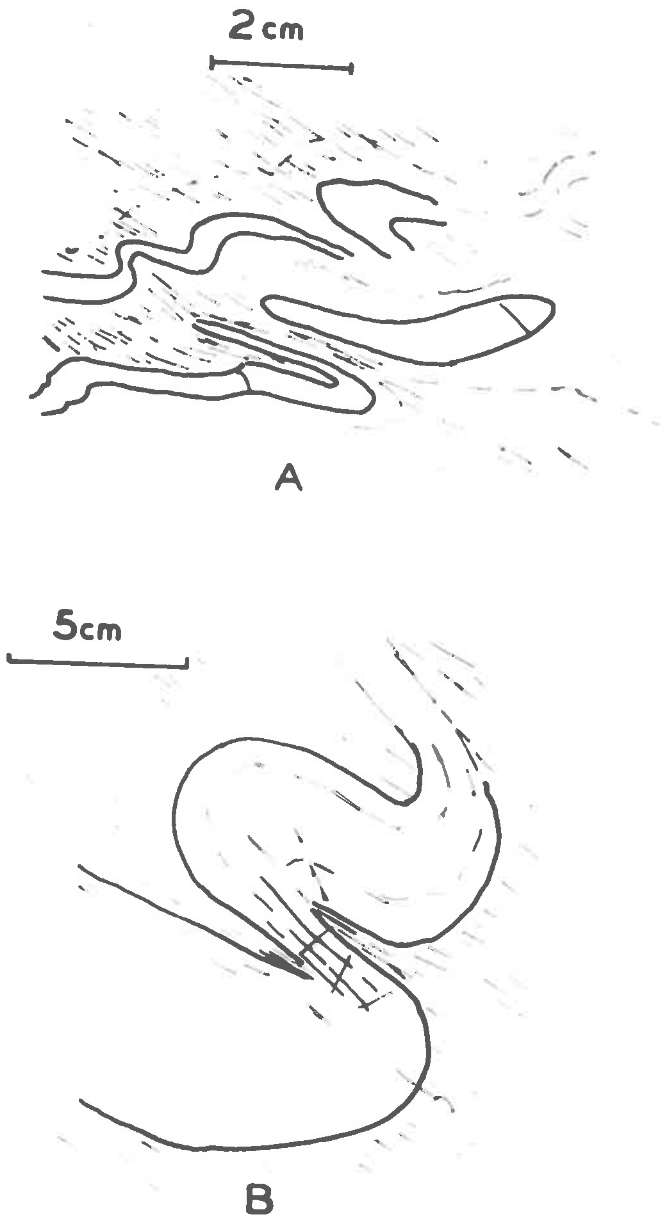


FIG 50

role in the deformation. In many cases the style is intermediate between concentric and similar, and is less asymmetric than the previous examples (Fig. 51). In phyllites strictly similar folds are observed, generally however in thin section or on polished blocks. Where differences in lithology are great enough to be easily visible in hand specimen the style tends to be similar to that of the previous group.

STYLE OF MACROSCOPIC B_2 FOLDS.

Owing to difficulties due to lack of outcrop, faulting and low plunge, the style of macroscopic B_2 folds west of the Houghton Complex is difficult to determine. In the River Torrens the anticlines are characterized by long east limbs with low dips and short west limbs with steep dips. The associated synclines are more appressed than the anticlines. In some cases faulting of the west limb is observed, notably in the folds in quartzite north of Crouch's Hill where a thrust fault cuts out much of the west limb of a large asymmetric anticline.

To the east of Cudlee Creek on the southeast side of the Houghton Complex a series of quartzites forms a group of hills around Mt. Bera. These have been heavily dissected by the River Torrens and good exposures of the quartzite are visible. The transverse profile of the folds formed in these quartzites (Fig. 52) shows many similarities to the style of the mesoscopic folds. In the area where the quartzites are

Fig. 51. Folds in laminated quartzite (oblique profile).

River Torrens. (Grid ref. 747 917).

FIG 51

Fig. 52. Transverse profile of folded quartzites.

Mt. Bera.

Grid ref. A. 842 970

Grid ref. A′ 852 953

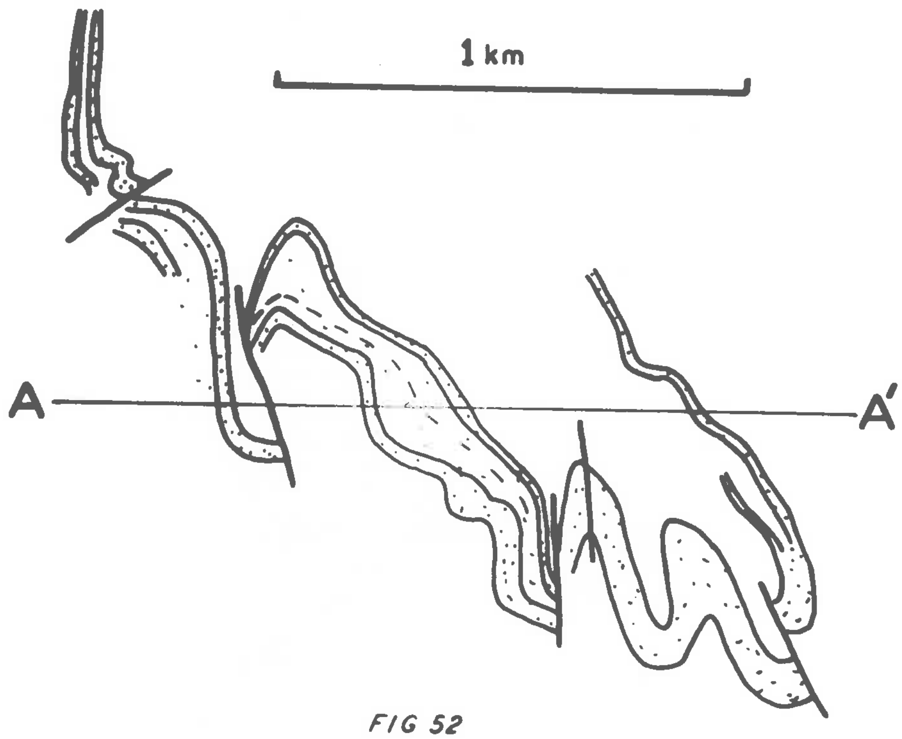


FIG 52

thickest long east limbs dominate the structure with shorter faulted west limbs. In the more highly folded parts of the quartzites the folds are more symmetrical and faulting is less common. The right-hand fault at the top left of the profile may be a conjugate fault, such faults also being observed in microfolds.

In no case was there any evidence of macrofold styles conflicting with the fold styles described above and it is concluded that the macrofolds, which can be mapped in an area of one or two kilometres square, are genetically related to the mesoscopic folds and do not represent an earlier phase of folding. Macroscopic analysis supports this conclusion but there remains the possibility that very large folds existed before this main period of deformation (See Chapter IX).

STYLE OF MICROSCOPIC B_2 FOLDS.

In their broader features the style of folding on the scale of a thin section (Fig. 53) shows close similarities to the style of mesoscopic folds but a better picture of the deformation is obtained from a study of microscopic folds as no problems of "poor outcrop" and "doubtful correlations" are encountered.

In the less micaceous layers it is clearly seen that the fold surface is controlled by sharp changes in lithology and that the outline of the folds at these junctions is concentric in pattern. The adjacent micaceous bands may show no correspondence of structure whatsoever or just reflect the gross

Fig. 53. Style of microscopic folds in interlayered phyllites and quartzites. Spec. 157-637.

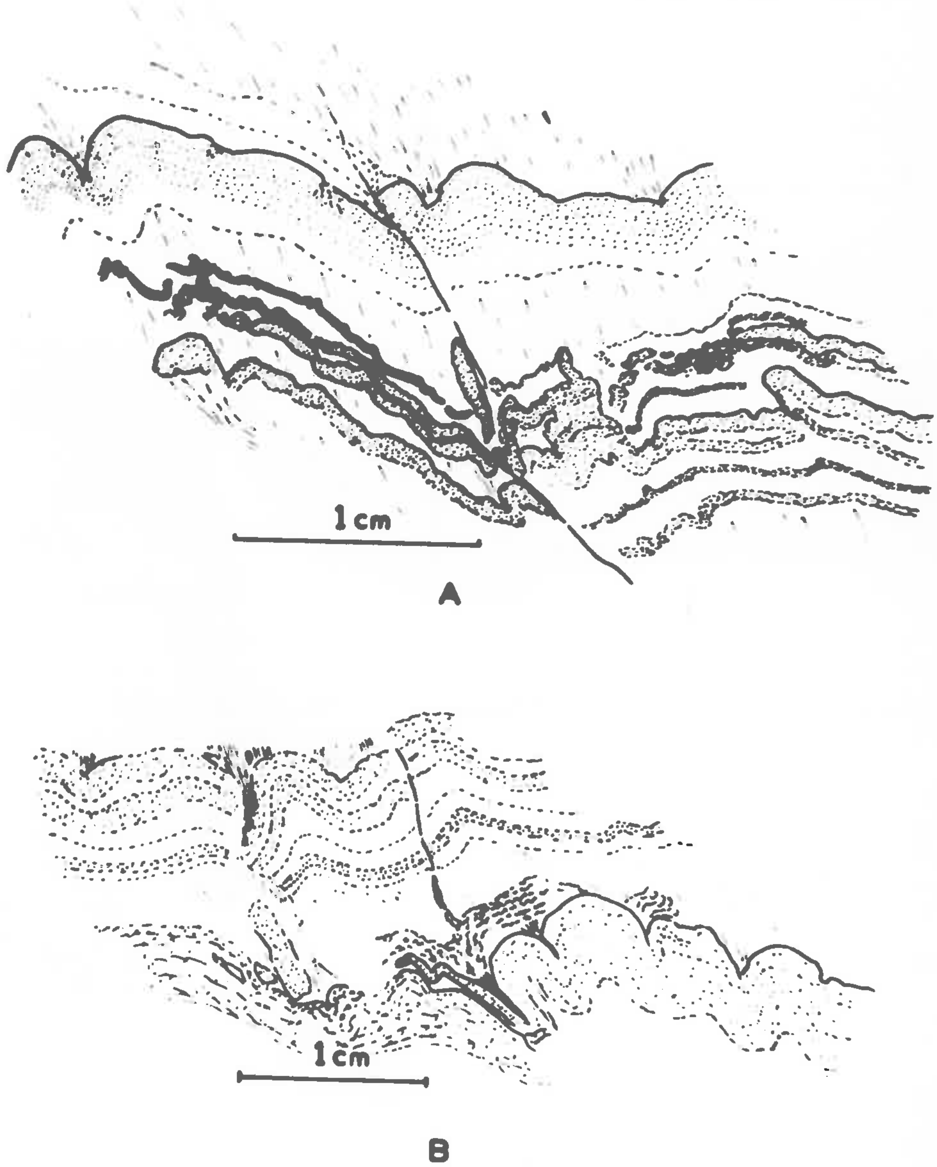


FIG. 53

features of the folding (Fig. 53B). If, as is commonly the case, one of the contacts between "silt" and phyllite is more pronounced than the other, then the less pronounced contact will have a smoother, less indented profile than the other contact. In fact the folds may die out without any detachment surface, the intensity of folding becoming greater towards the sharp contact (Fig. 53A). In contrast a strong decollement is commonly observed at the sharp contact (Fig. 53B centre).

In thinner quartzitic layers (less than 1 mm.) folding is less regular and small scale faulting along cleavage planes commonly results in the layers being quite discontinuous. If the breaking is regular, structures similar to cleavage boudins result. In some cases "intrusion" of the broken material along the cleavage results in diapiric structures (Fig. 54).

The faulting observed in mesoscopic structures is also observed in thin section. As in mesoscopic folds the west limb is faulted although some conjugate right-hand faults are observed (Fig. 54). In some cases the faults are clean breaks although more commonly the west limb is extremely attenuated (Fig. 55). This observation is in accord with the mesoscopic evidence (e.g., Fig. 48 A), and in most cases the apparent shearing is parallel to the local cleavage (Fig. 55). A further consequence of the concentration of deformation in zones parallel to the cleavage is that the boundaries between different lithologies vary in sharpness in different parts of

Fig. 54. Diapiric structure of quartzite in phyllite. Drawn from photomicrograph of Spec. 157-637. Part of field of Fig. 53A.

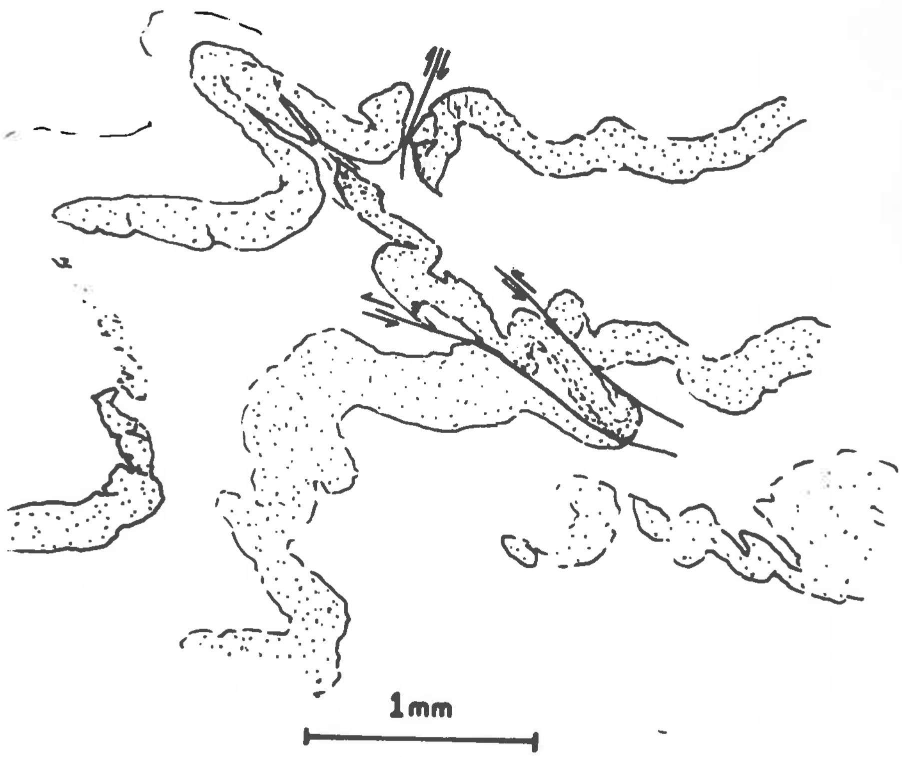


FIG 54

the folds. The east limbs are more nearly parallel to the cleavage than the west limbs, and may have sharper boundaries than the west limbs, which are in fact often only extended crestal regions. Deformation may have been concentrated in planes parallel to the cleavage resulting in differential slip parallel to the east limbs. In the west limb, the cleavage is at a high angle to the layering , there is no comparable slip parallel to the layering and the boundaries maintain their gradational character (Fig. 56). These observations suggest that the folding has taken place by a combination of rotation of west limbs and "slip" parallel to the east limbs. Fig. 57 shows this rotation in more detail. The east limb is considerably thinned compared with the crestal region and is parallel to the cleavage. The cleavage is somewhat rotated in the crestal region, the sense of rotation being anticlockwise and consistent with the folding (A certain amount of grain enlargement appears to have taken place in the core of this antiform although this phenomenon is not common).

Where deformation is concentrated into a narrow domain parallel to the cleavage, shearing through the metasilts is observed (e.g., Figs 48A and 55). The quartz in these layers is notably reduced in grain size and tends to be excluded from the deformation zone; the amount of micaceous material is increased and the rock may look more like the surrounding phyllites (Fig. 48A).

A phase of post-kinematic mineralization is seen as the

Fig. 55. Sheared antiform in quartzite. The shear zone is parallel to the local foliation.
Photomicrograph of part of field of Fig. 53A.

Fig. 56. Folds in quartzite. The limb parallel to the cleavage has sharp boundaries. The other limb has more diffuse boundaries.
Spec. 157-637D.

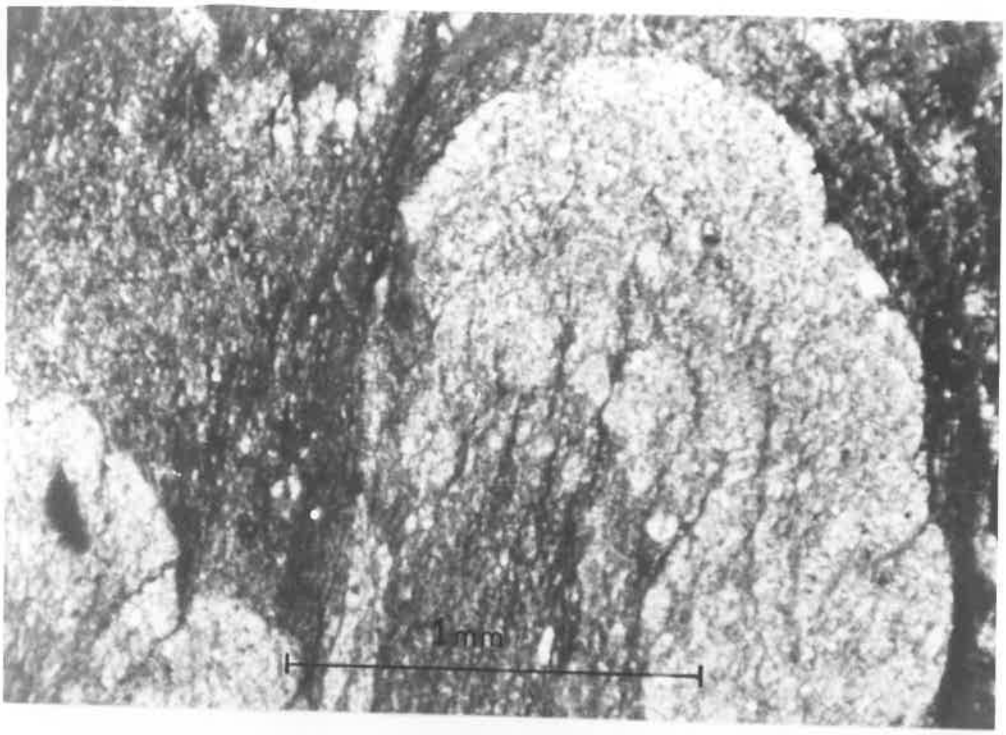


FIG. 55

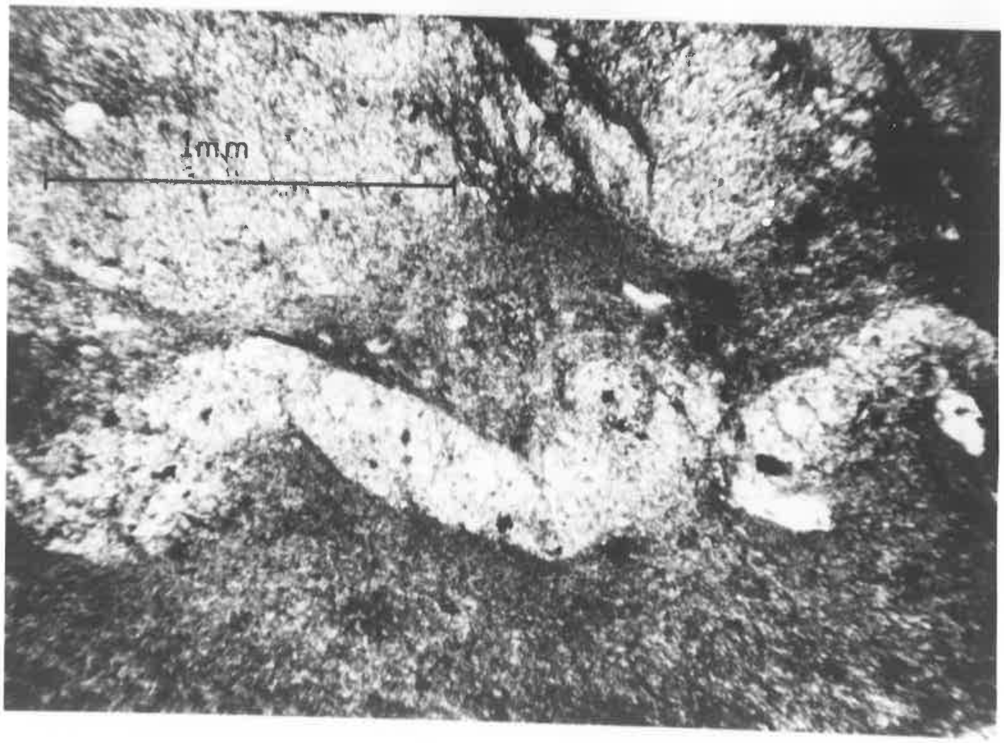


FIG. 56

Fig. 57   Small fold in phyllite. Quartz is notably elongate between foliation planes.
Spec. 157-1242.

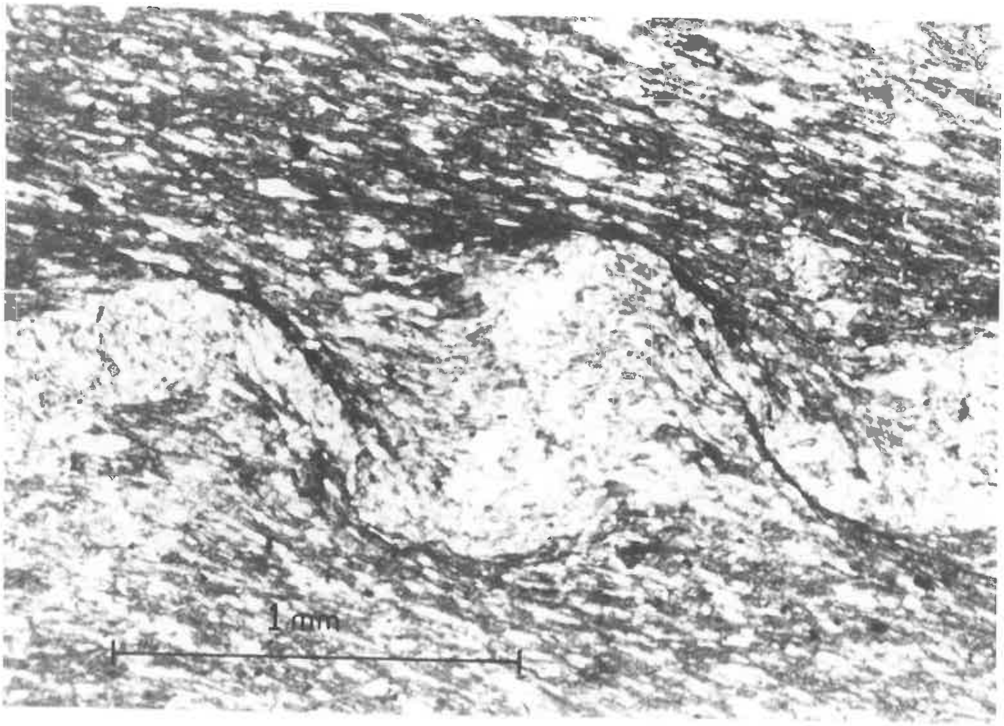


FIG. 57

enlargement of dolomite grains at the expense of quartz and mica and in the appearance of rare cross-cutting muscovite and green biotite.

Phyllites with essentially no metasilt layers are typically featureless and show a cleavage as the only important structural element. In some of the specimens faint traces of layering are visible and show typical similar-fold styles with much of the layering shredded out parallel to the cleavage (Fig. 58).

Cleavage.

Two types of cleavage are associated with this phase of deformation, a slaty cleavage in phyllites and a fracture cleavage in quartzites. The fracture cleavage has the appearance of fairly close but irregularly spaced joints which in some cases grade into slaty cleavage where they pass into adjacent phyllites. The attitude of the fracture cleavage is more variable than that of the slaty cleavage. Commonly the slaty cleavage is subparallel to the axial surface of the folds (if not, it is subparallel to the east limbs of appressed folds) whereas the fracture cleavage is always at a higher angle to the layering than the slaty cleavage (Fig. 59) and fans around the crests of folds in a manner compatible with this (Fig. 60). Where the contacts between different lithologies are gradational, smoother transitions in attitude are observed.

In thin section the cleavage is very irregular in development and varies with the quantity of micaceous material present.

Fig. 58. Similar folds in phyllite. The phyllitic groundmass is rich in mica and carbonate.
Specimen 157-569. Drawing of thin section.

FIG. 58

Fig. 59. Closely spaced joints (fracture cleavage) in quartzite. Slaty cleavage in phyllite. The fracture cleavage is at a higher angle to the bedding than the slaty cleavage. River Torrens. (Grid ref. 784 921).

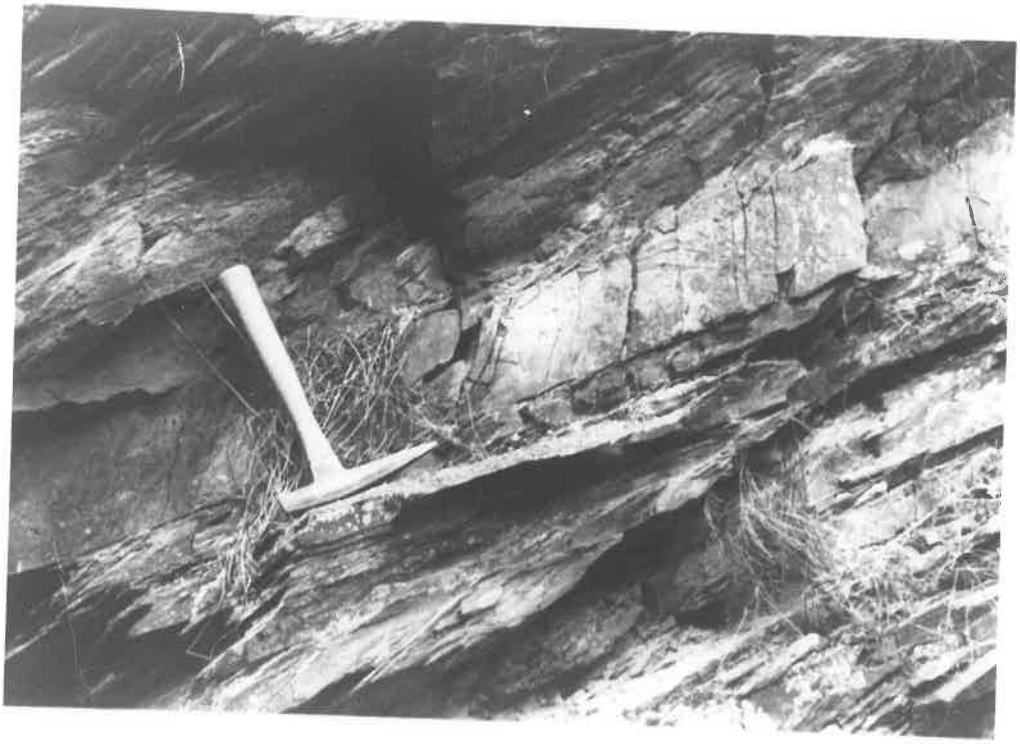


FIG. 59

Fig. 60.   Fracture cleavage in folded quartzite.
River Torrens. (Grid ref. 746 927).

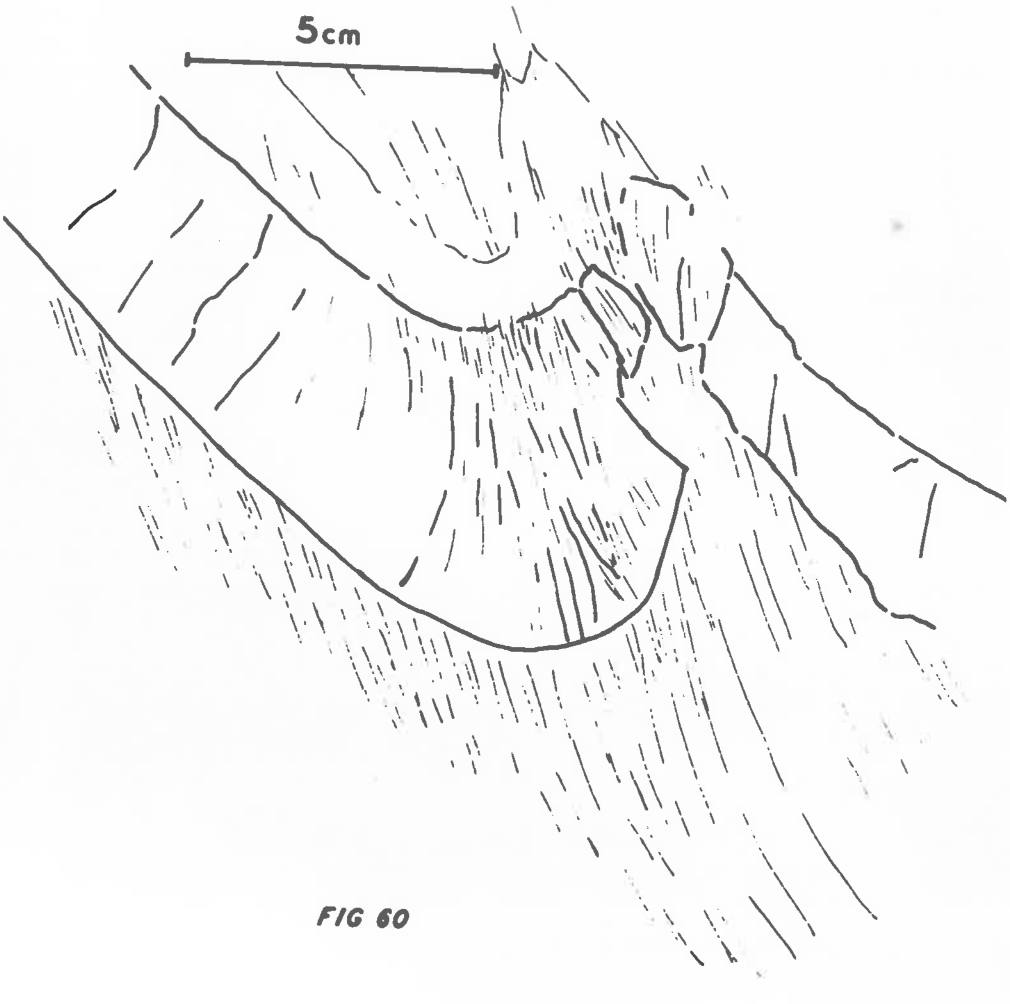


FIG 60

In the metasiltstones the fine-grained micas commonly show a planar preferred orientation which is less apparent in the more quartz-rich rocks. In the micaceous rocks not only is there a stronger preferred orientation of the micas but the rocks tend to be laminated with the concentration of micas in certain zones (e.g., Fig. 57). The quartz grains are commonly elongate and occur as eyes in the foliation. Elongate quartz grains are not observed in the quartz-rich layers, suggesting that the elongation is not produced by flattening but rather by recrystallization, the sericites exerting a marked influence on the grain shape of the growing quartz crystals (Voll, 1960). Commonly the cleavage is much better developed in the region of the cores of folds (Fig. 61) and the grain size of the quartz and mica is noticeably finer in these regions. Whether this is accompanied by any removal of material from these zones, as is noted in the development of S_3, is not known but would be a possible mechanism for release of stress in such highly deformed regions.

MECHANISMS OF B_2 FOLDING.

The concepts of "flexural slip" or "slip" folding as simple mechanisms do not seem to apply in the deformation pattern deduced. Slip surfaces are only locally developed although it is possible that they may have been masked by later recrystallization. However, the model of S passive deformation on planes (or in domains) parallel to the cleavage is applicable to the deformation of the most abundant rock types in the Torrens group, that is phyllite; but S-active

Fig. 61. Synform in quartzitic layer in phyllite.

Spec. 157-637. Part of Fig. 53A.

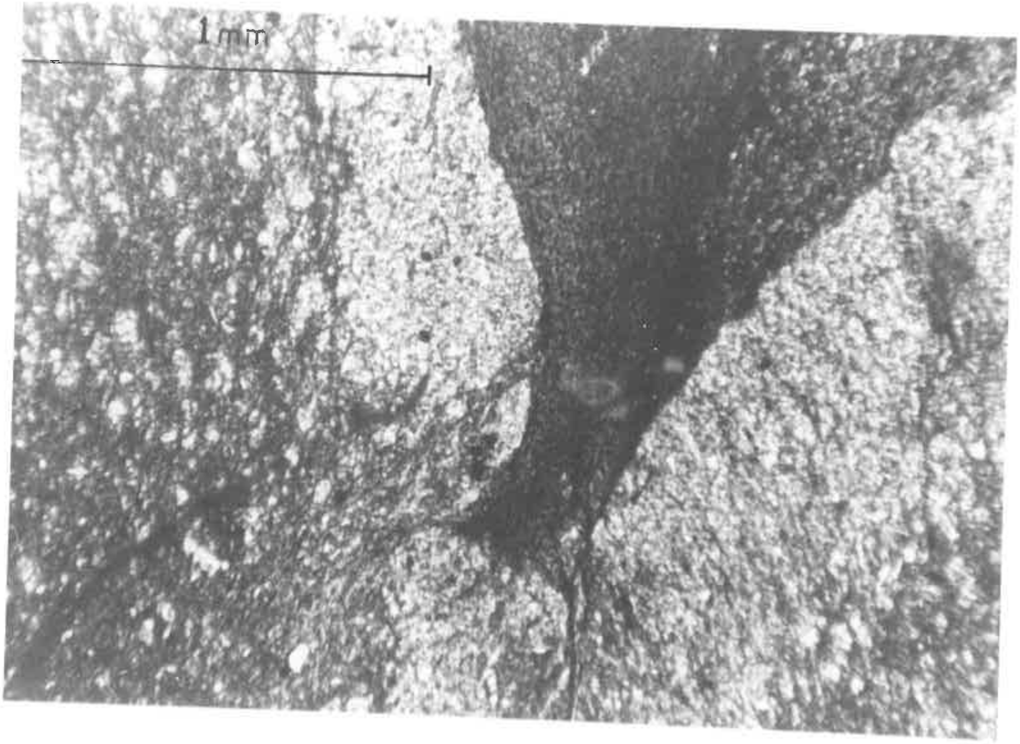


FIG 61

deformation has considerably modified the fold pattern where lithological variations have rendered the rocks markedly anisotropic. In fact it is the S-active component of the deformation which is most apparent on a mesoscopic scale, as folding is rarely visible unless the rock is markedly anisotropic.

The cleavage seems to have developed at an early stage in the deformation and in the quartz-rich layers has been passively rotated during the deformation. This period of deformation seems to be related to the main period of metamorphism of the Torrens group although much of the folding may have occurred after the formation of the metamorphic minerals.

CHAPTER VIII.

B_3 FOLDS. FOLDS WITH CRENULATION CLEAVAGE PARALLEL TO THE AXIAL SURFACE.

B_3 folds are seen to consistently overprint S_2 structures and are therefore later than the S_2 structures. Only rarely are B_3 structures seen to overprint B_2 folds as the B_2 and B_3 axes are roughly parallel, and B_2 folds are rare in the areas where S_3 is well developed. However folds and microcrenulations in the cleavage are extremely common and the B_3 folds appear to be related to these folds.

The distribution of B_3 folds is relatively restricted. They are best exposed in the road cuts in the Torrens Gorge just to the east of the main quartzites (Stonyfell Quartzite) near the mouth of the Gorge but are found as far north as Tea Tree Gully and Upper Hermitage. The occurrence of microcrenulations in S_2 is more widespread and they are found throughout the whole area west of the Houghton Complex. These crenulations are rare to the east; however this may partly be the result of differences in rock type and exposure in this area.

MESOSCOPIC STYLE OF B_3 FOLDS.

B_3 folds are best developed in the laminated dolomitic phyllites stratigraphically below the Stonyfell Quartzite in the Torrens Gorge. Examples are more rarely found in quartzitic phyllites provided that the quartzitic laminations are thin. The dolomitic phyllites can be traced to the north of the River Torrens and in the neighbourhood of Tea Tree Gully and Hermitage they are seen to be less dolomitic and more quartzitic. Associated with the decrease in dolomite, B_3 folds are less common but folds in S_2 are abundant, either as a crenulation cleavage or as a larger scale folding of S_2.

In contrast to the B_2 folds, B_3 folds have a pseudo-orthorhombic symmetry; the limbs are commonly symmetrical, in length as well as orientation and both limbs may be highly attenuated. A crenulation cleavage is strictly parallel to the axial surface of the folds (Fig.62A), although in some outcrops two cleavages are seen at low angles to one another and symmetrical to the axial surface of the mesoscopic fold. In no case could a "fanning" of the cleavage be proved for a single mesoscopic fold although in some instances the crestal regions show slight deviations from the planar attitude of the foliation (Fig. 62B).

The size of an individual fold rarely exceeds 1 m. in amplitude and no macroscopic structures could be interpreted as B_3 structures, although the macroscopic flexing of S_2 may be related to this phase of deformation. The folds in the thinly laminated phyllites have a similar style, the individual microfolds being remarkably persistent down the dip of the axial surface. However the interlayered thicker dolomites commonly show concentric style with only minor attenuation of the limbs (Fig. 62A). In some of the folds the laminations in

Fig. 62. B_3 folds in phyllite and dolomite.

A. A crenulation cleavage is parallel to the folds in the more massive dolomite beds. The dolomites show concentric shear planes.

B. The attitude of the crenulation cleavage is somewhat variable but is parallel to the axial surface in the apices of the folds.

River Torrens (Grid ref. 734 930).

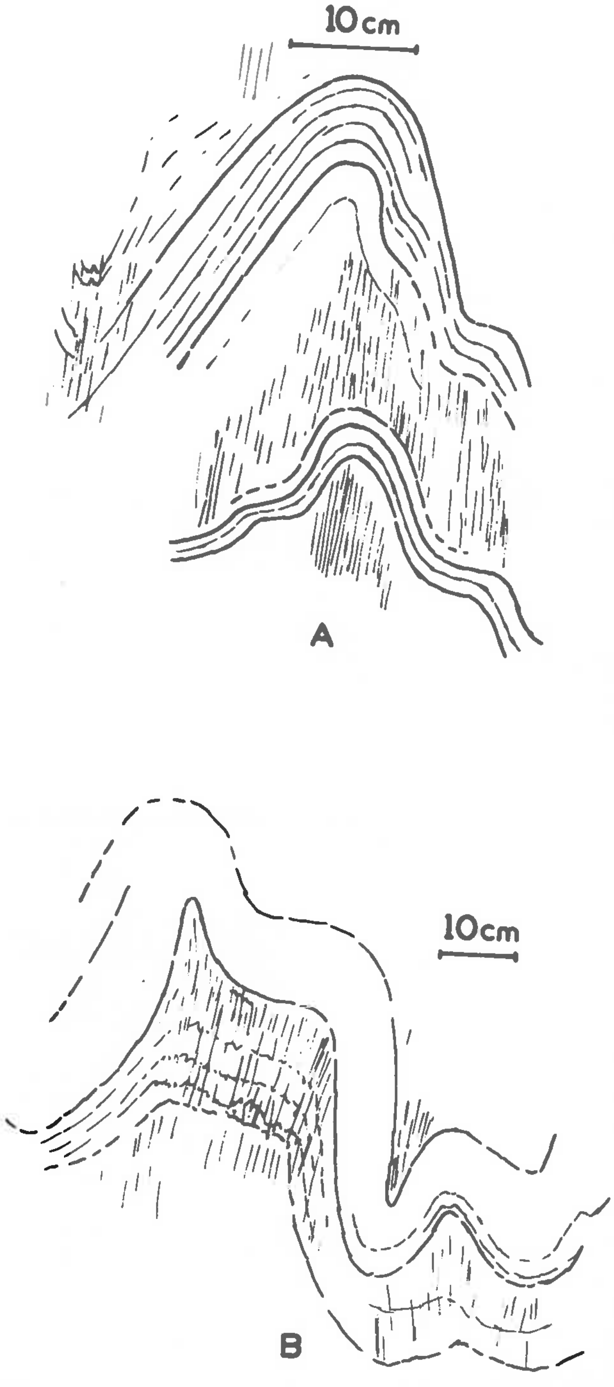


FIG. 62

the dolomites are remarkably persistent and sharp (Fig. 62A), an observation not in keeping with the general nature of layering in the dolomites, and it is thought that these planes may have been domains of intense deformation and that the folds may approximate to the model of concentric shear. Certainly no other mesoscopic folds in the region contain such convincing evidence of a flexural slip mechanism.

The cleavage associated with the B_3 folds is a crenulation cleavage[1] and appears as a series of closely spaced parallel planes which are commonly the steeper limbs of microfolds, or may be discrete planes showing small offsets of the quartz-rich layers. In some cases a layering is developed in the more phyllitic layers parallel to the cleavage. In many outcrops this new layering and the original layering can be seen in adjacent parts of the outcrop (Fig. 63).

Folds in S_2 are common in the poorly laminated phyllites. Apart from the microfolds the common styles are gentle waves in S_2 (Fig. 64) or kinks where the development of S_3 occurs only in limited zones (Fig. 65A). In these regions it is common to find that S_3 does not affect the more quartzitic layers and abuts against them (Fig. 65B).

1. The terminology of cleavages used in this thesis follows that of Knill (1960) except that crenulation cleavage is used instead of "strain slip cleavage" as the latter implies a genetic significance. Knill himself points this out but still uses strain slip cleavage in his suggested classification. Rickard (1961) formally proposed the use of crenulation cleavage to "designate cleavage planes whether micaceous layers or sharp breaks, which are separated by thin slices of rock containing a crenulated cross-lamination".

Fig. 63. Layering parallel to S_3 (vertical) Original layering visible crossing the top third of the picture.
River Torrens (Grid ref. 734 930).

Fig. 64. Folded $S_1 = S_2$ surfaces with crenulation cleavages parallel to the axial surface of the folds.
Upper Hermitage (Grid ref. 748 985).

FIG. 63

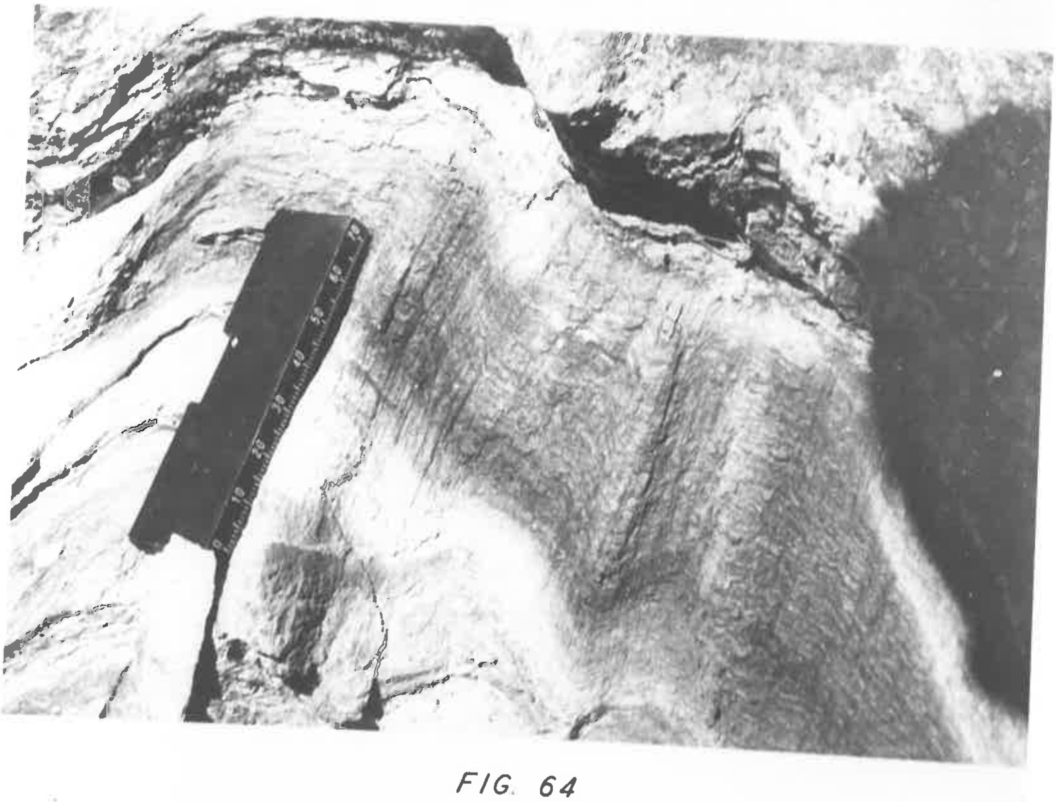


FIG. 64

Fig. 65. A. Folded S_2 surface with differential development of S_3.

Lower Hermitage. (Grid ref. 765 996).

B. Appressed rootless B_2 fold overprinted by S_3.

Ansteys Hill. (Grid ref. 745 952).

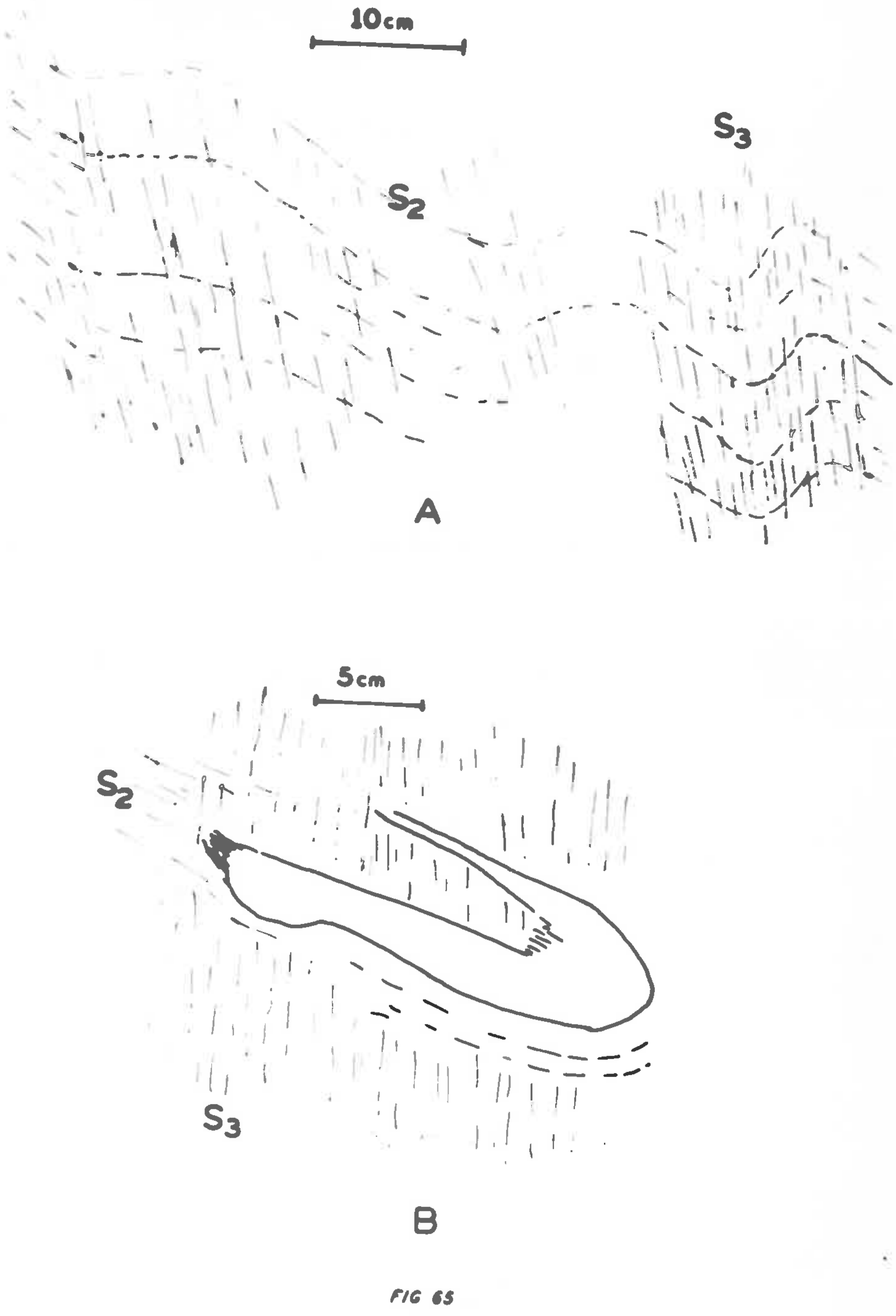


FIG 65

STYLE OF MICROSCOPIC B_3 FOLDS

The development of crenulations is strictly related to the presence of abundant mica in the deformed layer. Where mica is relatively rare, crenulations are not observed. In the micaceous layers, deformation shows as asymmetric waves in S_2 which rapidly die out at the contact with more quartzose layers. The regularity of the corrugations depends on (a) the degree of uniformity of the micaceous layers; (b) the composition of the mica-rich layers and (c) the thickness of the layers.

In very thin (less than 0.2 mm.) and thick (greater than 3 mm.) micaceous laminae S_3 tends to be very irregular and in the thin laminae may be absent. S_3 is most regularly developed in laminae of intermediate thickness and constant composition. Attempts were made to correlate wavelengths of the corrugations with the thickness of the laminae with only limited success owing to variabilities introduced by the irregularity of the layers; a further difficulty is that the wavelength decreases with increasing appression. However there is a suggestion that the wavelength increases with the thickness of the layers (cf. Fig. 66).

Where S_3 is not perpendicular to the layering, the corrugations are markedly asymmetrical and the two limbs have a distinctly different appearance. One of the limbs (e.g., the east limb of antiforms in the middle layer of Fig. 66) is commonly at a low angle to the layering (30° in this case)

Fig. 66. Crenulation cleavage in laminated phyllite.
Specimen 157-627.

Fig. 67. Part of field of Fig. 66.
The left limbs of antiforms of microcrenulations are markedly appressed and contain less quartz than the right limbs.
Crossed polars.

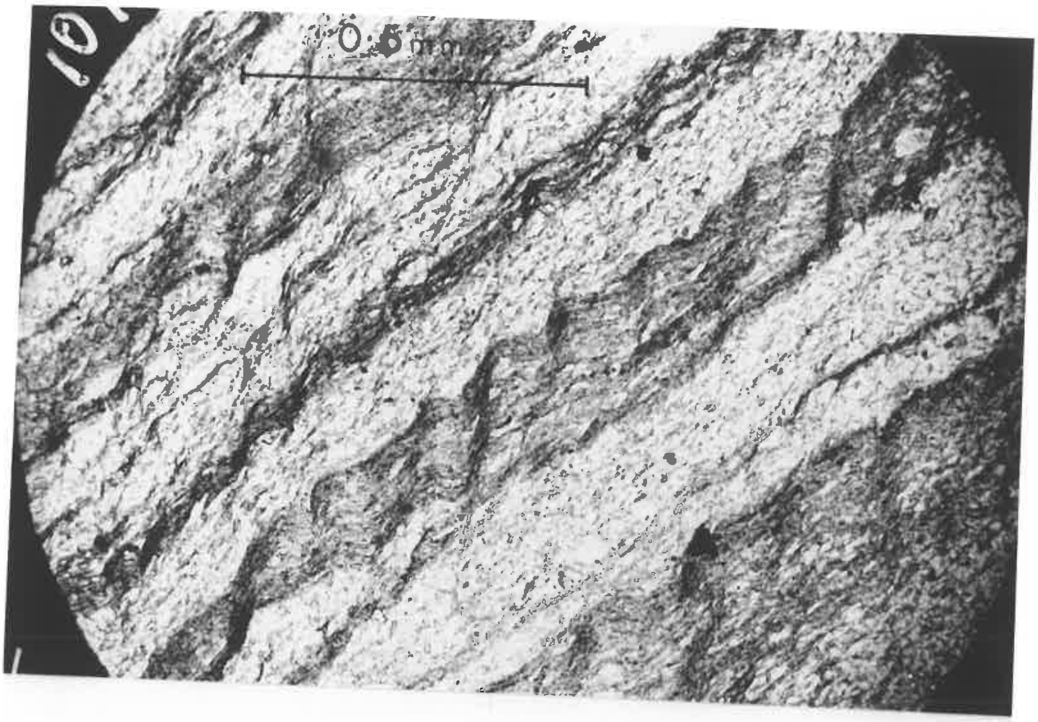


FIG. 66

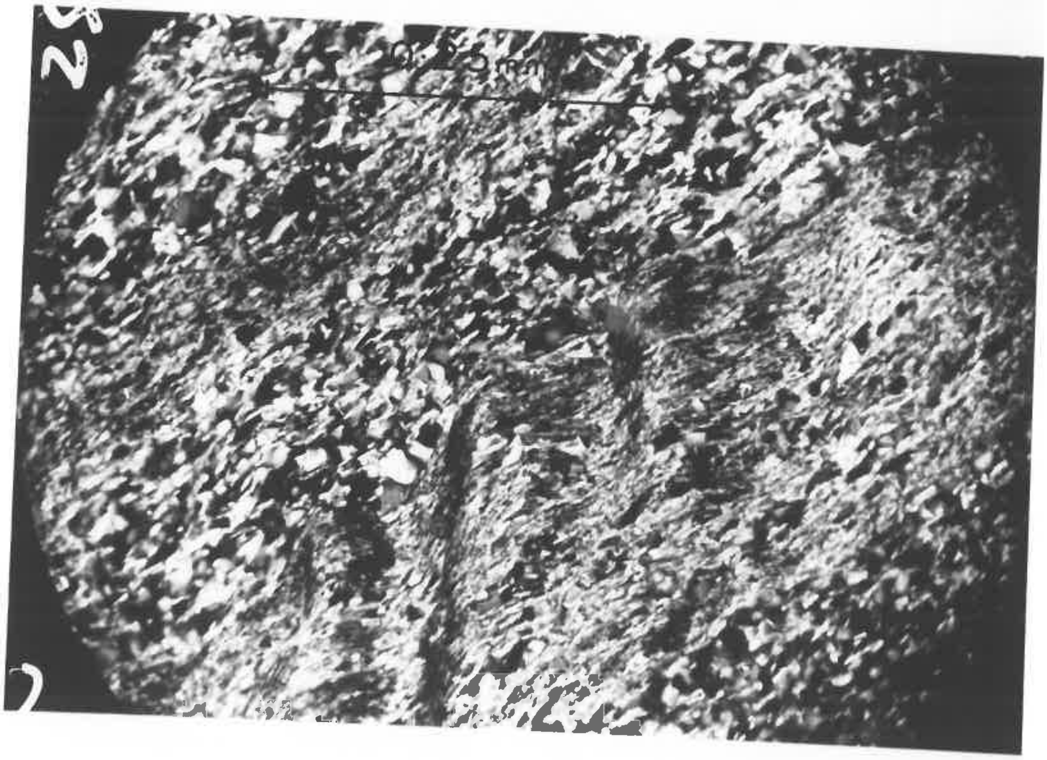


FIG. 67

and contains sericite, chlorite and quartz with a small amount of opaques, the quartz and opaques not markedly elongate. The west limb is at a somewhat higher angle to the layering but in the opposite sense and differs markedly in appearance from the other limb. Individual micaceous grains are of slightly finer grain size than in the east limb and are more closely spaced, and where the degree of appression is pronounced the quartz grains tend to be markedly elongate and thin and there is a suggestion that quartz decreases in amount (Fig. 67). In most cases the appressed limb is richer in very fine-grained opaques. At the contact between the micaceous and the quartzose layers the quartz grains are commonly larger in the troughs of the synforms (Fig. 67), possibly representing a process of simple grain growth consequent on deformation or movement of material from the more appressed limbs.

In rocks with less mica the appearance of S_3 is somewhat different. Discrete S_3 planes tend to develop rather than appressed corrugations although transitions between the two types suggest that the discrete S_3 planes have passed through a stage of being corrugations. The transition stages show that the change is accompanied by a marked increase in the degree of preferred orientation of the micas parallel to S_3 (Hoeppener, 1956) and to some degree of breakdown of the micas, the S_3 planes characterized by very fine-grained micaceous minerals and abundant opaques (Fig. 68).

In many cases differential slip along S_3 has not taken

Fig. 68. S_3 cutting S_2. Micas are parallel to S_3 in S_3 zones. These zones are crowded with an opaque dust.
Spec. 157-934.

Fig. 69. Differential development of S_3 in zone in phyllite. Individual S_3 planes have acted as planes of slip. (Note attitude of S_3 is vertical).
Spec. 157-731 U3.

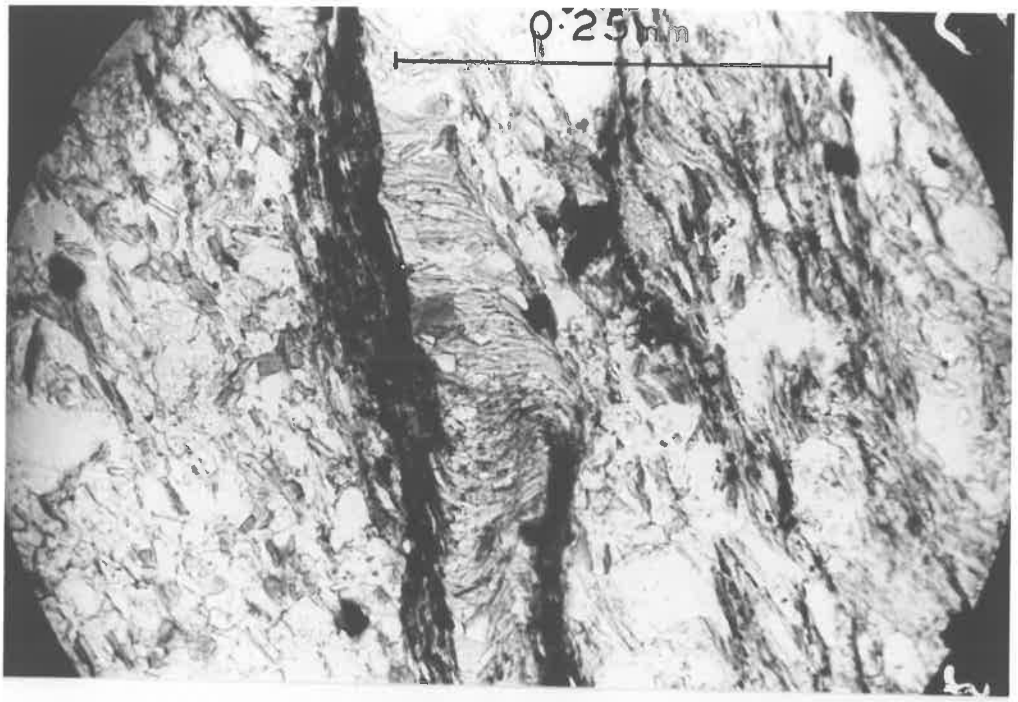


FIG. 68

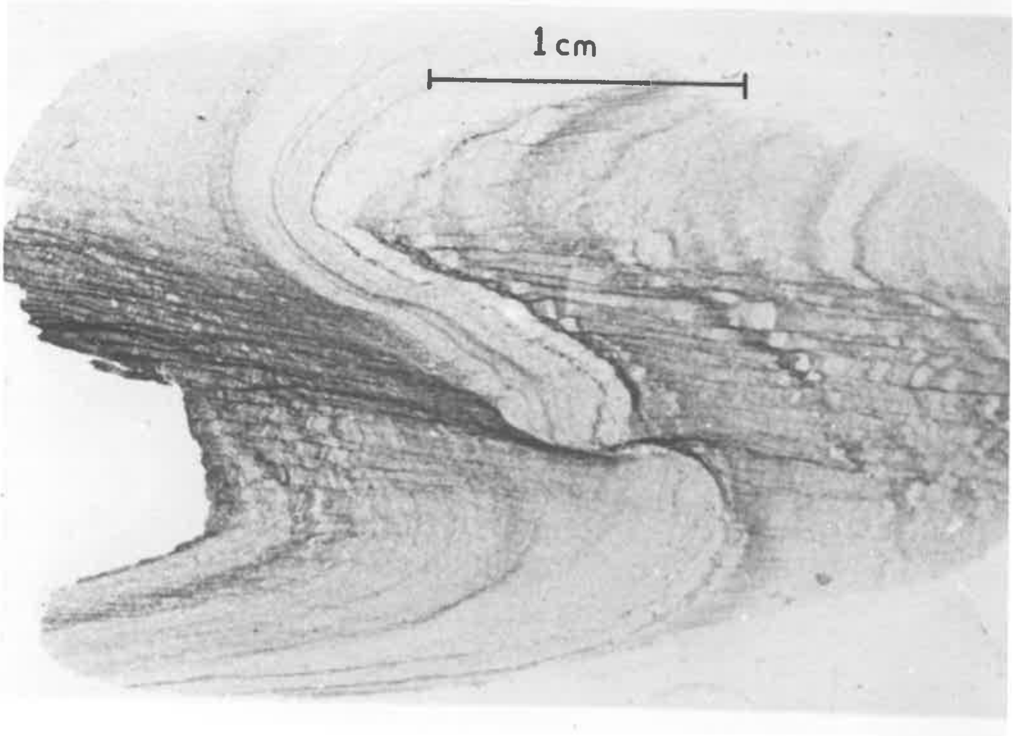


FIG. 69

place, although differential movements by folding are apparent (e.g., Fig. 66). However in some thinly laminated phyllites, slip along discrete S_3 planes has taken place (Fig. 69). The individual slipped blocks (or microlithons, De Sitter, 1956) may retain their original outline (Fig. 69) or may be boudinaged, especially in the more appressed structures.

In the crests of many small scale B_3 folds two crenulation cleavage surfaces are observed, making low angles with each other and having different senses of rotation (Fig. 70). On the limbs only one of the S_3 planes is observed and it is always the plane which makes the smaller angle with the limb. If this is a general relationship it may indicate that the fanning of crenulation cleavages commonly recorded is in reality the development of two cleavage planes at an angle to one another.

In a few localities a new layering is produced parallel to S_3 (Fig. 63). In some cases the layering is the result of transposition of the original layering by folding on S_3, but in many cases the new layering is the result of the differential development of S_3 in certain zones (see also Flinn, 1958 who reports an irregular striping in phyllites caused by crenulation cleavage). In such cases the original layering can still be observed in the quartz-rich portions and the new layering may be quite broad, layers up to 5 cm. being fairly common (Fig. 63). In some cases the layers developed are very regular and cut through alternations of quartzose and micaceous layers

Fig. 70. Development of two S_3 surfaces in the crest of a B_3 fold.

Phyllite. Spec. 157-934.

Fig. 71. Layering developed parallel to S_3 (horizontal) cutting transposed sedimentary layering (sloping). (S_3 is vertical in the outcrop).

Specimen 157 - 731.

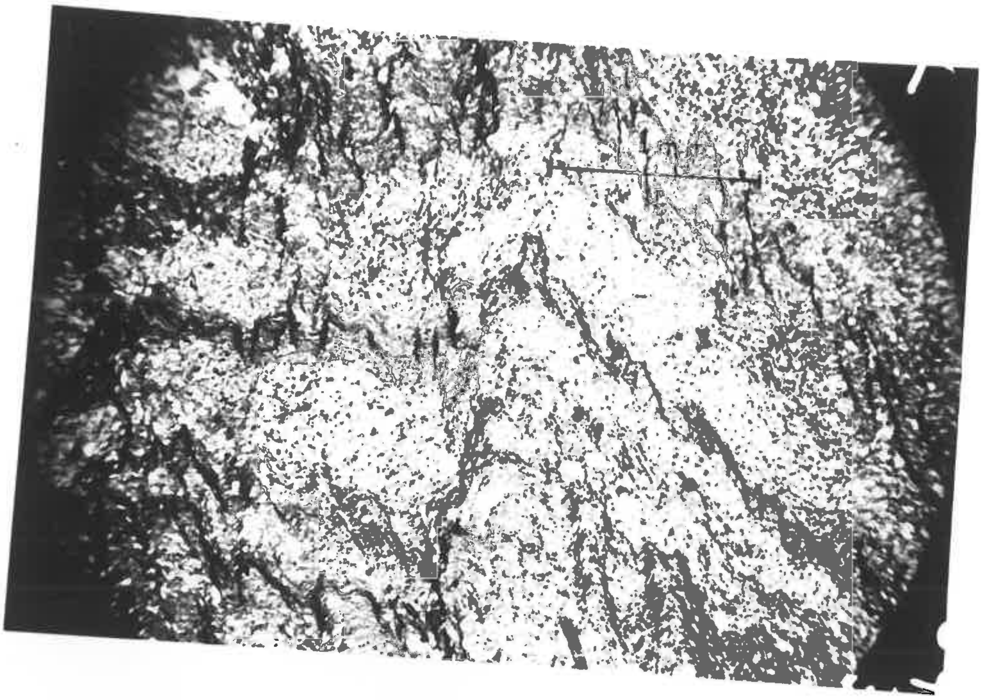


FIG. 70

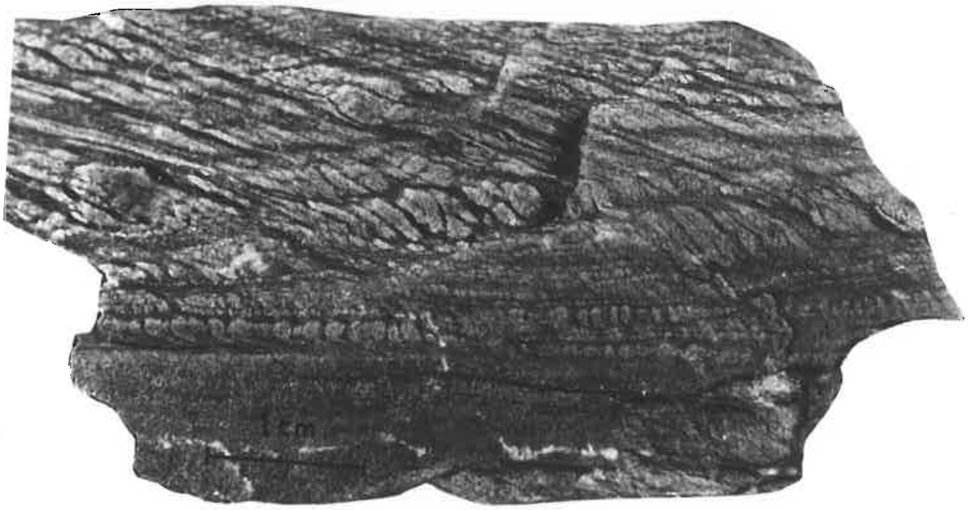


FIG. 71

(Fig. 71) but more commonly the best development of layering occurs in the more phyllitic beds.

In thin section this layering appears less apparent as it is easier to distinguish the original layering. S_3 planes are more abundant and closer together in the darker layers (Fig.69) and it is this difference which gives rise to the mesoscopic layering. Measurements on the thicknesses of original layers involved in the deformation suggest that considerable thinning of the layers has taken place. Measurements perpendicular to the layering and parallel to the axial plane give values inconsistent with either concentric or similar style. (Table III).

The apparent thinning in many of the structures is correlated with a decrease in the quantity of carbonate present. Measurements were made on the S_3 zones and the relatively undeformed crestal regions in 157-731J to confirm this common visual impression (Table IV). Firstly a particularly rich quartz-dolomite-mica layer was chosen, then runs across several layers were made. Difficulties in dealing with restricted bands made the counting of the requisite 2000 grains impossible and so the results should be treated with caution, but it is believed that they indicate the order of changes involved. Thus there is a very real change in composition and the layering is in fact compositional and not simply textural. Quartz as well as carbonate must have migrated out of the S_3 zones to account for the amount of mica present. Rickard (1961) and Dunroth (1962) report a similar increase in the amount of mica in the long limbs of the microflexures and conclude that

TABLE III.

VARIATIONS IN THICKNESS OF A QUARTZ-CARBONATE-MICA LAYER AROUND AN S_3 FOLD.

Measurements from thin section 157-731 U3

Thickness in crestal regions		Thickness on appressed limb.	
Parallel to S_3	Perpendicular to S	Parallel to S_3	Perpendicular to S
1.85 cm.	1.62 cm.	.83 cm.	.59 cm.
1.61 cm.	1.49 cm.	1.55 cm.	.66 cm.
1.3 cm.	1.3 cm.	1.5 cm.	.81 cm.

TABLE IV.

VARIATIONS IN QUARTZ, CARBONATE AND MICA CONTENT IN DIFFERENT PARTS OF AN S_3 FOLD.

Specimen 157 - 731J.

Readings in the quartz-dolomite-mica layers.

	Crestal region	S_3 zone.
Quartz(+ feldspar)	30%	21%
Carbonate	54%	8%
Mica	16%	71%
Number of grains counted	961	853.

Readings in several bands.

	Left hand side	S_3 zone	Right hand side.
Quartz(+ feldspar)	26%	20%	29%
Carbonate	55%	14%	43%
Mica	19%	66%	28%
Number of grains counted	1301	1253	1424

the quartz has migrated to the crestal regions. No significant increase in quartz in the crestal regions appears to have occurred in the Torrens Gorge examples.

Discussion.

It is a common observation that crenulation cleavage occurs only in rocks which have abundant micaceous minerals already possessing a high degree of planar preferred orientation (White, 1949; Gillott, 1956; Rickard, 1961). In general this implies the prior formation of a slaty cleavage or foliation although in certain circumstances a shale might develop a crenulation cleavage. Examples of crenulation cleavage associated with cross-folding are common (e.g., Knill, 1959; Rast, 1958; Weiss and McIntyre, 1957). Crenulation cleavage can also occur as a later stage development in one tectonic phase, especially if the crenulation cleavage bears some symmetrical relation to the pre-existing slaty cleavage (Agron 1950; Cummings and Shackelton, 1955).

In the Torrens Gorge crenulation cleavage is also restricted to micaceous rocks with a good slaty cleavage. The crenulation cleavage is also homoaxial with the slaty cleavage and B_2 folds and is thus thought to be related to the same or parallel system of stresses as the B_2 folds.

The mechanism of folding of the slaty cleavage is in some doubt. The restriction of crenulations to portions where micas are sufficiently abundant to overlap suggests that a concentric slip mechanism may have occurred. Although this mechanism would adequately explain the presence of microcrenulations there seems no reason why this should result in the regular pseudo-orthorhombic B_3 folds. Rather it seems that the folding expresses an overall shortening of the rock perpendicular to S_3 and that the thicker, more massive beds which outline the B_3 folds have adjusted to what could have been axial stresses by folding in a predominantly concentric style. Orthorhombic and monoclinic folds resulting from the application of axial stresses on rocks possessing a pre-existing planar anisotropy have been produced experimentally by Paterson and Weiss (1962). However, an orthorhombic stress system would also produce the folds described above.

The development of a new layering parallel to S_3 is also considered to indicate a shortening perpendicular to S_3. In these rocks this layering is restricted to the more dolomitic micaceous layers and indicates the ease of migration of the dolomite. As noted previously Rickard (op. cit.) reports that such layering can also be achieved by exclusion of quartz, presumably under conditions of higher grade metamorphism than existed in the Torrens region, as the amount of quartz removed in these rocks is not very great.

MISCELLANEOUS STRUCTURES IN THE TORRENS GROUP ROCKS.

Mesoscopic structures other than those described above are rare in the area mapped. A few occurrences of east-west crenulation cleavages have been observed at the western end of the Torrens Gorge and also along the Tea Tree Gully-

Houghton road. They are too limited in number to be considered with the other structures but may be related to the better developed east-west crenulations in the Macclesfield - Strathalbyn area 25 miles southeast of Adelaide. Mr. R. Offler (personal communication) considers these east-west crenulations in the Macclesfield area contemporaneous with the major north-south crenulations and hence indicative of a single episode of triaxial strain. In the area examined by the author, the east-west and north-south crenulations were not seen in the same outcrop so no relationships between the two cleavages could be determined.

In two outcrops good kink folds are observed. In the better of the two outcrops (Grid ref. 867 980, a road cut east of Gumeracha Weir,) the axial surface of the kinks strikes 340° and dips 65° to the west. In thin section (Fig. 72) the kink folds are seen to have overprinted and rotated an earlier crenulation-type cleavage. In the other outcrop (Grid ref. 857 973, in cut for Mannum-Adelaide pipeline), kinks occur with axial surfaces parallel to prominent joints, and are similar to joint drags described by Knill (1961). The kinks observed had axial surfaces which strike 035°, dip 25° to the west; and strike 025°, dip 40° to the west. The relationship of the kinks to other recognizable structures is not known.

Fig. 72. Kink folds in dolomitic phyllite.
The kink folds (S_4 axial surface) has rotated both the laminations and a crenulation cleavage (S_3).
Spec. 157-1336.

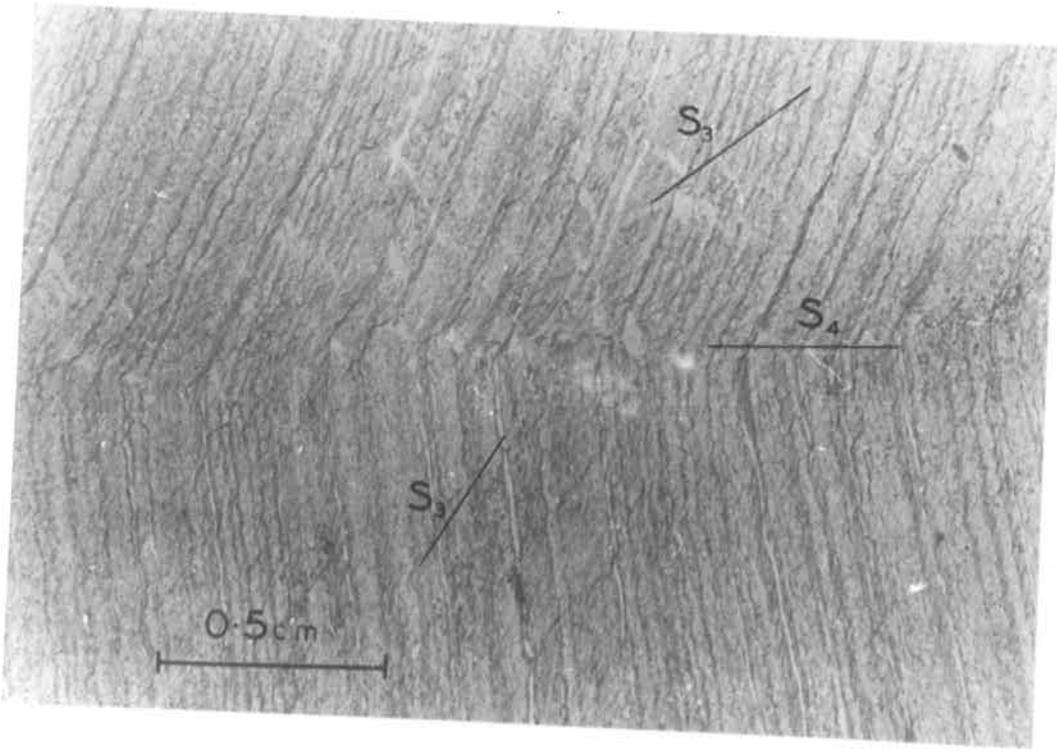


FIG. 72

CHAPTER IX.

RELATIONSHIPS BETWEEN STRUCTURES OF THE HOUGHTON COMPLEX AND TORRENS GROUP.

A discussion of the orientation of the structural elements has been left out of the previous sections so that a unified analysis of the structural relationships between different rocks and phases of deformation can be undertaken.

In view of the absence of macroscopic fold surfaces in the Houghton Complex extensive use has to be made of methods employing statistical analysis of the structural elements. These methods are in common use in areas such as the Scottish Highlands and review papers regarding multiple deformations have fully treated their use (Weiss, 1957 ; Weiss, 1959 ; Ramsey, 1960).

Orientation of structural elements in the Houghton Complex.

Due to the large scale retrogression in the Houghton Complex the visible structures are very few. The sericitic foliation is the most prominent structural element and shows a marked degree of preferred orientation. The statistical maximum is 350° 58°E and there is a tendency for the readings to spread about a great circle whose pole plunges at 55° in a direction 093° (Fig. 73B). Deviations from the norm are not great; in the southern part of the area the maximum of foliation is 340° 60°E and in the north it is 355° 75° E. The regional variations have been summarized in Plate II and selected readings plotted on Plate I.

Fig. 73. Collective diagrams of structural elements of the Houghton Complex.

A. Poles to layering 531 points.
Contours 1-2-3-4-5% per 1% area.

B. Poles to foliation 512 points.
Contours 1-3-5-7-9-11% per 1% area.

C. Lineation 149 points
Contours 1-3-5-7-9% per 1% area.
(Maximum concentration 12%).

HOUGHTON COMPLEX

COLLECTIVE DIAGRAMS

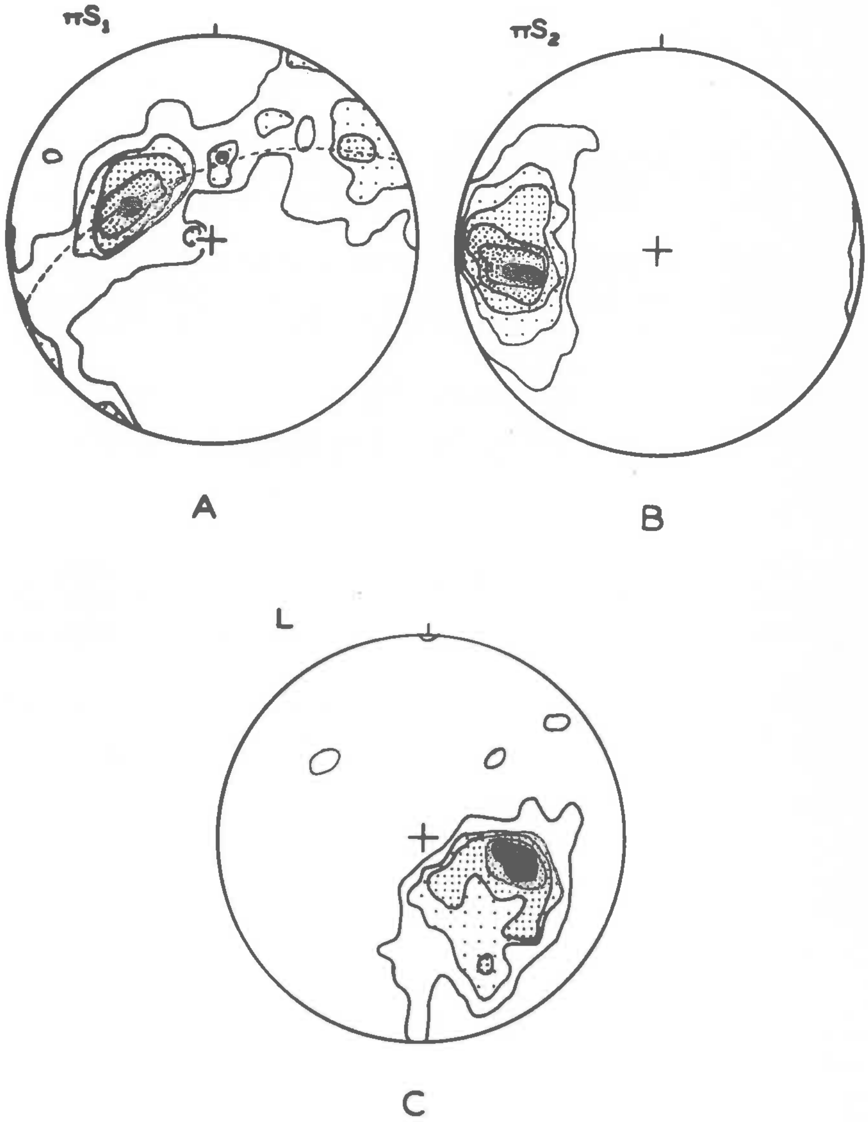


FIG 73

The down dip lineation in the sericite schists has a maximum plunging 50° in a direction 104° which is essentially the same as the axis of macroscopic warping of the foliation.

Layering is observed locally in the highly retrograde rocks but is subparallel to the foliation, and thus no distinct macroscopic analysis was possible. In the southwest part of the Complex layering is fairly common, especially in the feldspar gneisses. A collective plot of readings on layering shows a spread about a diffuse girdle whose axis plunges 38° in a direction 160°, significantly different from the warping axis of the foliation.

Plots of individual subareas show a wide variation in fold axes but give fairly well-defined axes within each area. The notable exceptions are north of Houghton(Subarea I), where the girdle of poles to layering is quite diffuse, and Kersbrook (Subarea VIII), where folding is not apparent.

The distribution of β axes in the Houghton Complex is similar to the orientation of the lineations (Fig. 75) whether of layering or foliation, although as is obvious from the collective plot of layering (Fig. 72A), the β axes of layering plunge in a more southerly direction than the β axes of foliation. The distribution of all the β axes lies roughly in a plane which is parallel to the statistical maximum of the foliation planes, although a somewhat better fit can be obtained using a plane of strike 16° dip 50° E. However the number of axes involved (17) makes the determination of the plane containing them very uncertain.

Orientation of Structural Elements in the Torrens Group.

Measurements for statistical analysis over the whole area were made on bedding S; cleavage S_2; and fold axes and bedding-cleavage intersections. Attitudes of bedding were taken only where mesoscopic folding was not apparent, and the correspondence of the linear elements with the statistical macroscopic fold axes indicate that the macroscopic folds for most subareas have axes parallel to the mesoscopic fold axes. Attitudes of crenulation cleavages have been measured where they occur, that is, in the southwestern part of the area, but a general macroscopic analysis of this cleavage is not possible as B_2, B_3, S, S_2 and S_3 are all roughly cozonal in this area. However, in other areas in the Adelaide Hills this is not always so and hence this area may give an atypical relationship of B_2 and B_3 folding. Discussion of B_3 geometry has therefore been limited to the style of the folds until more published data is available from these areas.

Collective plots of the various structural elements are shown in Fig. 74. Synoptic data from these plots are as follows:

Attitude of S	Statistical maximum orientation (7% per 1% area), strike 346°, dip 35° east.
	Poles to bedding spread about a very diffuse plane whose pole plunges 10° in a direction 9°.
Attitude of S_2	Statistical maximum (17%), strike 12°, dip 60° east. A secondary maximum (9%) coincides with the maximum of bedding attitudes.

Fig. 74. Collective diagrams of structural elements in the Torrens Group.

A. Poles to bedding. 729 points
Contours 1-2-3-4-5-6% per 1% area.

B. Poles to cleavage (S_2) 164 points
Contours 1-3-6-9-12% per 1% area.

C. Poles to cleavage (S_3). 53 points
Contours 2-6-9-12-15% per 1% area.

D. Fold axes (B_2) 170 points
Contours 1-3-5-7-9% per 1% area.

TORRENS GROUP

COLLECTIVE DIAGRAMS

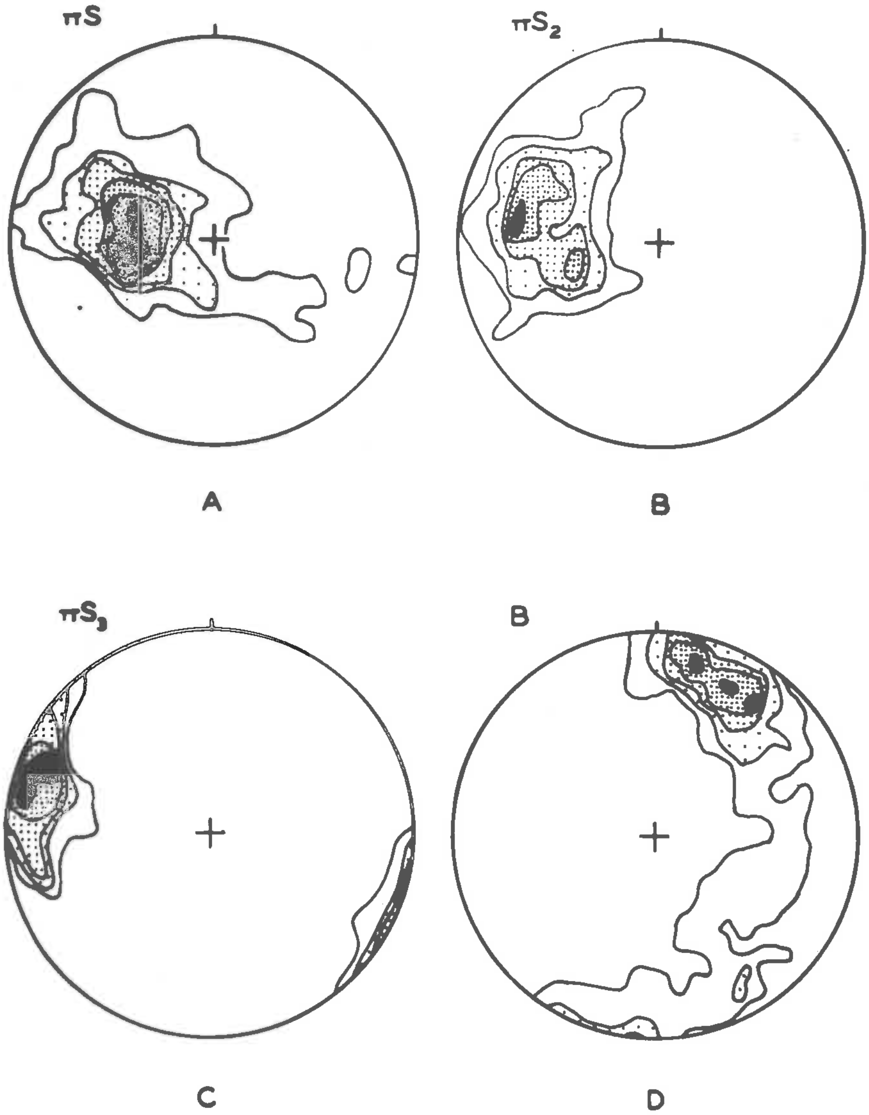


FIG 74

	S_2 tends to spread about a axis plunging 55° in a direction 92°.
Attitude of S_3	Statistical maximum (15%) strike 12°, dip 80-85° east. A plane in which the poles spread can be defined only approximately. Its pole plunges approximately 80° in a direction 45°.
Attitude of B_2	Statistical B_2 plunges at about 20° in a direction 015°. However the B_2 axes are fairly well dispersed in a plane which strikes 5° and dips 45° east. This is the only collective plot which expresses any major variation in structure throughout the area.

Subarea plots are shown on Plate II. In particular these show a macroscopic variation of B_2 folds which is in harmony with the variation of the mesoscopic B_2 folds. If the synoptic diagrams of β's of $\pi S, \pi S_2$ and the collective mesoscopic B diagram of the Torrens Group, and the β's of $\pi S_1, \pi S_2$ and the collective lineation diagrams of the Houghton Complex are compared (Fig. 75), it is seen that all the major linear elements in both the Houghton Complex and the Torrens Group lie approximately in a single plane, although the Houghton Complex linear elements are grouped quite differently from those of the Torrens Group structures.

This plane is approximately parallel to the foliation in the two sequences and it appears that the structures in the two sequences show a good degree of correlation and may be related genetically.

SYNOPTIC DIAGRAMS

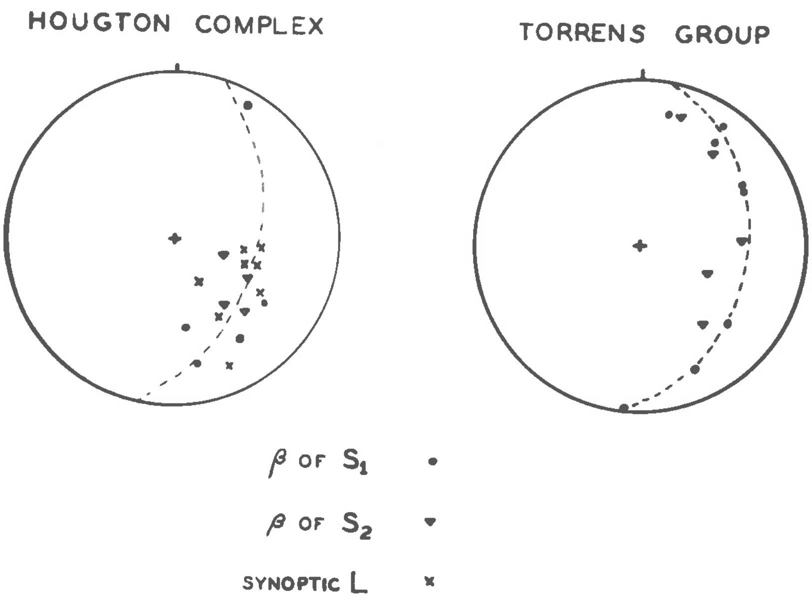


FIG 75

Spry has suggested that the production of the foliation in the Houghton Complex is contemporaneous with the folding in the Torrens Group, but evidence of retrograde fabrics in the pebbles in the Torrens Group conglomerates suggests that much of the retrograde metamorphism took place prior to the deposition of these conglomerates. It seems probable therefore that the retrograde metamorphism began before the deposition of the Torrens Group and continued during the period of folding of the Torrens Group and that the operation of the local stress field depended to a large extent on the presence of already active planar directions in the Houghton Complex.

The spread in the foliation attitudes in both the Torrens Group and Houghton Complex is similar and appears related to an axis of warping which plunges due east. Although microscopic evidence for folding on this axis is limited in the area mapped by the author, subsidiary structures on a vertical east-west axial surface are common in the Strathalbyn area as noted above. Cross-folds perpendicular to the main fold system are common elseqhere and are generally interpreted as expressions of one period of strain (Rast and Platt, 1957).

The spread of B_2 and β_2 axes in the Torrens Group suggests that the B_2 folds are overprinted on earlier structures; alternatively large-scale inhomogeneities in the deformation may have produced variable plunges (Harland and Bayly, 1958; Voll, 1960). Field evidence suggesting overprinting is lacking but the en echelon distribution of "Archaean" cores in the whole of the Mount Lofty ranges (Fig. 2) is consistent with the patterns obtained in experimental superposed folds. (O'Driscoll 1962).

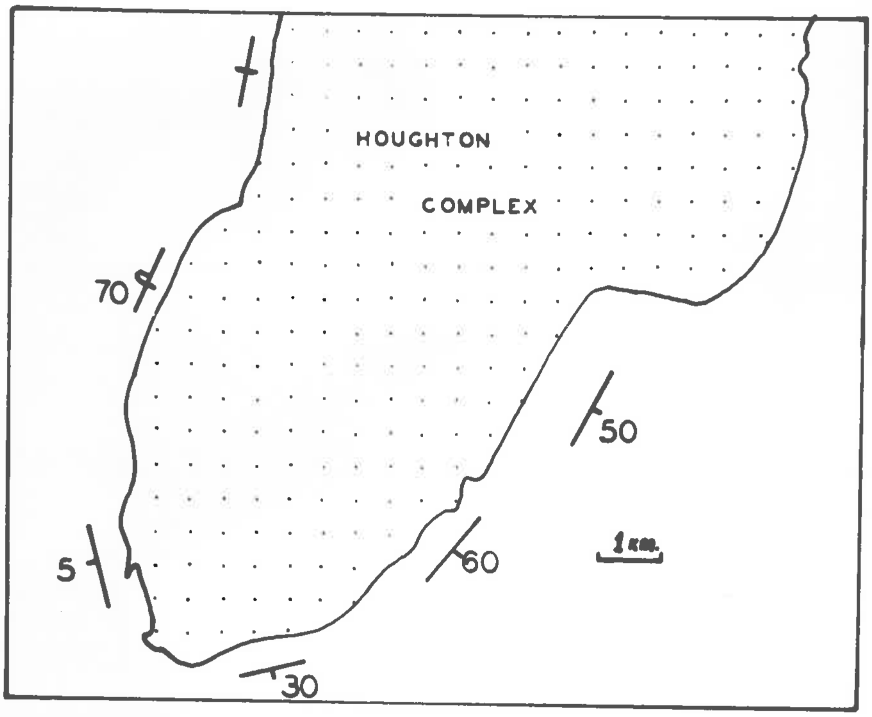
If it is assumed that very large pre-B_2 folds exist in the Mount Lofty ranges, it is possible to calculate the plunge of the axes of such folds. In any one subarea the attitude of β_2 will be the intersection of S_2 with the original attitude of the layering in that subarea (Weiss 1959, Ramsey 1960, McBirney and Best 1961). If another linear direction in the layering can be determined the attitude of the layering can then be calculated. The only other linear direction which can be determined in the area mapped is the contact between the Houghton Complex and the Torrens Group which may be the intersection of the form surface of the original fold with the topography.

Using these assumptions the calculation of the attitudes of the folded surface in any subarea is straightforward. The strike of the local contact is the strike of the pre-B_2 fold limb and the β_2 fold axis lies in this limb. Hence if these are plotted on a stereo net the great circle passing through their projections is the projection of that limb. The results of such an analysis are presented in Fig. 76, subject to the degree of uncertainty of the assumptions, the near parallelism of the fold axes and strikes of the contact in some subareas and the probability that the contacts are faults. The fold axis indicated by these calculations plunges to the southwest at about 10° and is not a probable B_2 axis.

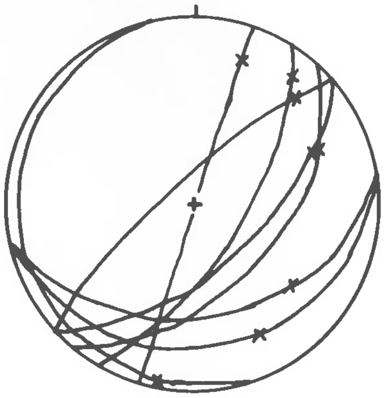
Fig. 76. A. Possible attitudes of limbs of B_1 fold.

B. Calculation of attitudes of limbs of B_1 fold. X's are the statistical fold axes of various small areas.

C. Uncontoured β diagram of the intersection of the limbs of B_1 fold.



A


B

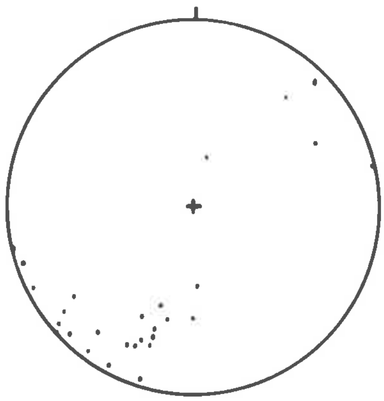

C

FIG. 76

The attitudes of the various parts of the supposed fold are shown in Fig. 76. The calculation of such a folded surface does not prove that it exists and it is hoped that better evidence for or against such structures may be found in other areas of the Mt. Lofty Ranges.

The relationship of the S_2 and S_3 planes is not clear. Their intersection is roughly horizontal and is parallel to the local fold axis. The attitudes of the S_3 planes are steeper than the S_2 planes in this area although in an area to the southwest, Mr. I. Pontifex, in the course of field mapping for an honours thesis, showed the author some examples of crenulation cleavage which were dipping at very low angles to the east. It may be that a symmetrical relationship between the slaty and crenulation cleavages can be proved by further investigations. Symmetrical crenulation cleavages have been reported by Peach and others (1909) and Agron (1950), and these authors conclude that such a symmetrical relationship implies that the crenulation cleavage is related to the same series of stresses as that of the slaty cleavage.

CHAPTER X.

THE WHEY WHEY CREEK AREA.

INTRODUCTION.

The second area investigated lies within the Olary Province as defined by Campana and King (1958). The rocks of the Olary Province fall into three main groups; a "crystalline basement" named the Willyama Complex by Mawson (1912); a sequence of metamorphosed sedimentary rocks of Upper Precambrian age and unconsolidated alluvial and pediment deposits. A small part of the Province was selected for study (see Fig. 4) to investigate whether the different metamorphic grade (compared with the Houghton Area) has affected the structural and metamorphic relationships between the basement rocks and the overlying Upper Precambrian.

Certain aspects of the geology of the region were chosen for special study after a preliminary investigation of the area; 1) The boundary relationships between the two metamorphic sequences were investigated in detail in the field. The basal rocks of the Torrens Group are very well exposed (in marked contrast to the poor exposures of the Paracombe - Mt. Gawler area); 2) The relationships of the structures in the Willyama Complex to those in the Upper Precambrian were studied, both on a mesoscopic scale and statistically on a macroscopic scale. These studies were supplemented by the examination of selected specimens in the laboratory. 3) The relationships between the structures in the Torrens Group to those in the overlying Sturt Group were examined.

A brief account of the petrology is presented to establish the metamorphic environment in which the rocks were deformed.

CHAPTER XI.

ROCK TYPES OF THE WILLYAMA COMPLEX.

In the Olary region Campana and King (1958) recognized three broad units in the Willyama Complex (which they called Archaean). These are (oldest first): 1. Weekeroo Billeroo schists, 2. Ethiudna calc-silicate group, 3. Arkosic quartzites, grading into granite-gneisses.

Unfortunately, the present author was unable to find any evidence for the relative ages of the sequence of metamorphic rocks in the part of the Complex around Whey Whey Creek. Campana does not state the evidence on which his succession is based and appears to have relied on the regional distribution of different lithologies.

In the mapping undertaken by the author the Willyama Complex was subdivided on the basis of dominant lithologies, and the distribution of these units is shown on Plate III.

1. Mica schists.

Extensive outcrops of brown and grey mica schist occur in several parts of the area. The main occurrences are in the neighbourhood of the central amphibolite (Subareas II, III and IV)[1]; south of the large granite in the northwest (Subarea VIII) and north of the granite in the northeast of the area (Subarea VII). Campana (1956) maps the rocks east of the eastern Torrens

1. Localities and the distribution of subareas are shown on Plate IV.

Group "sleeve" (Subarea VI) as mica schist but these rocks are granitic gneisses and granites. He does not differentiate between the mica schists and the granite in the northwest (Subarea VIII).

The mica schists in all three regions are very similar and are part of Campana's Weekeroo Billeroo schists. They form a monotonous group of rocks which are dominantly quartz-mica schists, commonly showing a silky lustre in hand specimen due to the concentration of muscovite in planes parallel to the foliation. Other rock types are rare but some quartzites (less than a metre wide) are observed as discontinuous lenses parallel to the foliation. In some parts of the area (e.g. in parts of Subarea II, near Locality 32) lenses of quartzose material less than 1 cm. in thickness are relatively common. These lenses which are commonly very tightly folded are thought to be transposed remnants of originally coarser-grained layers (Fig. 77).

The schists south of the amphibolite pass into a group of rocks transitional to the gneisses of subarea I. These rocks contain thicker layers (up to 20 cm. of quartzo-feldspathic material interlayered with the mica schist, the layers commonly being quite regular and parallel-sided. The foliation in the schists and the preferred orientation of the micas in the quartz-rich layers are parallel to the layering. The differences in composition probably represent original sedimentary differences, accentuated by transposition and metamorphic differentiation.

Pegmatites are common in the schists as interfolial bodies, although some small cross-cutting pegmatites are observed. Some of the pegmatites can be traced for many hundreds of metres and some may be up to 30 metres in width. More commonly they occur as small bodies, and many are folded with the layering. They are quartz-feldspar pegmatites containing black tourmaline and books of muscovite as common accessories. Beryl occurs in a few pegmatites. No pegmatites are found cutting the Upper Precambrian rocks.

Quartz veins are also common in the Willyama rocks. Some of these are cross-cutting and a few are observed crossing the contact into the Torrens Group.[1]

In thin section the mica schists are seen to consist essentially of quartz and muscovite with minor biotite, chlorite and feldspar; light brown to blue pleochroic tourmaline is a common accessory, and a colourless garnet is found in some specimens. Staurolite, chloritoid and sillimanite occur together in one locality just south of the amphibolite (211-24), the sillimanite occurring in needles cutting all other minerals. Clusters of muscovite are common in the northeast outcrops (Subarea VII) suggesting original andalusite; K. J. Mills reports

1. Quartz veins parallel to the foliation are very common in parts of the Upper Precambrian, especially in the northern part of the area.

(personal communication) the presence of andalusite in the schists about a mile east of the area mapped.

In the mica-rich rocks (50-75% mica, e.g. 231-33 and 40) the planar orientation of the micas is very marked and the micas are commonly in aggregates more than a centimetre in length making estimates of the grain size difficult. Where individual grains are observed 0.2 mm. is the common grain size. Both biotite and chlorite occur in crystallographic continuity with muscovite. Clusters of biotite and chlorite are also observed, some parallel to the foliation and some cutting it. The quartz and feldspar grains occur between mica flakes and are clear, generally rectangular grains, and are elongate parallel to the mica {001} planes. Where clusters of quartz and feldspar are observed the grains are more nearly equidimensional and the above example illustrates the marked control exerted by the micas on the habit of the felsic minerals.

With an increase in quartz and feldspar the micas show less tendency to aggregate and occur as isolated grains still showing a marked preferred orientation. In such rocks a yellow to bronze pleochroic biotite tends to be more important quantitatively and in some rocks is more abundant than muscovite.

Most of the schists show poor or no lineations where they have not developed crenulation cleavages. Some however show a fairly strong lineation parallel to the small appressed folds previously mentioned (Fig. 77). In these rocks (e.g. 231-57 and 60) the micas are concentrated in layers about a millimetre apart, parallel to the foliation, and in these layers they have a strong planar orientation which gives the rock a well-developed fissility. In between the mica-rich layers are lensoid zones of quartz, feldspar and mica. The mica shows a good planar orientation even in these zones, but in some lenses the micas outline microfolds even where individual micas are not in contact with one another (Fig. 78). Whether these structures should be grouped with the more obvious crenulation cleavages or with the normal foliation is not known.

In the area around the western and northeastern margins of the amphibolite, the orientation of micas is not parallel to the layering but is parallel to the axial surface of small folds (231-61, 62 and 65). These rocks are richer in quartz and finer-grained than the typical mica schist and can be mistaken for the Torrens Group mica schists. However, they are overlain by the basal conglomerate and in the absence of unsuspected structural complications they are included in the Willyama Complex. Moreover, they grade into typical mica schists over a short distance (231-60).

In thin sections of the layered mica schists, the micas show a good planar preferred orientation. This orientation is parallel to the axial surface of small folds, and the preferred orientation is stronger in rocks with higher

Fig. 77. Appressed folds in laminated mica schist.

Scale in millimetres.

Subarea II. Plate IV.

Fig. 78. Rotated attitude of micas in quartz-rich layer in mica schist.

Spec. 231-60. Willyama Complex.

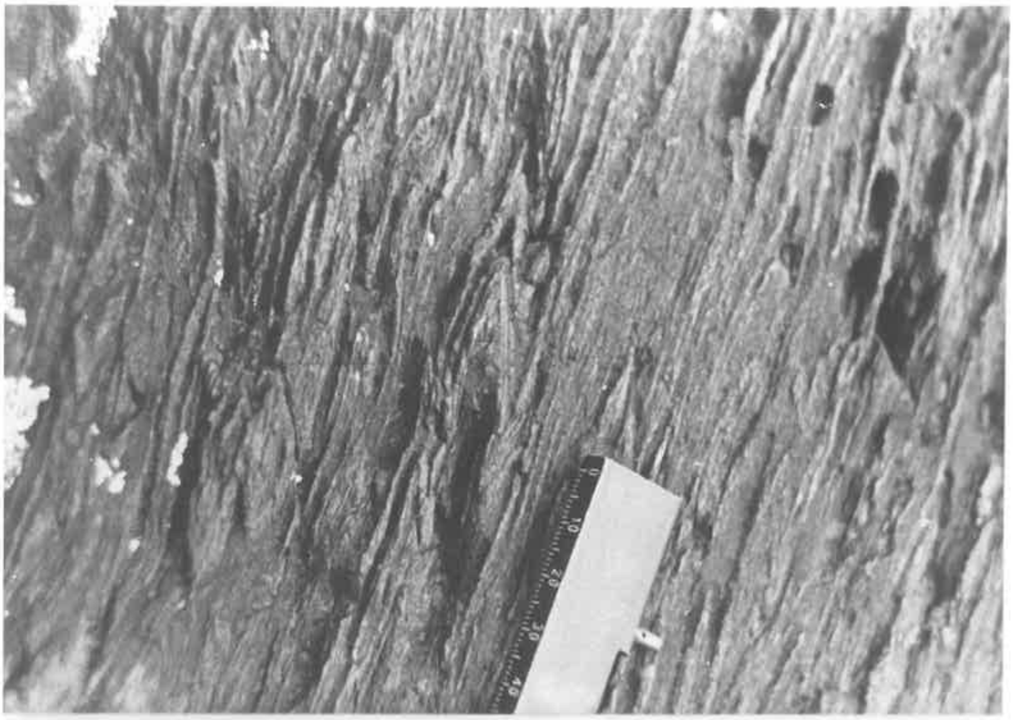


FIG. 77

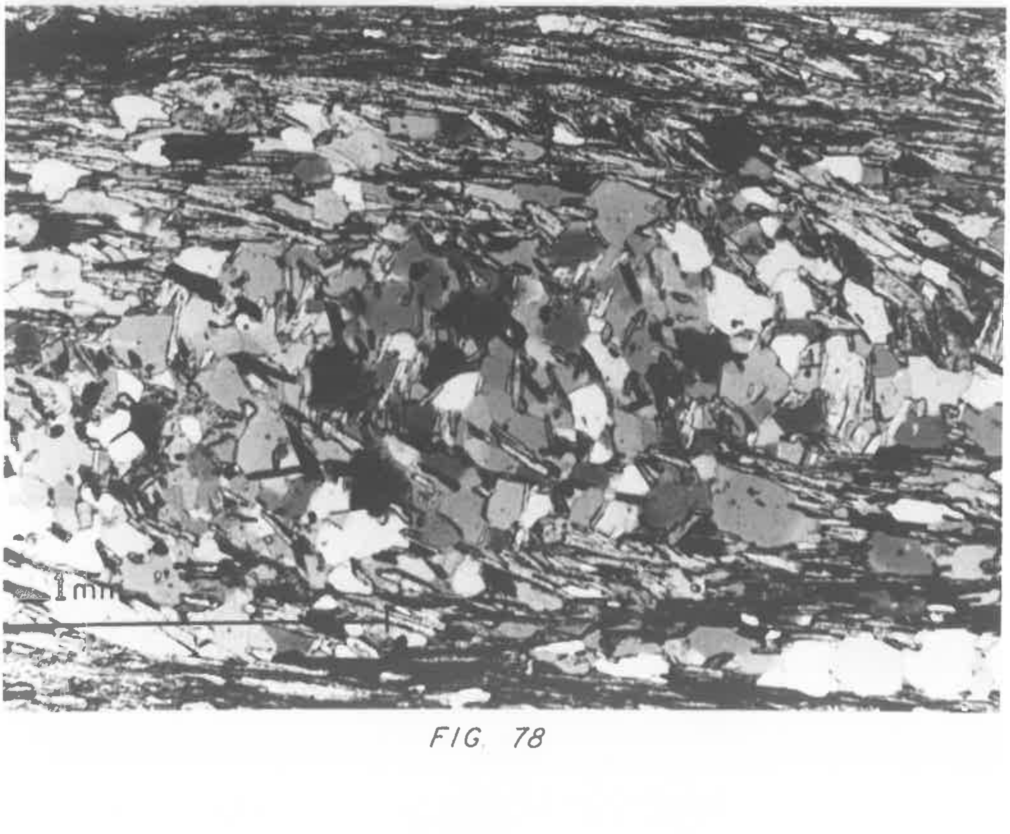


FIG. 78

proportions of mica (e.g. 231-61, with 20% mica, shows a strong preferred orientation whereas 231-65, with less than 5% mica, shows a poor preferred orientation). In a few of the mica-rich laminations the micas lie parallel to the layering and they may outline weak micro-crenulations (Fig. 79) or show a well-developed crenulation cleavage. The crenulation cleavage is subparallel to the foliation in the less micaceous laminations and suggests a genetic relationship, although there is no indication that the foliation is a crenulation cleavage.

2. Platy gneisses.

With an increase in the quartz and feldspar content the mica schists grade into quartz-feldspar gneisses. In the region south of the amphibolite the transitional rock types are very striking and consist of interlayered mica schists and platy gneisses. Near the gneiss unit, thin (3 to 20 cm. quartzo-feldspathic interlayers are more abundant and the rocks grade into quartz-feldspar gneisses with micaceous partings.

With a decrease in mica the rocks grade into quartz-feldspar rocks which resemble aplites in hand specimen. A large zone of these rocks occurs south of Localities 38 and 76 in Subarea 1. These rocks may be well-layered, the layering defined by accessory minerals. The rocks consist essentially of quartz and albite (An_{0-5}) with rarer potash feldspar and accessory rutile, haematite, amphibole and apatite. The

Fig. 79. Weak microcrenulations in schist.
Spec. 231-63. Willyama Complex.

Fig. 80. Irregular layering in migmatite.
Subarea VI. Plate IV.

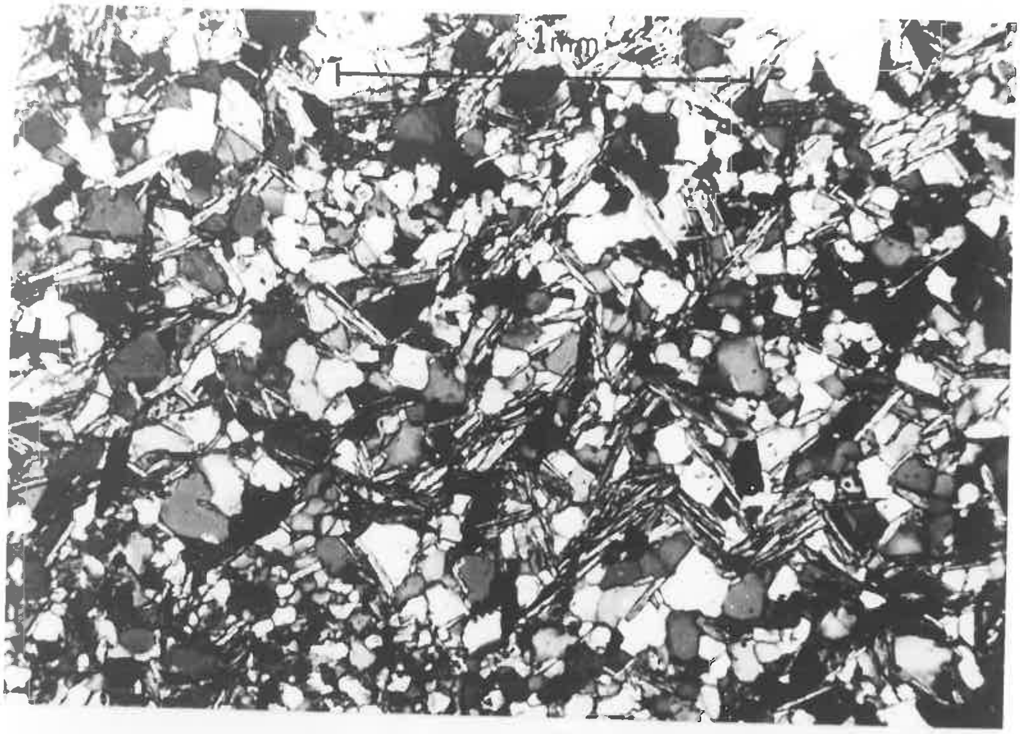


FIG. 79

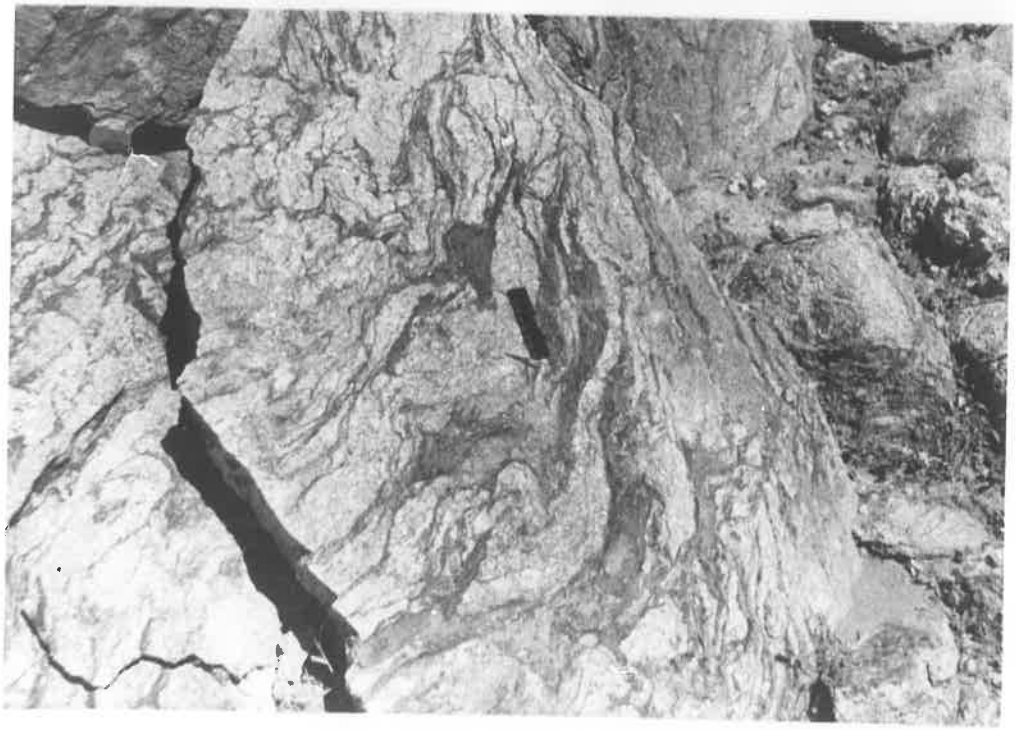


FIG. 80

texture is granoblastic. Quartz commonly shows strain shadows and the plagioclase shows rare offsets of the twin lamellae along microfaults. Campana and King (1958) interpret these rocks as recrystallized arkoses.

The typical gneiss of this subarea (231-24, 25 and 26) is characterized by the association quartz and feldspar (both microcline and albite) with minor muscovite and biotite. The rocks are well layered, due to the different proportions of mica to felsic minerals. Some layers are very rich in mica (biotite most common) and the mica is always oriented parallel to the layering.

In some parts of the gneiss complex, significant pods of mica schist are found (e.g. 231-27 and 29). These are not significantly different from the major occurrences of mica schists although some facies of them are very rich in biotite (e.g. 231-29B) with a very good planar preferred orientation.

Although the layering in the gneisses is remarkably planar over much of the area another type of layering is observed in the south. This layering consists of anastomosing veins of mica-rich material which divide the quartzo-feldspathic rocks into distinct lenses. In thin section (231-25) the mica-rich zones are curved, irregular and discontinuous and appear as zones of retrogression, rich in sericite with some biotite. The areas between these zones appear to be little altered and are typically quartz, feldspar, biotite gneisses.

3. Granite gneisses.

Limited areas of granitic gneisses occur in the eastern part of the area mapped and constitute the bulk of the rocks of the Willyama Complex south of the mapped area. The rocks vary from migmatites to granite gneisses with a planar foliation. The migmatites are characterised by an irregular development of micaceous zones (Fig. 80) which apparently are unrelated to the other structure elements of the complex. These micaceous zones have a similar appearance to the zones of retrogression in the platy gneisses except that they are composed essentially of dark brown biotite rather than sericite.

The other gneisses which are grouped with these migmatites are well-layered quartz-feldspar gneisses. The contacts between the gneisses and migmatites are irregular and it is possible that the migmatites are derived from the gneisses. The gneisses consist of a granoblastic aggregate of quartz and plagioclase (An_{0-3}) with minor dark brown biotite and accessory zircon, apatite, sphene and muscovite. The layering is defined by varying proportions of biotite and the accessory minerals. Strain effects are rare, some of the micas are bent slightly and a few plagioclase grains show small offsets of the twin lamellae. Quartz commonly shows no strain shadows.

4. Granitoid rocks. ("Granites" on Plate III")

The areas marked granite on the map are a varied suite of rocks but were included together as they appeared to be structurally similar. They have not been studied in detail and only a brief description by area is given here.

The granite area in the northwest (north of Subarea VIII) is mostly non-layered. The granite grades through a zone of interlayered granite and mica schist into relatively unaltered mica schist along its southern margin. Elongate inclusions of the mica schist in the granite occur in various places along its southern margin. The approximate southern limit of granitic foliae within the schist is marked on the map by a broken line through the centre of the schist.

The "granites" (231-91 and 96A) consist essentially of quartz, microcline and plagioclase (An_{0-5}) with accessory muscovite and dark brown biotite with pleochroic haloes . Microcline is more abundant than plagioclase. The original textures appear to have been modified somewhat although relic hypidiomorphic texture is apparent. The quartz grains show strain shadows, granulation and some recrystallization; the plagioclase shows a flexing of the albite twin lamellae and the microcline shows crinkled twinning. The micas are bent but in general are not kinked. Over much of the outcrop the granite shows a foliation marked by the preferred orientation of the accessory micas. Layering is rare except along the southern transition zone.

The other areas of granitoid rocks are found along the eastern margin of the mapped area. In this region several rock types have been grouped together for simplicity. The northern area of these granites (Subarea VI) is divisible into two distinct groups which are distinguished on the map by a dotted line through Subarea VI. The northern part of this mass is very similar in appearance to the granite in the northwest; massive, but with a good preferred orientation of micas.

The northern contact of the granite with the mica schists is sharp. The schists contain numerous pegmatitic foliae near the contact (Fig. 82), but otherwise appear to be unmetamorphosed. The foliation in the schist is parallel to the contact with the granite, even though the contact is curved. This relationship suggests that the granite has pushed aside the schist during intrusion.

The southern part of this granite mass consists of inter-layered granitic material and well-layered gneisses. The boundary of the "granite" is considered to be at the southern-most occurrence of massive granite in the area. The granitic portions are indistinguishable from the main mass of granite. The gneissic portions are well-layered (Fig. 81) quartz-feldspar gneisses (231-69 and 70) similar to the granitic gneisses south of the "granite".

The granite in the extreme southeast part of the area is a fine-grained quartz-feldspar-biotite granite. The granite is commonly massive but in parts the biotite has a good preferred orientation; the granite is rarely layered. In the southwest portion of this granite mass, migmatites and gneisses occur in patches and appear to be more numerous beyond the

Fig. 81. Pegmatitic schlieren in mica schist close to contact with granite. Subarea VII. Plate IV. Locality 84.

Fig. 82. Layering in quartz-plagioclase gneiss. Subarea VI. Plate IV.

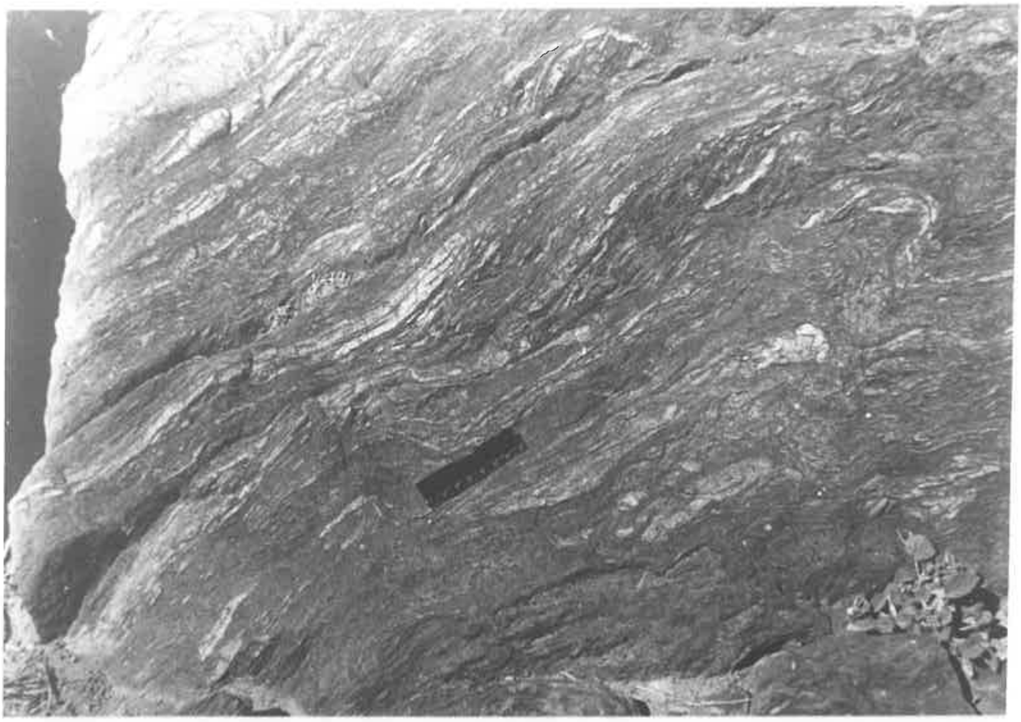


FIG. 81

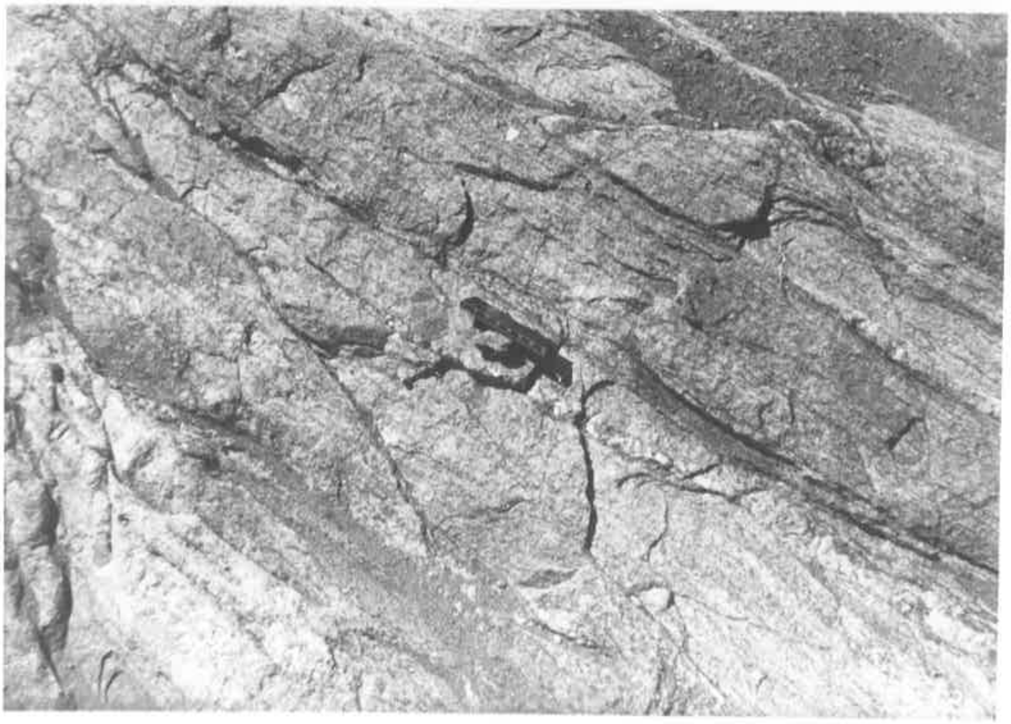


FIG. 82

southern boundary of the map.

Amphibolites and pegmatites.

Bodies of amphibolite and pegmatite are widespread in the Willyama Complex. In the area mapped all of the amphibolites are considered to be igneous in origin; none can be equated with Campana's Ethiudna calc-silicate group. Most of the small bodies of amphibolite are cross-cutting lentimular bodies composed essentially of actinolite and plagioclase. Garnet, sphene and opaques are common accessory minerals. Other cross-cutting bodies (e.g., 231-71) are composed essentially of a dark green chlorite ($2V_Z = 15^o$; $\beta = 1.615$; birefringence .006; probably a prochlorite) with accessory biotite, sphene, apatite, quartz and some opaques. Both these types are considered to be metamorphosed basic dykes.

A much larger outcrop of amphibolite occurs in the northern part of Subarea III. This mass was thought to be a series of soda-rich volcanic rocks by Jones, Talbot and McBriar (1962). The rocks are composed essentially of actinolite and albite and accessory epidote. Relic porphyritic and ophitic textures are fairly common in the mass and layering is observed in parts. The volcanic nature of the rocks was suggested by the layering and the presence of structures believed to be pillows. The contact relations with the surrounding schists were thought to be tectonic. However, the contacts are highly irregular in places and marginal alteration of the schists is common along parts of the northern contact. Both the outer margins of the amphibolite and the surrounding schists have been albitised and structures resembling agmatites are observed in parts of the marginal zones of the amphibolites (Fig. 83). It seems more likely that the amphibolite is an intrusive basic igneous rock whose margins have been sheared and albitized.

Pegmatites are extremely common in the Willyama Complex, either as thin interfolial bodies or as larger masses of slightly cross-cutting nature. Individual pegmatites have not been shown on the map; the two large pegmatite masses in the southern part of the area are complexes of pegmatite and minor gneissic or schistose material. The pegmatites are typical of the Olary Province, consisting essentially of quartz, microcline (commonly perthitic), and minor albite with accessory muscovite, tourmaline and rare beryl.

Fig. 83. Blocks of amphibolite (darker grey), in a matrix of fine-grained albite. Agmatite. Subarea III. Plate IV.

FIG 83

CHAPTER XII.

STRUCTURES IN THE WILLYAMA COMPLEX.

The large scale structure of the Willyama Complex is comparatively simple. The structure is dominated by east-northeast trends and variations in these trends are rare (Plate III). The only major departures from this simple structure occur in Subarea VII, where the granite appears to have pushed aside the mica schist, and between Subareas I and III, where the mica schist appears to be faulted against the platy gneisses. At this latter contact the trend of the platy gneisses swings into near conformity with the contact. Another large scale variation in the foliation trend occurs in the extreme southwest part of the area. Here the gneisses show a smooth swing of about 90° in the foliation trend, but alluvium obscures the relationship of this structure to others in the Willyama Complex.

The most prominent mesoscopic structures in the Willyama Complex are foliation and compositional layering. Both of these structures show a more or less well-developed planer preferred orientation parallel to the strike of the boundaries between the major rock units in the region. The layering varies from planar in the platy gneisses to lensoid and irregular in the migmatites. Only rarely is the platy layering not parallel to the local foliation, even where the layering is folded.

The layering is locally folded on a small scale and several styles of folds are common:

1. Folds without a crenulation cleavage parallel to the axial surface. (B_2 folds).
2. Folds with a crenulation cleavage parallel to the axial surface. (B_3 folds).
3. Kink folds with a crenulation cleavage parallel to the axial surface. (B_4 folds).
4. Kink folds without a crenulation cleavage parallel to the axial surface. (B_{4a} folds).

Within these groups variations in style are common and appear to be related to differences in rock type. The styles tend to be mutually exclusive and do not commonly overprint one another. Exceptions are the B_{4a}folds which overprint the other styles in limited areas.

1. <u>B_2 folds.</u>

Folds without a crenulation cleavage parallel to the axial surface are comparatively rare. They are found in both the gneisses and schists and show different characteristics in each rock type. In the gneisses moderate variations in the attitude of the layering may represent a macroscopic variation related to the rare mesoscopic folds. However, the amount of variation is not sufficient to establish an accurate macroscopic fold axis (See Subarea I, Plate V). Mesoscopic folds, though rare, have a fairly consistent fold style. They are always small and generally involve only a limited number of layers. The style is concentric and the limbs are nearly parallel in most examples. Polyclinal folds tend to be present in the less highly appressed examples (Fig. 84). The micas are most commonly oriented parallel to the layering even in the cores of tight folds. Some micas tend to be aligned subparallel to the axial surface of a few folds, especially in mica-rich layers, but a new axial plane foliation is never strongly developed.

Small folds which may be related to the mesoscopic folds in the gneisses occur in mica schists with thin quartz-feldspar laminations. The folds have straight limbs and sharp crests (Fig. 77) and are commonly rootless. Irregular lenses of quartz-rich material may be the transposed remnants of these folds. The foliation associated with these folds is parallel to the axial surface of the folds. However, this does not distinguish them from B_3 folds in mica schists, since these also develop an axial plane foliation in very tight folds (e.g., Fig. 88). The plunge of the B_2 folds in the schists is steep and contrasts with the variable plunge of B_2 folds in the gneiss and with the moderate plunge of B_3 folds. The B_2 folds in schists have been included in this group because of their lack of crenulation cleavage, but they may be related to either B_2 or B_3 folds.

2. B_3 folds.

Folds with a crenulation cleavage parallel to the axial surface are common in limited areas, especially in the zones

Fig. 84. Style of folding in gneiss. Folds are slightly polyclinal.
Spec. 231-24A.

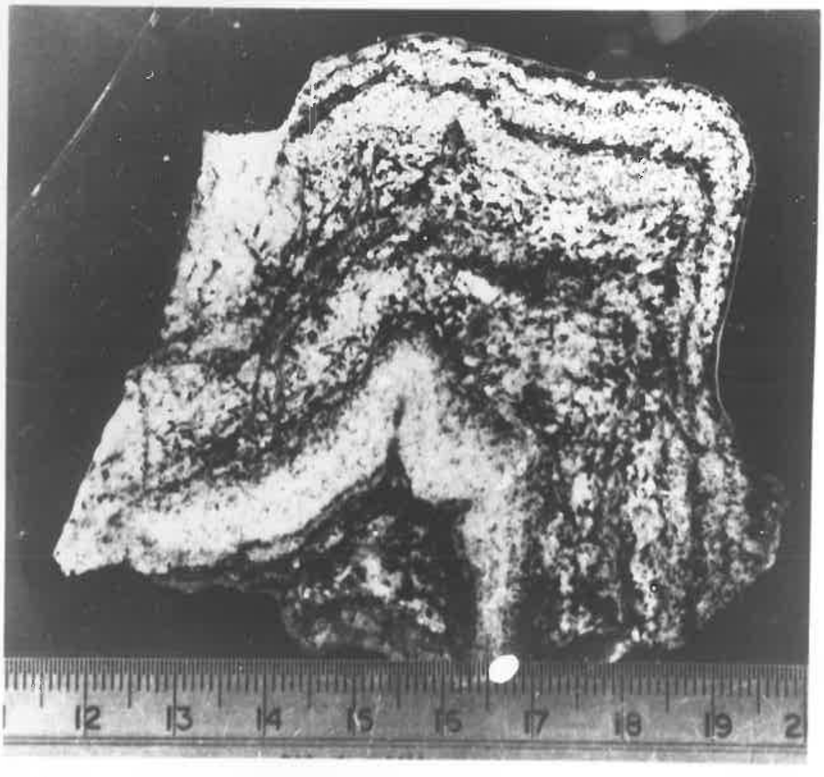


FIG 84

of interlayered schist and platy gneiss. The best localities for these folds are in the transition zone between Subareas I and II (Localities 38, 75 and 76). B_3 folds are also observed in some of the thicker schist bands in the gneisses of Subarea I. The crenulation cleavage is present alone in mica schists with no gneissic layers.

Folds of this type are never very large (most commonly less than one metre amplitude). The mesoscopic style is highly appressed concentric, with rounded crests and long fairly uniform limbs (Fig. 85). Some apparent thinning of the quartz-feldspar-rich layers is seen in the limbs of the folds but is much more marked in the micaceous layers. In most cases the gneissic layers are continuous although in other cases certain parts of the fold may be considerably thinned. Boudinaged gneiss layers were not observed. Thin interfolial pegmatites are commonly folded with the B_3 folds (Fig. 86); no pegmatites were observed to cut B_3 structures; however, since the B_3 folds occur in such a restricted area it does not follow that there are no post-B_3 pegmatites.

Associated with the B_3 folds is a vertical, axial-plane crenulation cleavage (Fig. 86). This cleavage is observed only in the mica-rich layers and has many of the characteristics of the crenulation cleavage in the Houghton area, except that the micas are coarser-grained and hence some of the relationships are more evident. In thin section the cleavage is seen as closely spaced planes which are the long

Fig. 85. Typical B_3 folds in interlayered mica schist and platy gneiss. Oblique profiles. Taken looking east. Subarea II. Locality 75.

FIG. 85A

FIG. 85B

Fig. 86. Transverse profile of B_3 folds in interlayered schist and gneiss. A thin pegmatite is folded in conformity with the layering. A crenulation cleavage (S_3) is parallel to the axial surface of the B_3 folds.
Subarea II. Locality 76.

Fig. 87. Crenulation cleavage (S_3) in mica schist. Quartz is virtually absent from the limbs of the microcrenulations.
Subarea VII. Spec. 231-56.

FIG. 86

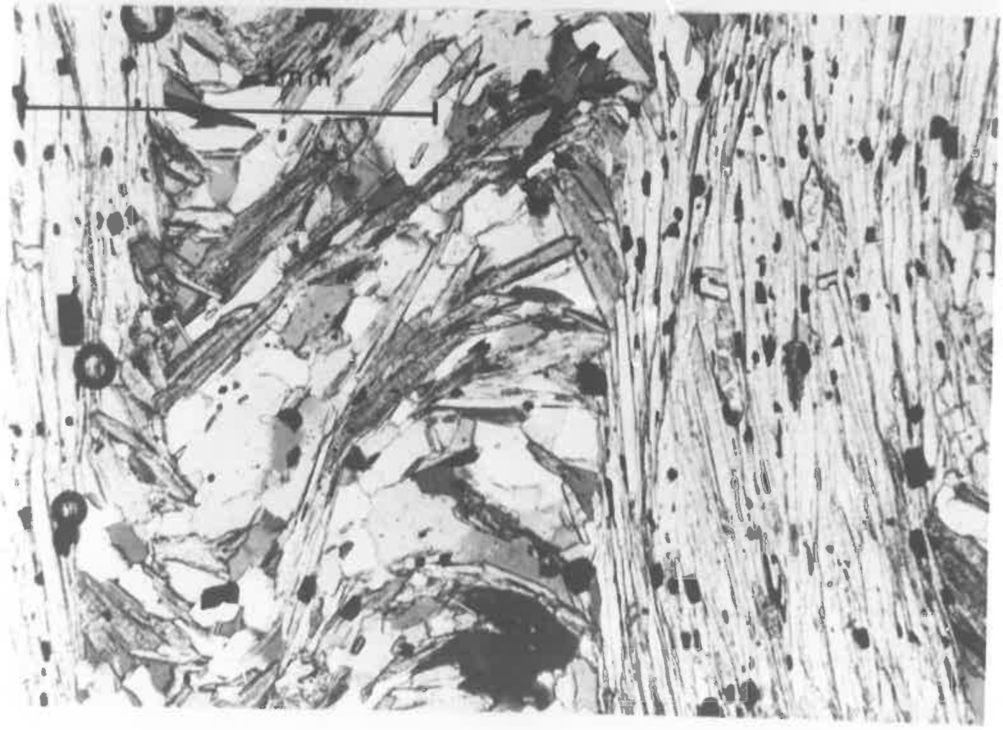


FIG. 87

appressed limbs of microfolds. As at Houghton it is only one of the limbs which is appressed, and quartz and feldspar are obviously excluded from this limb (Fig. 87). However this metamorphic differentiation does not give rise to the pronounced coarse layering seen in the Torrens Gorge, although the rocks may appear laminated. In contrast to the Houghton area, the cleavage planes are only very rarely planes of discrete slip; this can probably be correlated with the higher content of mica in the Weekeroo rocks, and may reflect lower differences in competency between the schist and the quartz-rich layers under the conditions of higher grade metamorphism.

A few specimens of B_3 folds were collected and polished surfaces and thin sections examined. They are commonly more highly appressed than the larger folds (Fig. 88) and the two limbs may be almost in contact with only a few millimetres of schist between them, but the overall concentric nature of the folding is still maintained. In all such folds the micas in the quartz-feldspar layers are oriented parallel to the layering. In the mica-rich layers the orientation of mica is determined by its position in the crenulations, but in the cores of appressed folds the mica has a strong preferred orientation parallel to the axial surface.

In many outcrops more than one crenulation cleavage is observed, although only one crenulation cleavage is axial surface to B_3 folds. The other crenulations are commonly only waves in the foliation surface and have extremely

Fig. 88. Microscopic style of B_3 folds. A good axial plane foliation is developed in the cores of the folds. Spec. 231-75.

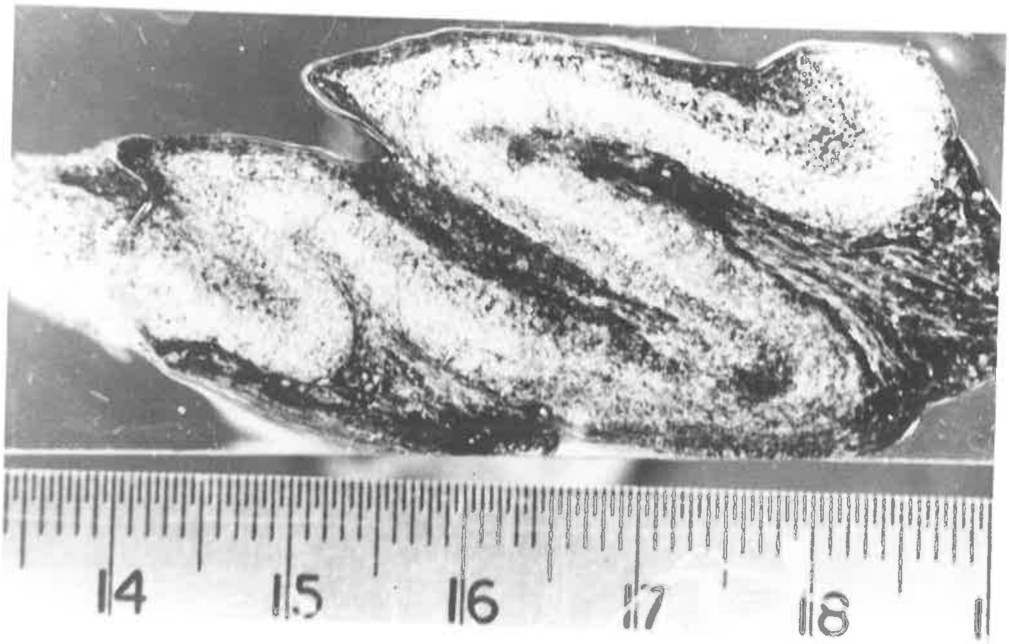


FIG. 88

variable attitudes. It is not known what relationship these have to any larger scale structures.

In all of the samples studied only a few examples of kinked or warped micas were observed. Even in the tightest crenulations the individual mica flakes are unstrained. That the micas are unstrained indicates a period of post-tectonic crystallization at the biotite grade of metamorphism. The rocks of the Torrens Group are also metamorphosed to the biotite grade and it is quite probable that the post-tectonic recrystallisation in the Willyama schists was contemporaneous with the neomineralization of the Torrens Group rocks.

3. B_4 folds.

In the fault block of Willyama rocks north of the amphibolite (Subarea 4) are a number of folds which differ from the B_3 folds. The style and degree of appression vary considerably; commonly the folds have kinked crests, although many have sharply inflected but rounded crests (Fig. 89). The axial surfaces (S_4) dip at moderate angles to the southeast and many individual folds are polyclinal. Some of the folds have a crenulation cleavage parallel to the axial surface.

The attitude, style and restricted occurrence of these folds separates them from the B_2 and B_3 folds. The occurrence of B_4 folds in the vicinity of an important fault suggests they might be related to the fault; however, no comparable signs of deformation are visible in the overlying Torrens

Fig. 89. Kink fold in platy gneiss and schist.
Photo taken looking northeast.
Subarea IV. Locality 48.

Fig. 90. Kink fold in mica schist. The axial surface of the kinks is vertical north-south. View towards the north.
Subarea II.

FIG. 89

FIG. 90

Group rocks.

4. B_{4a} folds.

Kink folds (B_{4a}) with essentially vertical north-south axial surfaces are found in some of the mica schists (Fig.90). They are never tight structures; the angle between the limbs is always less than 40°. They are commonly associated with north-south joints which may act as axial planes to individual kinks. As the axial surface of these folds is essentially perpendicular to the steeply dipping foliation, all the kink axes plunge steeply.

These kink folds overprint crenulation cleavages in mica schists, No overprinting of B_3 folds was observed, as kink folds are not found in the transition gneisses where the B_3 folds are abundant.

In the gneisses and pegmatites very small scale offsets on vertical north-south joints are extremely common. These offsets and the B_4 kinks in schists may be expressions of the same strain and in turn may be related to the large scale north-south faults in the area.

CHAPTER XIII.

UPPER PRECAMBRIAN ROCK TYPES.

Lying with marked unconformity on the Willyama Complex are rocks of presumed Upper Precambrian age. They were correlated with similar rocks in the Adelaide region by Mawson (1912), although in the area investigated by Mawson only the equivalent of the Sturt Group is present. Campana (1956) correlated some of the lower beds in the Plumbago Sheet area with the Torrens Group of Adelaide. Rocks of the Sturt Group are also widespread in the Plumbago area. The general distribution of the Upper Precambrian rocks in the Plumbago-Weekeroo area is shown on Fig. 4 and the distribution of the Torrens Group and Sturt Group in the area mapped on Plate III.

General appearance of the rocks of the Torrens Group and Sturt Group.

The original sedimentary nature of the younger sequence of rocks is readily apparent. The rocks are well-layered and show an abundance of small scale sedimentary features such as ripple marks, cross bedding and convolute laminations, and in the basal beds, rare suncracks. Facies changes in many horizons can be distinguished easily and original variations in the quartzites and glacial horizons are particularly prominent. The macroscopic structures are relatively simple and individual marker horizons can be followed for comparatively large distances.

Many of the original sedimentary features, however, have been profoundly modified by deformation and metamorphism. Campana and King (1958) concluded that the rocks were not regionally metamorphosed but the present author does not agree with this conclusion. The monomineralic rocks such as quartzite and dolomite are simply recrystallized but all the other rocks show considerable neomineralization. Muscovite and biotite occur in all of the "silts" and commonly define a good foliation. Other, rarer metamorphic minerals are tremolite and actinolite. Scapolite is widespread in many of the Sturt Group "silts" near the top of the sequence observed by the author. If this mineralogy is compared with that of the mica schists of the Willyama Complex, it is seen that the Upper Precambrian rocks lack only garnet as a regional mineral (the occurrence of staurolite, chloritoid and sillimanite in the Willyama Complex is local, and they are obviously relic minerals in the rocks in which they occur). It is concluded that the present metamorphic grade of the Willyama Complex and the Upper Precambrian rocks is essentially similar.

The overall simplicity of structure is also misleading. Although mesoscopic fold structures are rare in the Torrens Group, a considerable degree of transposition of thin competent beds has taken place by rotation of small faulted blocks. On a large scale the considerable thinning of sedimentary units on the limbs of macroscopic folds also indicates a considerable amount of deformation.

In this section the basal unit is described separately and the other rocks are discussed only in terms of their importance to the structural history of the area.

1. The basal unit of the Torrens Group.

The contact between the Willyama Complex and the Torrens Group is readily mapped except in the southwestern part of the area where alluvium obscures most of the rocks. The best exposures of the contact are found on the western margin of Subarea A and parts of the western margin of Subarea B.

The rocks lying directly on the unconformity are for the most part metamorphosed conglomerates. In a few localities impure quartzites rest on the Willyama rocks. The conglomerate is quite variable in particle size, composition and thickness, and these variations show a correlation with variations in the underlying Willyama Complex.

The maximum thickness of the conglomerate is about 15 metres. The thickest sections are found adjacent to the schists in the Willyama Complex and even local occurrences of schist in the gneiss unit show a corresponding thickening of the conglomerate at those localities. The pebbles show a corresponding increase of size(up to 10 cm. diameter) in the thicker sections. Adjacent to very massive rocks the conglomerates are thin or may be absent (Fig. 91).

The composition of the basal beds varies sympathetically with the rocks in the underlying Willyama Complex. In all

Fig. 91. Contact between pegmatite of the Willyama Complex (right hand side) and impure quartzite of the Torrens Group.

Subarea A. Locality 102.

Fig. 92. Pegmatitic and quartz debris resting on pegmatite of the Willyama Complex.

FIG. 91

FIG 92

the outcrops pebbles of vein quartz are prominent. Adjacent to the gneisses,pebbles of quartz-feldspar gneiss are very common and pebbles of quartz-feldspar-biotite gneiss are also present. Adjacent to the schists, gneiss pebbles are less common and vein quartz pebbles predominate. As might be expected, schist pebbles are not abundant. In contact with pegmatites, conglomerates are less common (Fig. 91). Pockets of pegmatite debris occur in places against the pegmatite and probably represent depressions on the original surface (Fig. 92). Pebbles of amphibolite and albite rock are found in the conglomerate resting on the amphibolite and in some exposures epidote (from the amphibolite) is abundant in the matrix.

The shapes of the pebbles are quite variable but ratios of long to short axes are not commonly greater than 3:1; more spherical pebbles are more common. The pebbles are rounded to subrounded although some angular pebbles are present. In most outcrops dimensional preferred orientation of the pebbles is only poorly developed and where strongly developed can nearly always be correlated with other fabric elements in the rocks.

Vertical variations in the basal unit are also fairly regular. The incidence of conglomerate decreases with height above the base and the pebbles are smaller (less than 0.5 cm. diameter). Although the composition of the basal bed varies with the Willyama rock type, vein quartz pebbles are always predominant in the higher conglomerates. More commonly however, the higher beds of this unit consist of fairly uniform impure quartzites, metagreywackes and interlayered mica schists. The quartzites commonly show cross bedding, heavy mineral laminations, and ripple marks; suncracks were observed in two localities. These structures are consistent with deposition in a shallow water environment.

In thin section both metamorphic and relic sedimentary features can be observed in the conglomerates. Essentially unaltered clastic grains are abundant in many of the finer-grained rocks (e.g. 231-35) and the pebble outlines in many cases appear little affected. The recognizable clastic grains are predominantly plagioclase (albite) (Fig. 93) and are not modified apart from marginal replacement by a fine-grained quartz-biotite aggregate. Quartz is less common as recognizable clastic grains. This appears inconsistent with the high content of quartz in most of the gneisses and with the great abundance of vein quartz pebbles. However, many polygonal aggregates of fine-grained quartz could be interpreted as clastic grains whose boundaries merge into the matrix. The apparent anomaly could thus be interpreted as the greater tendency of quartz compared with feldspar to recrystallize to a fine-grained aggregate. It is unlikely that all the clastic quartz grains were originally aggregates, as aggregates of pure fine-grained quartz are not common in the Willyama Complex.

Fig. 93. Clastic grain of plagioclase in recrystallised groundmass of conglomerate.
Spec. 231-7.

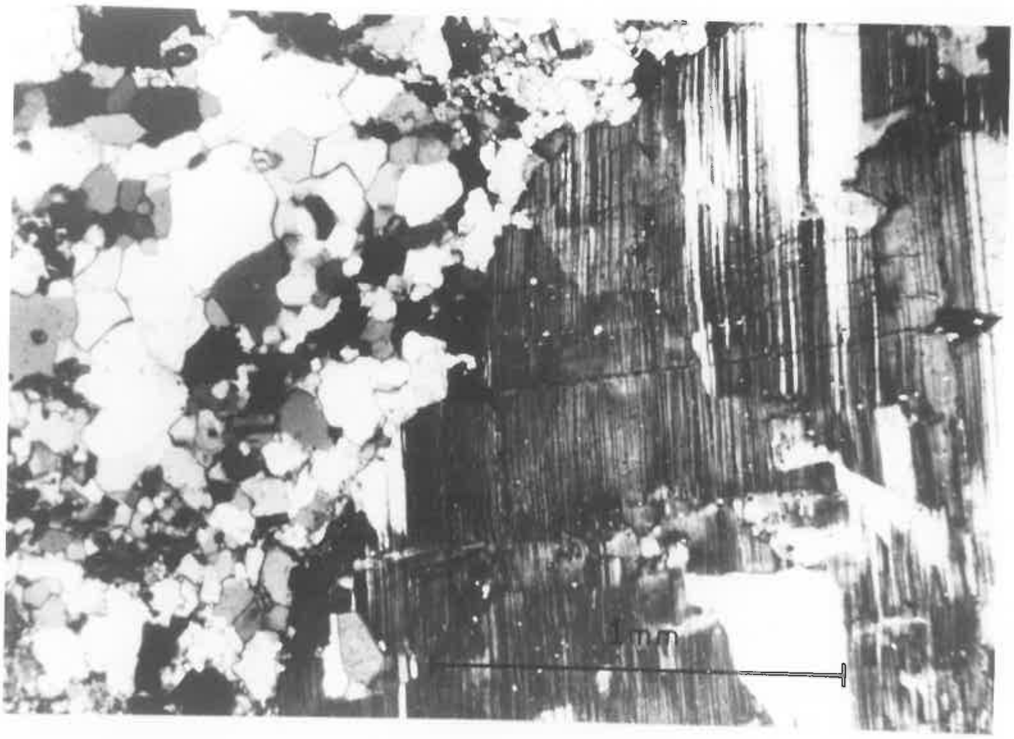


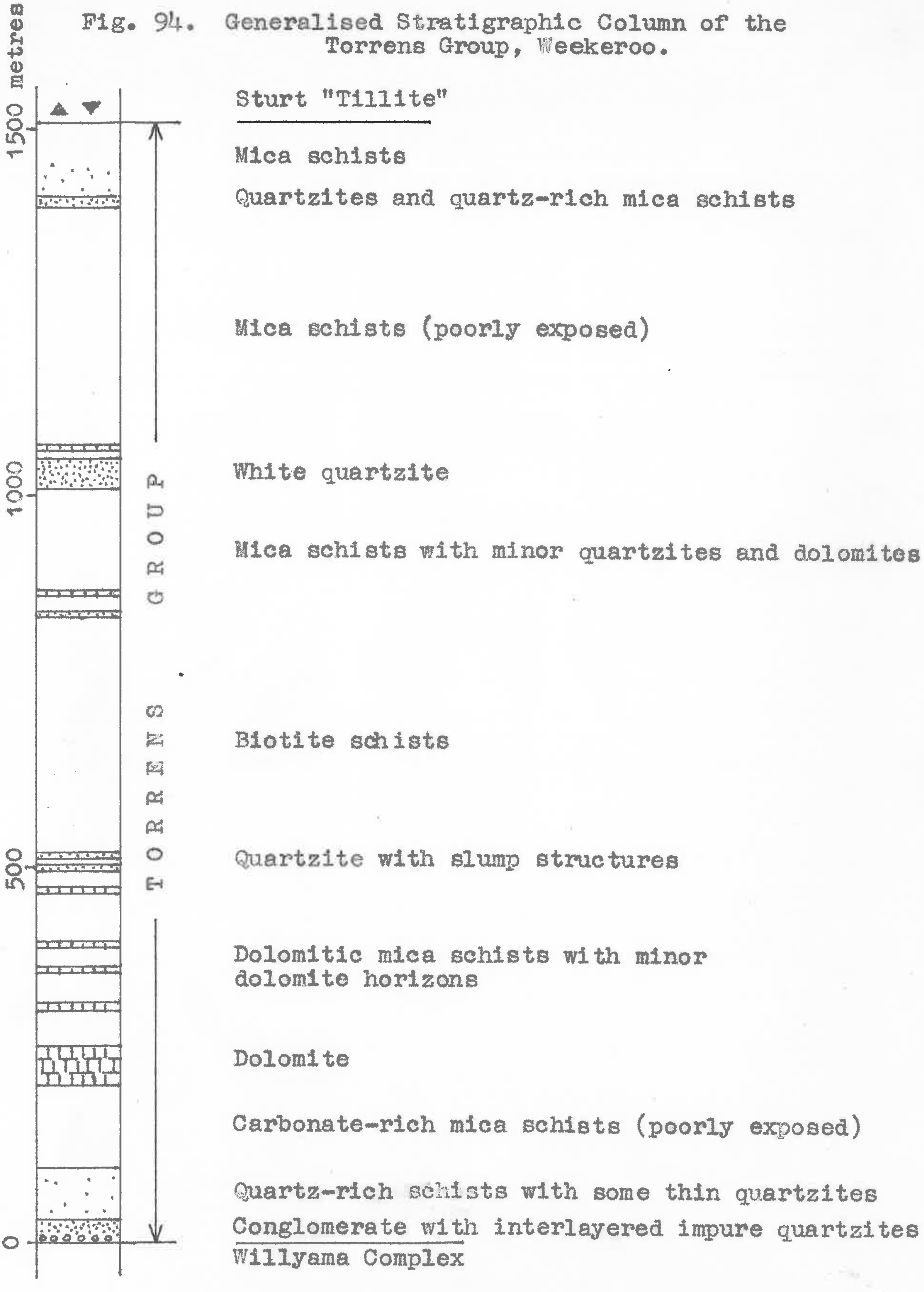
FIG 93

Many other textural features may be interpreted as of metamorphic origin. Biotite is a common mineral in the matrix of the conglomerates and in many cases is present in sufficient quantity to impart a poor foliation to the rock. The matrix of the conglomerates is an aggregate of fine-grained quartz and mica, and the textures resemble those of the fine-grained mica schists higher in the sequence. The grain size of the matrix is variable even within a single thin section varying from about .01 mm. to more than . 2 mm. These variations may reflect differences in original grain size but are also correlated with differences in composition; parts of the rock fairly rich in mica are fine-grained; the coarse-grained portions are pure quartz aggregates.

2. <u>The remainder of the Torrens Group sequence.</u>

The rocks lying above the basal beds consist of a sequence of mica schists, dolomite marbles and pure and impure quartzites. A stratigraphic column showing tentative thicknesses of each unit has been constructed (Fig. 94). Such a column is difficult to construct owing to the considerable changes in apparent thickness associated with deformation. The thicknesses, taken from the transverse profile (Fig.96), were measured in a direction parallel to the axial surface in the crestal regions. This method is open to the objection that the amount of elongation (and hence thickening) which may have taken place in a direction perpendicular to the fold axis is unknown. However, the column probably gives

Fig. 94. Generalised Stratigraphic Column of the Torrens Group, Weekeroo.



a fairly good indication of the relative positions of the more massive mappable horizons.

Most of the section consists of fine-grained mica schists, which vary from rocks which have greater than 50% biotite to rocks which superficially resemble siltstones, but which also contain biotite showing a good preferred orientation. Commonly these rocks are laminated in parts, but many of the rocks appear quite massive. In thin section (e.g., 231-3, 14, 44) they are seen to consist essentially of quartz and brown biotite with less muscovite and rare feldspar. Carbonate is common in some specimens. The grain size of the micas is fairly constant at about .05-.1 by .02 mm. with the quartz grains somewhat smaller, .03 mm., although rising to 0.1mm. in quartz-rich schists. The accessory zircons in the rocks have a smaller grain size (ca. .005 mm.) and may indicate the grain size of the original rocks, which would lie just within the range of siltstones.

Dolomite marbles, both pure and impure, occur in many parts of the sequence and have only been differentiated from the mica schists where they are prominent marker horizons. A dolomite about 60-70 metres thick forms a distinct marker horizon near the base of the sequence. The dolomites are commonly fine-grained, varying in grain size from .02-.10 mm. Quartz and muscovite are common accessories in the grey dolomites (e.g., 231-1 and 2). In the region near the east end of the amphibolite, coarse tremolite occurs in the dolomites (231-77A) and appears to be a metasomatic replacement product of the dolomite.

Quartzites also occur as thin layers throughout the sequence; however, the main quartzite near the middle of the sequence is the only reliable marker quartzite. Other quartzites tend to be irregular and lensoid and are not reliable for correlation purposes over the whole area. In hand specimen many of the quartzites show a crude foliation which in some instances is due to the preferred orientation of micas. In some cases a "foliation" is present in pure quartzites and appears to be due to the elongation of original quartz grains, although in thin section these appear as elongate groups of polygonal grains, the grains within each group having closely similar orientations (231-45 and 46).

3. Rocks of the Sturt Group.

The base of the Sturt Group has been placed arbitrarily at the base of a unit of mica schists which contain abundant large granitic boulders. Above this unit large boulders occur sparsely in all types of rock and have only been indicated on the map where they are persistently developed. Small pockets of schist or dolomite with abundant boulders have not been distinguished on the map. Boulders are rare in the beds above the quartzite outlining the syncline south of the northwest granite. However, no unit in the Sturt Group can be said to be completely free of boulders.

The basal unit of the Sturt Group is marked as "Tillite" on the map. This is the most persistent of the boulder horizons in the Sturt Group and has been interpreted as a lithified boulder clay by Campana and King (1958). The horizon is very variable, containing numerous granite boulders in some parts and being almost free of boulders in other parts. A typical outcrop is shown in Fig. 95. The boulders range in size up to 50 cm. diameter and are most commonly well-rounded, although boulders less than 5 cm. diameter tend to be subangular. The dominant boulders are of massive granitoid rocks; many other rock types are represented; gneiss, quartzite, vein quartz and dolomite are present in many localities. The matrix of the rock is a characteristic dark grey structureless fine-grained biotite granofels (231-78). In thin section the clastic nature of grains down to about 0.1 mm. is still apparent. The grains, mostly of quartz but also of aggregates of dolomite or quartz (and feldspar) vary from rounded to subangular and show a wide variation in grain size. The matrix in which these fragments occur has a grain size varying from 0.02 to 0.05 mm. and consists essentially of quartz, olive-green biotite, and carbonate, with accessory muscovite and granular opaques. This matrix is entirely recrystallized (and neomineralized) and its original nature is therefore no longer discernable. The biotite has only a poor preferred orientation and in most outcrops the rocks do not exhibit any marked schistosity. However in the thick "tillite" in the eastern part of the mapped area, schistosity

Fig. 95. Mica schist with granite boulders.

Basal Sturt Group.

East of Subarea D. Plate IV.

FIG. 95

is well developed in rocks which trend at right angles to the regional foliation.

Campana (1956) shows the rocks in contact with the granite in the northwest also as basal tillite. However, the "tillite" which occurs close to the contact is stratigraphically higher than the basal "tillite" and is better correlated with the boulder horizons just to the north of Locality 78 (in the southern part of Subarea H). The whole of the southern contact of the granite with the Sturt Group is a fault, and the outcrop pattern along the poorly exposed northern contact suggests that this is also a fault. The true stratigraphic relationships are therefore unknown in this area.

An isolated pocket of granite debris is seen at Locality 95 in close proximity to the granite. The outcrop consists of boulders of granite (similar to the adjacent granite) varying in size up to 1.5 metres in diameter. The matrix of the rock is also granitic and the rock appears to be an infilled depression on the original granite surface. The relationship to the underlying granite appears to be similar to that of the smaller pockets of pegmatitic debris lying on the large pegmatite in the south (Fig. 92, near Locality 102). The outcrop is overlain by a coarse conglomerate with rounded boulders up to 20 cm. in diameter set in a finer grained matrix. Alluvium separates these outcrops from the Sturt Group outcrops just to the east. Campana and King (1958) have interpreted very similar boulder beds at

Old Boolcoomata Homestead (32 miles east-northeast of the Weekeroo occurrence) as basal tillite horizons and the map symbol used for the Weekeroo occurrence is also for "basal tillite". The present author believes that these granitic debris outcrops are fossil residual deposits and are not necessarily glacial in origin.

The majority of the rocks of the Sturt Group are fine-grained quartz-rich mica schists with minor dolomite marbles and quartzites. Scapolite is common as porphyroblasts up to 0.5 cm. in diameter in some horizons in the mica schists in the northern part of the area (231-13). Scattered boulders (most commonly granite) up to one metre in diameter occur in all parts of the sequence but are most common below the major quartzite which outlines the syncline in Subarea H. Boulders are common also just above this quartzite but are rare higher in the sequence. Sedimentary structures, such as convolute laminations and small scale cross bedding, are common in the quartz-rich mica schists.

The uppermost beds of the Sturt Group in the area mapped are a sequence of massive quartzites. These are medium-grained and the clastic nature of many of the grains is readily apparent (231-97). Quartz, feldspar and quartz-feldspar aggregates are the most abundant recognizable clastic grains[1] and the matrix is a granoblastic aggregate of quartz, reddish-brown biotite and carbonate.

1. Quite a number of grains consist of an aggregate of interlocking plagioclase lathes in random orientation. They appear to be foreign to the area.

CHAPTER XIV.

STRUCTURES IN THE TORRENS AND STURT GROUPS.

The overall distribution of Upper Precambrian rock types indicates a relatively simple structural pattern as Campana has shown (1958). The structure is dominated by a broad synclinal region trending west-southwest to east-northeast, between the southern and northern outcrops of the Willyama Complex. Corresponding anticlinal regions are dominant at the eastern margins of the older rocks. The structure is complicated by many macroscopic folds of a smaller scale than these and also by the presence of many large faults. These various structural elements are all thought to be intimately related but will be discussed separately before any analysis of their relationships is attempted.

The various rock types of the Upper Precambrian show rather different styles of deformation and each rock type will be discussed separately. Particular attention has been paid to the deformation of the basal conglomerate unit, and an intimate relation between the structures in the conglomerate and those in the directly underlying Willyama rocks can be demonstrated.

Structures in the Basal Conglomerate.

It was pointed out in the previous chapter that variations in the thickness and composition of the conglomerate can be correlated with the underlying rock type of the

Willyama Complex. The correlation between the degree of deformation and underlying rock types is also striking, particularly where there is a marked change in lithology in the Willyama rocks.

On a macroscopic scale the structure of the conglomerate is shown on the map (Plate III) and the transverse profile (Fig. 96). Adjacent to the gneiss unit in the south (Subarea I of the Willyama Complex), the outcrop pattern of the conglomerate is relatively straight and strikes about northwest-southeast and dips 60° to the northeast. A few small structures are observed in this region but will be discussed later. At the contact with the Weekeroo Schists (Subarea II) the strike of the conglomerate changes sharply to northeast-southwest, and then outlines a large anticline around the schists.

To the west the conglomerate occurs in a very appressed syncline in which the conglomerates on either limb are almost in contact. The limbs are exceedingly straight and are essentially parallel to the foliation in the immediately adjacent schist outcrops. The conglomerate then delineates a series of faulted folds at the eastern end of the amphibolite, before swinging again into conformity of strike with the foliation in the schists. A series of tight folds are observed before the conglomerate finally abuts against the north-south fault system in the west. Just to the north

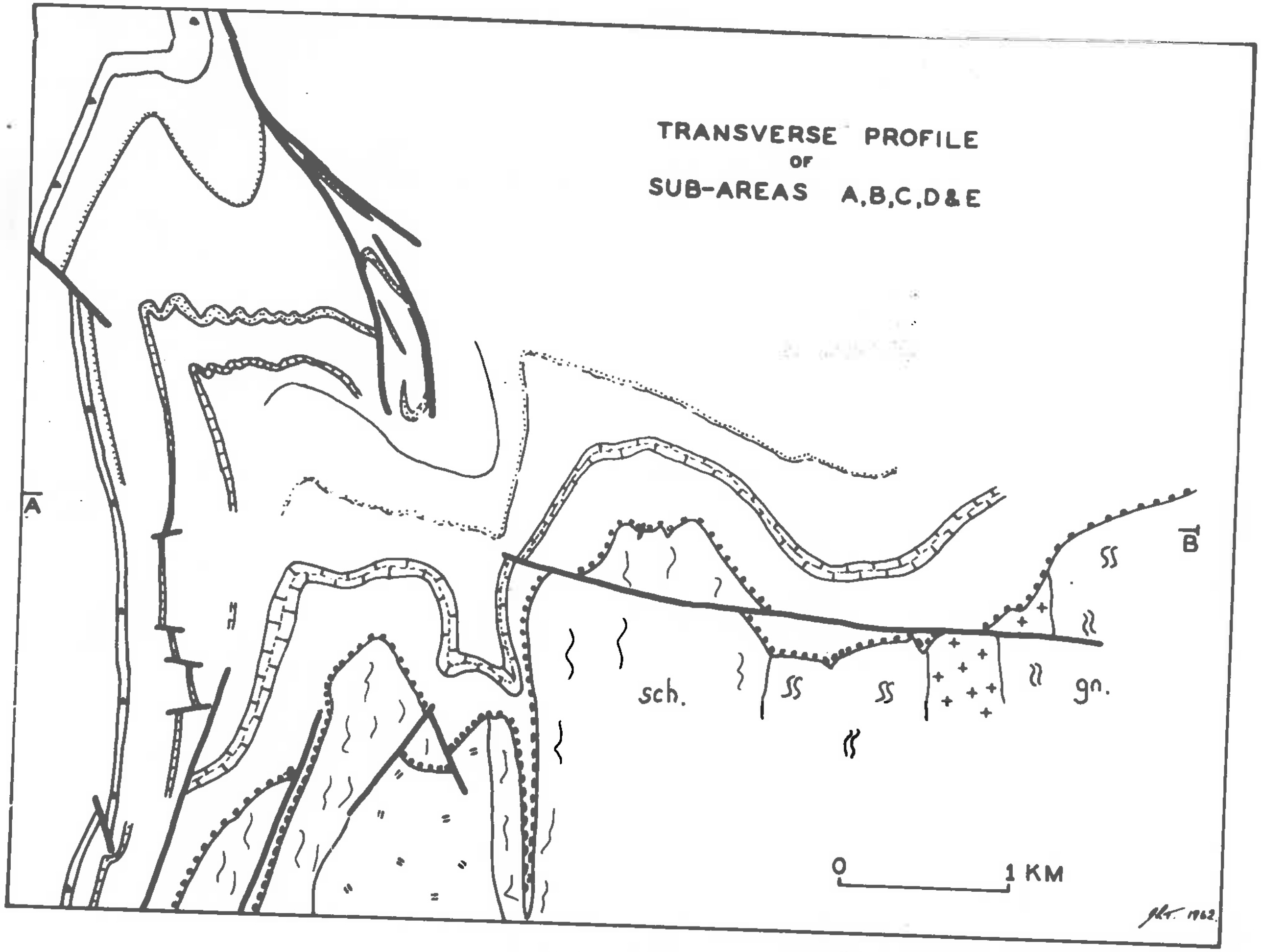


FIG 96

the conglomerate is repeated in a fault block (Subarea IV).

In the southwestern corner of the area the conglomerate reappears and can be followed with difficulty through the alluvium. Paucity of outcrop renders correlation of deformation of the conglomerate to the underlying rock type virtually impossible. However the greater complexity of folds in the extreme south can be correlated with a general increase in the amount of schistose material in the platy gneisses there.

The mesoscopic structures in the conglomerate also show sympathetic variations with rock types of the Willyama Complex. Where the conglomerate rests on massive units of the Willyama Complex it appears relatively little deformed although the matrix is completely recrystallized. The pebbles show no marked tendency for parallel orientation.

Adjacent to some of the more micaceous rocks, the conglomerate commonly shows a weak foliation and in some cases the pebbles tend to line up with their long axes parallel to the foliation even where they retain their original sedimentary shapes[1] (Fig. 97).

Where there are large differences in competency in adjacent rock units in the Willyama Complex the deformation of the conglomerate is more pronounced. Good examples of

1. Flinn(1956) notes a similar tendency for the long axes of undeformed pebbles to show a preferred orientation in the early stages of the deformation of the Funzie conglomerate.

Fig. 97. Conglomerate. The long axes of the pebbles have a preferred orientation parallel to the foliation (Hammer lying parallel to trace of foliation).

Subarea A. Plate IV.

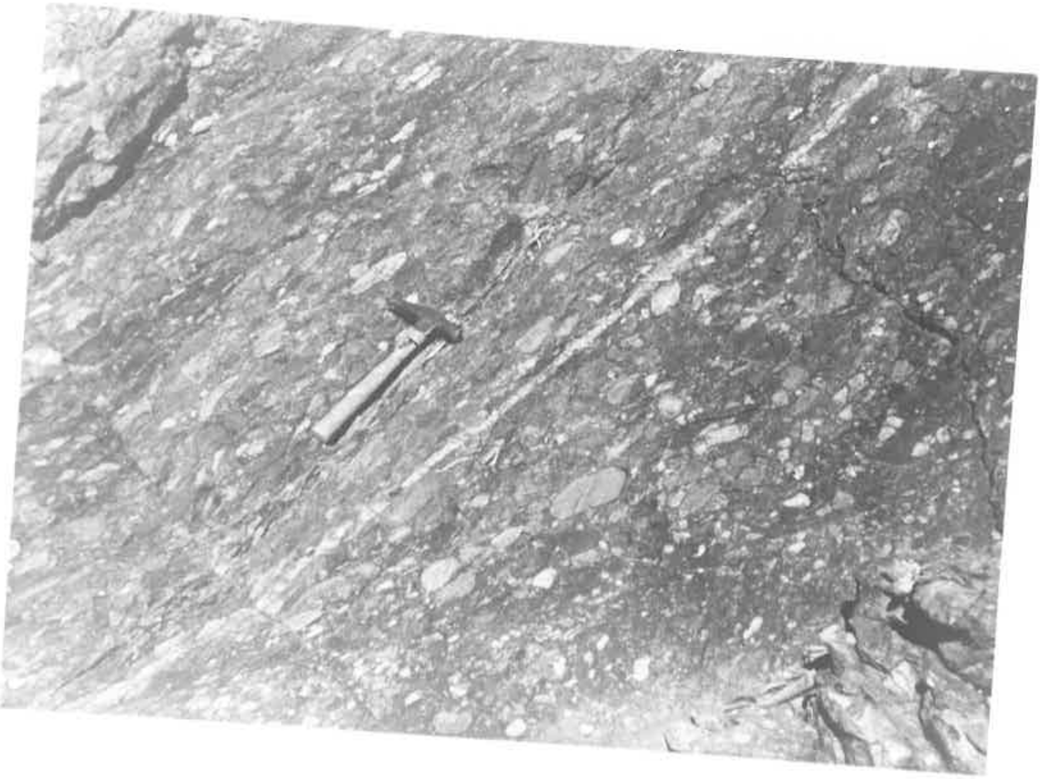


FIG. 97

such deformed conglomerate are found adjacent to the gneiss unit (Subarea I) at Localities 28 and 29. At both of these localities the conglomerate horizon is folded into a syncline with a very sharp axial region. In both localities the northern limb of the syncline is longer than the southern limb, and the axial regions of the synclines are close to the southern border of the mica schists (Fig. 98A)(at Location 28 the southern limb of the syncline is very insignificant). This gives an overall right-hand sense of movement in the conglomerate.(Fig. 98A). Crenulations in the underlying schist also indicate a right-handed displacement. It should be noted that the conglomerate is not folded on a small scale; the northern limb is remarkably straight. At the northern contact of the mica schists the conglomerate swings back to its normal northwest strike. (It is not possible to follow these folds very far into the Torrens Group as the beds above the basal conglomerate and quartzites are poorly-exposed mica schists).

In these synclinal regions (especially Locality 29) the pebbles in the conglomerate are highly elongate. The symmetry of the pebbles is orthorhombic although some pure quartzite pebbles are nearly spindle shaped. Measurements on the pebbles (Loc. 29)[1] are shown in Table IV.

1. Locality 29 is the only locality at which all three axes could be measured. Greater differences in the lengths of the axes are seen in other localities.

TABLE V.

MEASUREMENTS ON ELONGATE PEBBLES. (LOCALITY 29).

Measurements in centimetres.

long	Medium	short axis.	z		y		x	y:x
10	3.5	0.6	2.8	:	1	:	0.17	5.8:1
7	2.3	0.3	3.0	:	1	:	0.13	7.7:1
8	1.7	0.3	4.7	:	1	:	0.18	5.9:1
12	3.0	1.0	4	:	1	:	0.33	3 :1
17	1.6	0.3	11	:	1	:	0.19	5.3:1
19	4	0.8	4.8	:	1	:	0.2	5 :1
16	2.7	1.8	5.9	:	1	:	0.67	1.5:1

Fig. 98. A. Relationships between the relative senses of movement in the conglomerate and the crenulated mica schists. Diagrammatic representation of Locality 29.

B. Relationships of folding in the Torrens Group to the fractures in the underlying pegmatite. Locality 102.

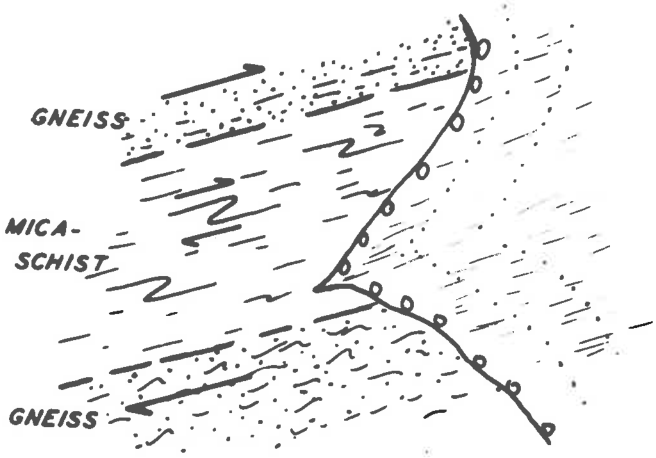


FIG 98A

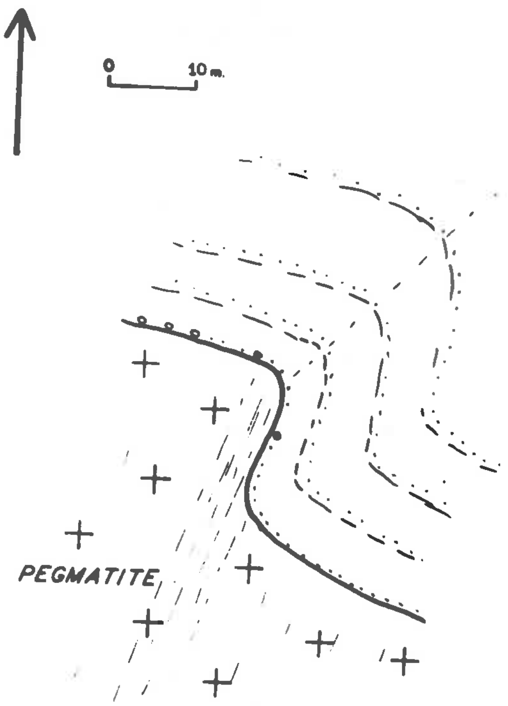


FIG 98B

The short axes x are perpendicular to the foliation. In only a few outcrops can an accurate relationship between bedding and elongation of pebbles be determined, and in these the pebbles plunge more steeply than the bedding-cleavage intersection (Fig. 99).

A somewhat different relationship is illustrated at the contact of the large pegmatite with the conglomerate. The pegmatite is cut by a series of north-south fractures possibly related to the large scale north-south faults of the region. At one point on the contact these fractures are more pronounced and have resulted in a total displacement of about 10 metres in the pegmatite with a corresponding fold in the overlying conglomerate (Fig. 98B). The sense of movement is right-handed and agrees with the movement on the larger fault to the west of Locality 102. The fold in the conglomerate is reflected in folding in the overlying sandstone. The fold appears to be concentric in style as the curvature decreases upwards.

North of the gneisses of Subarea I the Willyama rocks are dominantly mica schist (Subarea II). The contact between the two units is gradational and at this contact the strike of the conglomerate swings to the northeast and continues in this direction for about 700 metres. In contrast with the very tight folds described above the conglomerate appears relatively little deformed. The pebbles do not show any marked elongation and in most outcrops even parallel

Fig. 99. Deformed conglomerate. The pebble elongations (L) plunge more steeply than bedding cleavage intersections (B).

Subarea A. Locality 29 and 4.

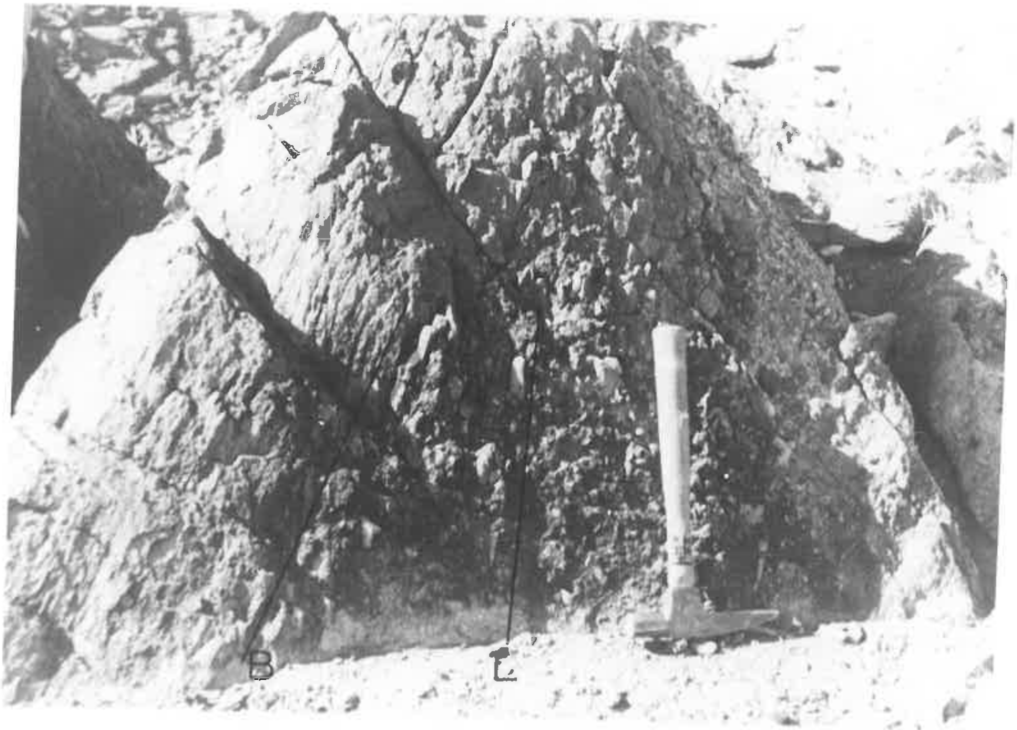


FIG. 99

orientation of the original long axes is not apparent. The conglomerate, however, commonly shows a crude foliation which is parallel to that of the underlying schists.

To the north (near Locality 33) the conglomerate outlines a broad anticline complicated by a series of small folds and faults in the crestal region. The faults are parallel to the foliation, and there appears to be no obvious reason for the localization of the faults. Similarly there appears to be no obvious reason for the localization of the large appressed syncline of Localities 35 and 36. The conglomerate in this syncline is highly deformed (Spec. 231-36). The long axes of the pebbles plunge steeply as in Locality 29, but the plunge on the syncline could not be determined. (This plunge may be quite different from that of the adjacent anticlines which plunge at 50° to 60° to the east-northeast). The long limbs of the syncline are essentially parallel to the foliation in the closely adjacent schists. It must be emphasized that over the whole length of the large anticline and the tight syncline there is no significant change in the attitude of the foliation in the Weekeroo Schists.

The conglomerate then outlines another anticlinal region even more broken than the anticlinal region to the south; the extent of this faulting appears to be correlated with presence of the underlying more massive amphibolite. The conglomerate is largely undeformed. Another very appressed synclinal limb is observed north of the amphibolite and shows similar relationships to the syncline south of the amphibolite. The conglomerate is highly deformed (231-104) and the gross layering in the conglomerate (here parallel to the foliation in the conglomerate) is parallel to the foliation in the Weekeroo schists. The pitch of the elongation of the pebbles is essentially 90°(i.e., it plunges steeply to the south, but outcrops which are not slumped about an axis parallel to the strike of the bedding are rare, so plunge measurements cannot usually be made.). In sharp contrast to the highly deformed limbs, the core of the syncline shows almost no deformation of the pebbles. Although it was not possible to measure the plunge of this fold, a measurement on bedding in the core gave a strike of 163° and a dip of 18° to the east. As noted before, the long axes of pebbles in the adjacent limbs plunge very steeply.

The northern limb of the syncline is cut by a series of strike faults and the conglomerate appears again as a highly deformed north dipping limb (Fig. 100) which is finally cut out by a large north-south fault.

Another outcrop of relatively undeformed conglomerate (231-100) occurs north of the syncline and is separated from the other conglomerates by a fault. Its position as a limb structure seems anomalous with the undeformed nature of the pebbles. Possibly the presence of two large faults bounding the block in which the conglomerate is found is in

Fig. 100. Deformed conglomerate. View looking down the long axes of pebbles. The intermediate axis lies in the plane of the foliation. Subarea III. Locality 47.

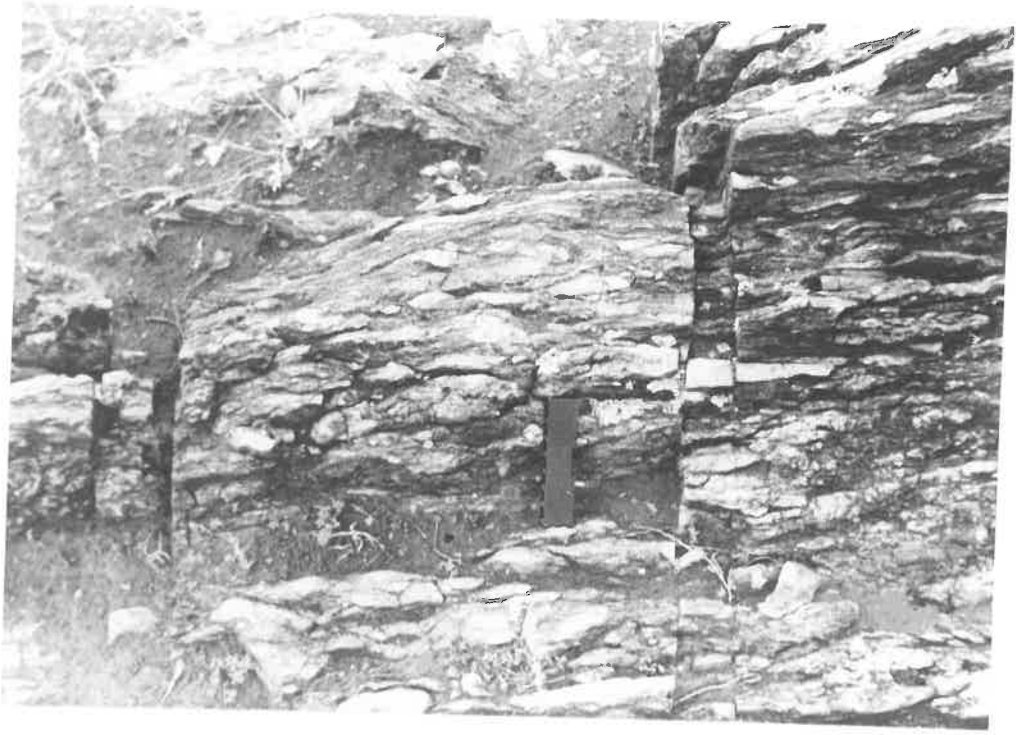


FIG 100

some way responsible for this apparent anomaly.

In thin section the relatively undeformed conglomerates (231-100) show abundant clastic pebbles of vein-quartz, microcline (some single large grains up to 5 mm.) and quartz-feldspar aggregates of various kinds which can be matched readily with rock types from the Willyama Complex. These pebbles are in a matrix of clastic grains of quartz, microcline, albite, mica and granular opaques. The smallest recognizable clastic grains are about 0.2 mm. in diameter and these in turn lie in a recrystallized matrix of quartz, feldspar and variable amounts of mica. Many of the opaque granules are euhedral and are possibly recrystallized.

In rocks richer in mica (231-24E) small clastic grains are still recognizable (Fig. 101) but the recrystallized matrix shows a good preferred orientation of the micas, resulting in a foliation which appears to "flow" round the clastic grains leaving the latter as augen.

In the conglomerates which show a fair degree of deformation of the pebbles (e.g., 231-35 and 36) the individual pebbles are still distinguishable from the matrix although the vein-quartz pebbles are highly strained or composed of polygonal aggregates of small quartz grains. Some parts of the matrix may be quartz-rich and the boundaries between pebbles and matrix are diffuse. Clastic sand-size grains are less abundant, and feldspars are the only commonly recognizable clastic grains. The clastic feldspar grains are

Fig. 101. Deformed vein quartz pebble in recrystallised groundmass. Conglomerate. Spec. 231-24E.

Fig. 102. Chequer albite grains in conglomerate.
Spec. 231-6

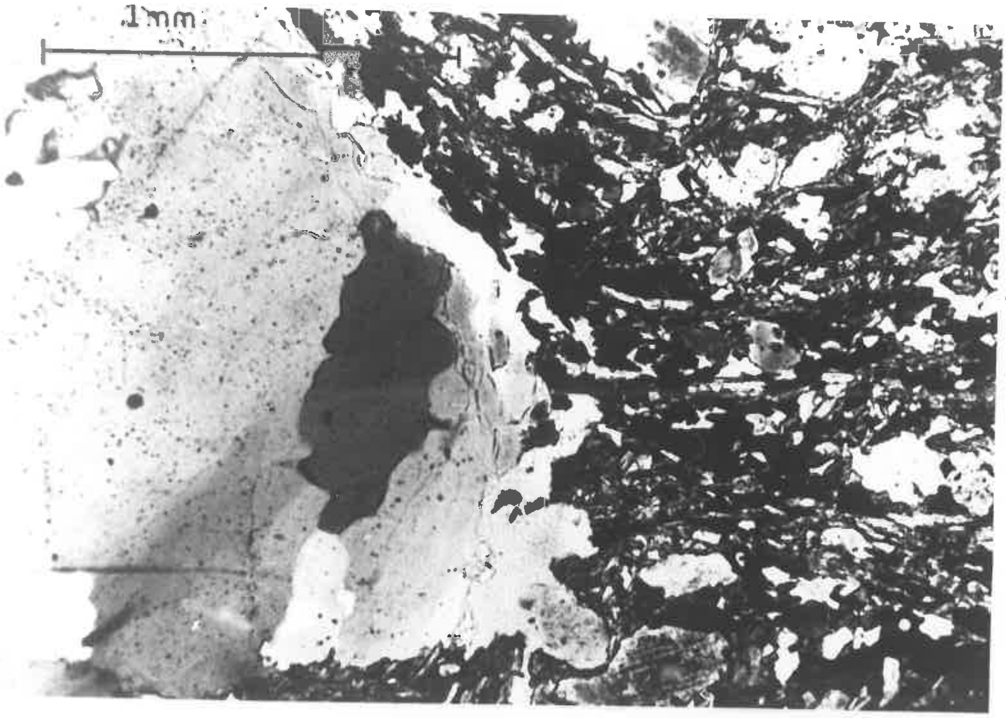


FIG. 101

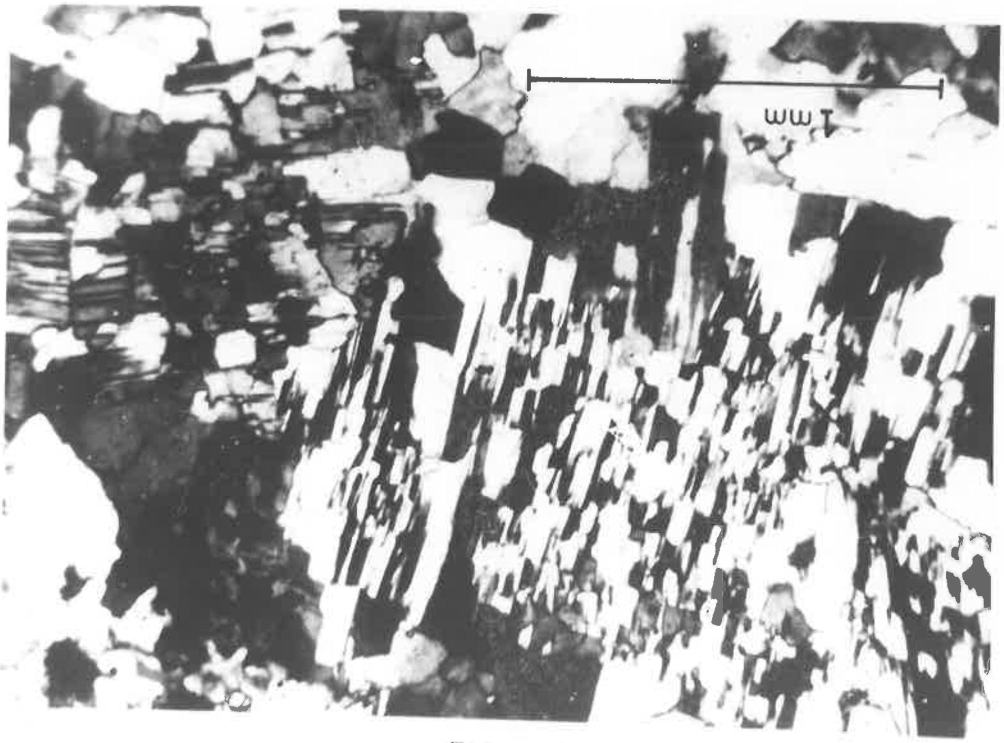


FIG. 102

commonly deformed; the microcline shows crinkled twinning, the plagioclase shows strain bands and bent lamellae and in some specimens chequer albite is common (Fig. 102).

In highly deformed conglomerates(231-104) the original clastic nature of the grains is not readily apparent. A few relatively undeformed grains are present but all other constituents are highly deformed and recrystallized. The "pebbles" are highly elongate bodies of polygonal quartz; the individual quartz grains are not noticeably elongate parallel to the foliation or lineation. The boundaries of "pebbles" are commonly indistinct and are marked only by changes in grain size. (Fig. 103). In other examples the matrix has a composition different from that of the pebbles, but it is not always completely certain whether changes in composition across the foliation-layering are changes due to the presence of pebbles or simply changes in the matrix itself.

<u>Structures in the Torrens Group above the conglomerate.</u>

The folds in the units above the conglomerate reflect the overall folding of the conglomerate unit although the minor irregularities observed in the conglomerate cannot be traced into higher beds. This dying out of minor irregularities can be correlated with the presence of a thick sequence of mica schists above the basal unit.

Mesoscopic folds are rather rare in quartzites and dolomites of the Torrens Group. Where observed they have axes consistent with the macroscopic axes. Small folds in

Fig. 103. Deformed quartz pebble (coarser grained) in contact with quartzite matrix. Conglomerate. Spec. 231-104.

Fig. 104. Layering in dolomite (S) showing transposition into S_2. Subarea A. Locality 1.

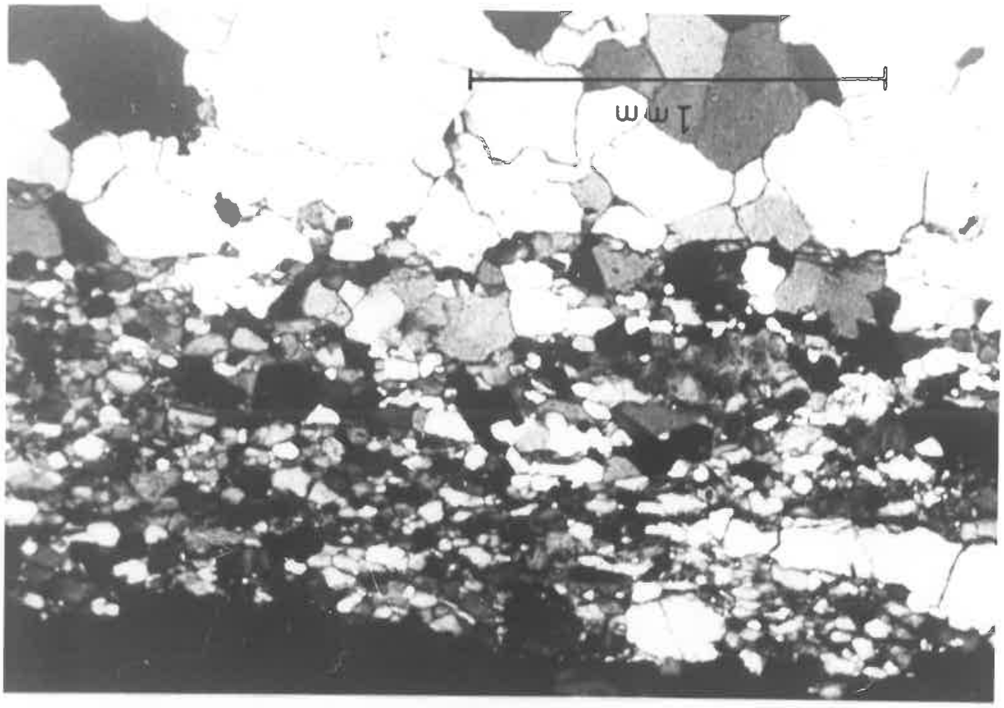


FIG. 103

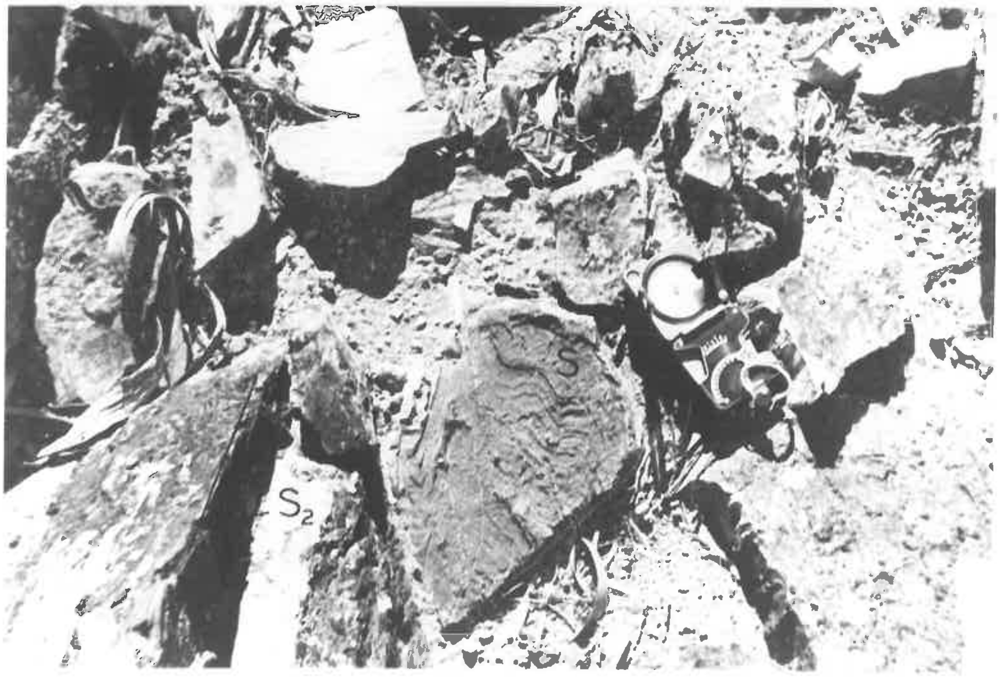


FIG. 104

mica schists have more variable axes. The style of the folds varies from concentric in the massive sandstones to similar in the mica schists.

Although mesoscopic folds are rare, evidence for deformation is abundant. Thin dolomite bands appear to be well and evenly layered in parts but with much contorted bedding in others. In some localities (e.g. some small folds around Locality 1) a transition between these two types is observed. The "bedding" is extremely regular in most parts of the folds, and there is a good lineation parallel to the fold axis, but in a few places small patches of dolomite show a more irregular lamination which is seen to be the untransposed remnants of the original bedding (Fig. 104). The evenly layered "bedding" is therefore interpreted as a transposition structure.

Evidence for transposition is also seen in many of the thin impure quartzites in the sequence. These may show relatively small scale folding but more often the outcrop shows a regular small scale faulting, and the individual quartzite bands are at an angle to the overall outcrop trend (The effect is pronounced enough to be represented on the map in the eastern "sleeve" west of Locality 14). Where the contacts with the adjacent schists are observed, the layering in the quartzite is commonly subparallel to the foliation. (Better examples of this are seen in the Sturt Group and are described in the next section).

The thick quartzite in the upper third of the sequence is folded into a series of spectacular folds in the vicinity of Localities 45 and 46 (Fig. 105). The folds show a classical concentric pattern with the limits of the folds defined by the thickness of the quartzite. Measurements along the top and bottom of the quartzite along its folded part indicate a shortening of 45% perpendicular to the axial surface of the folds, assuming no flattening of the quartzite as a whole. However, the foliation which is developed in parts of the quartzite suggests considerable flattening of individual grains.

The mica schists which form the major part of the Torrens Group sequence show very few structural features other than a well-developed schistosity in the mica-rich varieties. Laminations are fairly abundant and appear clearly on many foliation surfaces as bedding-cleavage intersections. The laminations are not so obvious on other surfaces and in many cases are shredded out in the foliation. The thicker competent beds in the schist show simpler fold forms. Another difference noted in the mica schists is that the bedding-cleavage intersections are more variable in a small area than the axes of folds in more competent horizons.

In a restricted region (the region between Localities 17 and 19) a crenulation cleavage is seen cutting the earlier structures. Well-developed chevron folds in S_2 are developed at Locality 19 (Fig. 106).

Fig. 105. Concentric folds in quartzite.
View from the north. Folds plunge
at 45° to the east.
Folds are those of localities 45
and 46 in Subarea C.

FIG 105

Fig. 106. Chevron folds in S_2 with a crenulation cleavage (S_3) parallel to the axial surface.
East of Subarea C. Locality 19.

Fig. 107. Crenulations in biotite schist.
Spec. 231-19J from same locality as Fig. 106.

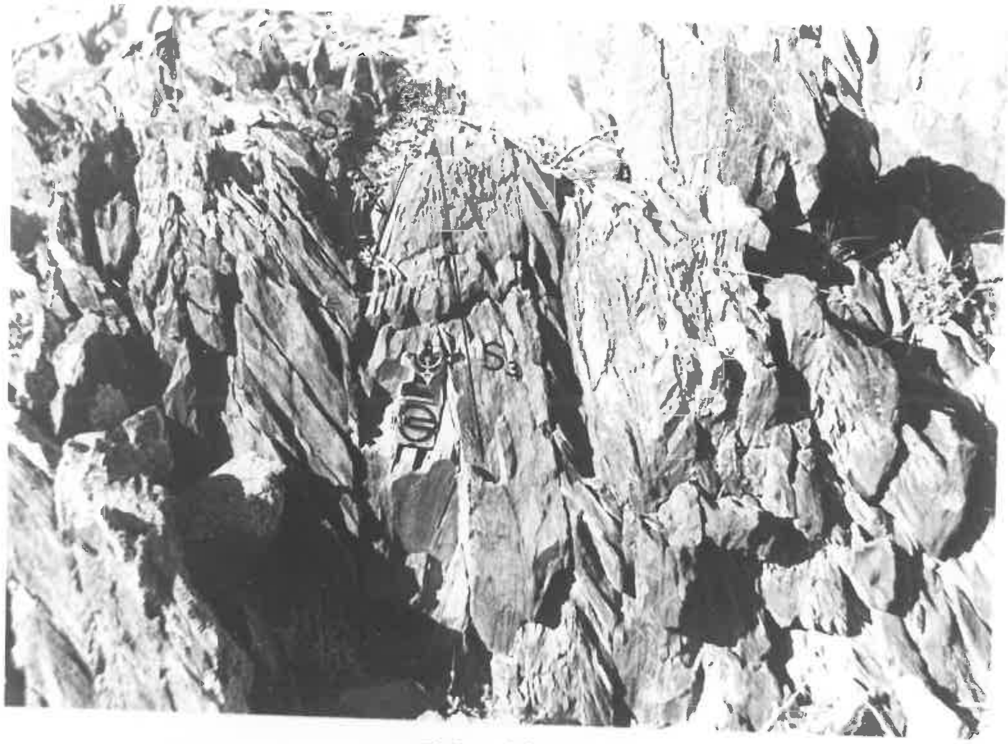


FIG. 106

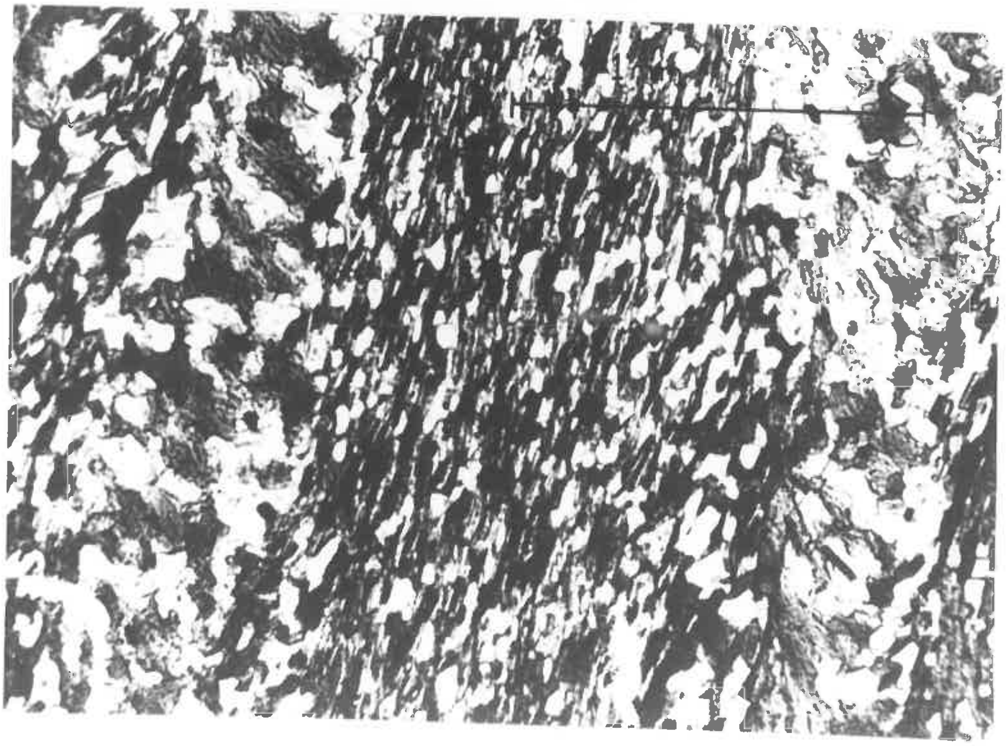


FIG. 107

These folds commonly have a symmetrical chevron style, the two limbs making an external angle of up to 130° with each other (Fig. 106). The structures overprint S_2 structures and in some cases bedding-(S_2) cleavage intersections occur and are folded around the chevrons; the angle betwwen the folded lineation and the axis of the chevron is rarely greater than 30°.

The rocks in which the chevrons are developed are biotite schists with 60-75% biotite (and minor muscovite), quartz and rare feldspar. Intergranular opaque minerals are common accessories. The chevrons develop as the mesoscopic expression of microcrenulations which develop most strongly in certain zones. Only rarely are the crenulations very appressed and then the micas have a strong preferred orientation parallel to the new cleavage and have a smaller grain size (Fig. 107). The folds have remarkably regular limbs for folds associated with crenulation cleavage; this may be a function of the lack of any competent layers in the sequence.

Owing to the very limited area of these folds, their significance in terms of the regional structure can not be determined. However it appears significant that the only outcrops of B_3 folds are adjacent to a zone of rather complex faulting and a genetic relation is suggested.

<u>Structures in the Sturt Group.</u>

No different styles of deformation are observed in the Sturt Group but certain features are better displayed in these rocks than in the Torrens Group rocks. Some apparent differences in deformation arise from the general lower plunge in the Sturt Group, and because the Sturt Group occurs more on the limbs of large structures whereas much of the Torrens Group is in the crestal regions of macrofolds. However this author has observed no structures or stratigraphical relationships which suggest that the Sturt Group lies unconformably on the Torrens Group, as suggested by Campana and King (1958).

The basal "tillite" shows almost no small scale structures. Over much of its outcrop it is on the limb of the major central synclinal structure and exhibits almost no variation in attitude except where affected by faults. In thin lensoid interbeds of impure quartzites the foliation is parallel to the layering in most areas. The only major series of fold structures affecting the "tillite" are at the eastern end of its outcrop. Here the folding of the "tillite" is in conformity with that of the underlying Torrens Group.

At the western end of these folds the plunge on the structure is about 45-50° (similar to the plunge of 44° of the Torrens Group in Subareas B, C and D). The fold is simple and is outlined by thin quartzites at the top of the "tillite". The base of the "tillite" is more difficult to follow, as the underlying beds have a similar composition but lack erratics. Just east of Locality 22 the foliation in the mica schists and tillite is strong and commonly shreds out the contact. Furthermore the schistosity in the "tillite" is differantially developed in certain zones and gives a pronounced large-scale layering to the rock. On the top of the hill formed by the "tillite" differential erosion of the schistose and non-schistose layers results in an outcrop pattern dominated by the schistosity. However, the structure of the true layering is revealed by the quartzite layers and is also apparent on the aerial photographs or when viewed from a distance in the field.

To the east of this east-plunging structure the structure changes in plunge to about 15° to the west, giving rise to a basin structure in some of the beds slightly higher than the "tillite". Then the tillite is folded sharply (near Locality 23) and most of the tillite is faulted out in the core of this structure. The "tillite" section is finally cut out by the major north-south fault to the east.

Another feature of the deformation of the Sturt Group is the presence of numerous small scale folds in many of the quartzites. These folds, which are quite regular in their development are never larger than a few metres in amplitude (Fig. 108) and would be termed parasitic folds by de Sitter, (1958). The folds are seldom large enough to represent on the scale of the map and do not affect the overall outcrop pattern in any significant way. On most of the limbs of large folds the folded surface in the small

Fig. 108. Small folds in quartzites. View looking east.
Northern part of Plate IV.

Fig. 109. Transposed layering in quartzite.
Subarea H. Plate IV.

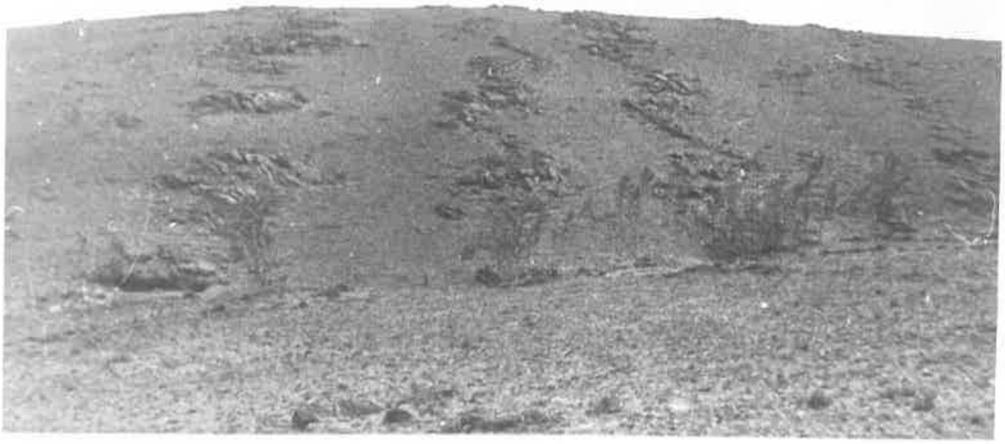


FIG. 108

FIG. 109

folds is continuous. However, in the crestal regions of the large folds, transposition structures are found in the thicker bands of quartzite. The quartzites then appear as a series of en echelon lenses subparallel to the foliation, the lenses arranged in a very regular pattern; the enveloping surfaces outline a smooth outcrop pattern (Fig. 109). In a few localities the generally massive quartzites show good compositional layering which appears to be bedding. This bedding is parallel to the length of the blocks (Fig. 110) and mica schist occurs between the quartzites.

Only the thicker quartzite layers have been deformed on this pattern; the thinner quartzites tend to show small folds similar to those of Fig. 108. In either case the overall result is that the long limbs of the small structures are closely parallel to the local foliation, and the structure seems to be a compromise between the constraint of the layering attitude as a whole (represented by the enveloping surface) and the tendency for the individual quartzite layers to be transposed parallel to the foliation.

This style of deformation is also common in many of the other thin quartzites in the Torrens and Sturt Groups, and these structures emphasize the role of transposition in the deformation of rocks which apparently show very broad simple structures. Transposition on a large scale may also be important in some parts of the area. In Subarea E, the northern part of Subarea J and the southern part of Subarea H, the layering and foliation are commonly parallel, and the

Fig. 110. Transposed layering in quartzite.
The layering (parallel to the hammer)
is approximately at right angles to
the enveloping surface.
Northern part of Plate IV.

FIG. 110

smooth straight outcrops may reflect a large scale transposition of the layering.

Faults.

Faults play a major part in the structure of the region and in many places have modified the fold structures to a considerable degree. It is difficult to separate different groups of faults, as large faults may change in strike by 90degrees in a comparatively short distance. It is not believed that such changes represent the intersection of different fracture systems, since the change in strike of the faults is usually gradual.

Faults with a small displacement are more common than can be indicated on a map of any reasonable scale. A double set of north-south faults is common in the central part of the area as are a series of small east-west faults. The faults cut folded or transposed structures and appear to be late structures.

The series of large curving faults are thought to be of different nature and appear to be related to the folding in the area. Of special importance are the large faults which bound the eastern sides of both the Upper Precambrian "sleeves". The fault on the east of the eastern tongue can be traced over most of its length except for a small part near the creek just south of the main "tillite" (i.e., southeast of Loc. 22), where alluvium covers the contact. Although the fault contact is rarely observed, the position of the fault can always be located to within 5 to 10 metres. In only a few places does the fault cross regions of significant relief (e.g. Localities 53-55) and the straight nature of the contact in these regions suggests that the fault is essentially vertical.

The movements on the fault must also have been essentially vertical, as the highly curved nature of outcrop of the fault is not consistent with a large lateral movement on a vertical fault. The apparent left lateral movement on the fault indicated by the disposition of the schist outcrops is probably due to the granite intrusion on the eastern side of the fault having displaced the schist boundary in a northerly direction.

The faults along the eastern and northern margins of the western Torrens Group sleeve (east of Subarea J and the ground around Subarea IV) are not so well exposed. The north-south part of the fault-system runs largely through partially covered pediment but its trace can be reasonably accurately located and established by plotting scattered low outcrops. North of the area where the fault diverges into at least three branches, the outcrop along the western branch is fairly good, and the position of the fault is well located. Its northern extension is not mapped beyond the point where displacement of beds can no longer be seen, although it seems quite likely that it could continue as a foliation fault for some distance to the northeast.

The middle branch (passing just north of Subarea IV) is covered by alluvium. The approximate trace of the fault has been determined on the occurrence of quartz-haematite breccias in the middle of the alluvium and the cutting out of the main dolomite. The southern branch of the fault, which runs along the southern boundary of Subarea IV, is considered to be well established, as the outcrop along its length is somewhat better exposed than along the central fault.

It is not known whether the two southern faults are merely branches of the main northern fault or whether they should be considered as part of a different system. In particular the northern and central faults cut out some of the sequence whereas the southern fault repeats the sequence. A possible solution is suggested in Fig. 111, a sketch section through Locality 48, but the author admits that he has no real evidence to support such an interpretation.

Movements in the western sleeve seem consistent with an east-west shortening in which the Sturt Group has over-ridden the Torrens Group and has come into contact with the Archaean. This interpretation would explain the buckling of the Sturt Group outcrops pattern, the much greater squeezing of the more incompetent Torrens Group schists and the considerable thinning of the beds above the "tillite". The faults would be hinge faults as suggested by the considerable increase in apparent horizontal displacement from north to south.

Fig. 111. Interpretation of faulting in Subarea IV.
North-south cross-section through Locality 48.

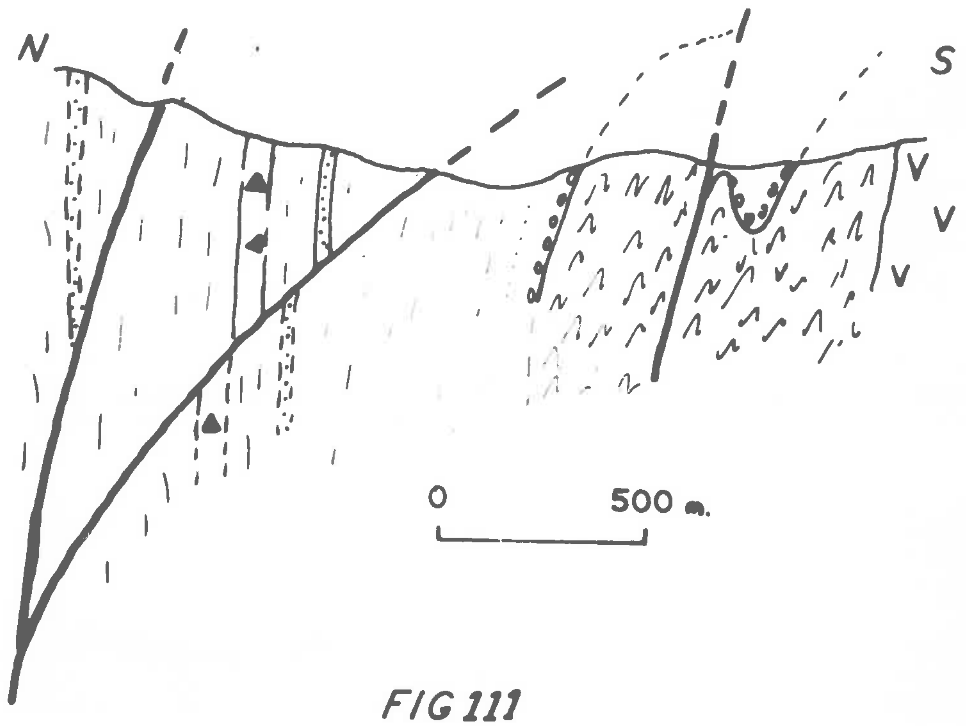


FIG 111

CHAPTER XV.

MACROSCOPIC STRUCTURAL ANALYSIS OF THE WHEY WHEY CREEK AREA.

Geometry.

Measurements in the Willyama Complex of layering (S_1); foliation (S_2); fold axes (B_2), without a crenulation cleavage parallel to the axial surface; crenulation cleavage (S_3); fold axes (B_3), with S_3 as axial surface; and axial surfaces of kinks (S_4), wer measured where possible and collective plots constructed (Fig. 112A). Measurements in the Upper Precambrian of bedding or transposed bedding (S); cleavage or schistosity (S_2); fold axes (B); pebble elongations (L) and rare crenulation cleavage (S_3) are shown in collective diagrams (Fig. 112B).

Subarea plots of S_1, S_2,S_3, S_4, B_2 and B_3 in the Willyama Complex and S_1, S_2 and B in the Upper Precambrian are shown on Plate V. In subareas where less than 25 measurements were taken of any particular structural element, those points were not contoured. A glance at Plate V therefore gives some idea of the relative abundance of each structural element in individual subareas (Subarea IV is an exception, as the subarea is small and S_4 is the only structural element which can be measured with any degree of certainty).

Both the collective and subarea plots of the structural elements in the Willyama Complex confirm the structural simplicity observed in the field. S_1, S_2 and S_3 are all

Fig. 112A. Collective diagrams of structural elements of the Willyama Complex.

πS_1	127 readings;	1-3-5-9 % contours per 1% area 13% Max.
πS_2	380 readings;	1-3-6-9 % contours per 1% area 19% Max.
πS_3	110 readings;	1-3-6-9 % contours per 1% area 18% Max.
πS_4	31 readings;	4-8-12-16% contours per 1% area 20% Max.
B_2	52 readings;	2-6-10-14% contours per 1% area 20% Max.
B_3	77 readings;	2-6-10-14% contours per 1% area 27% Max.

WILLYAMA COMPLEX

COLLECTIVE DIAGRAMS

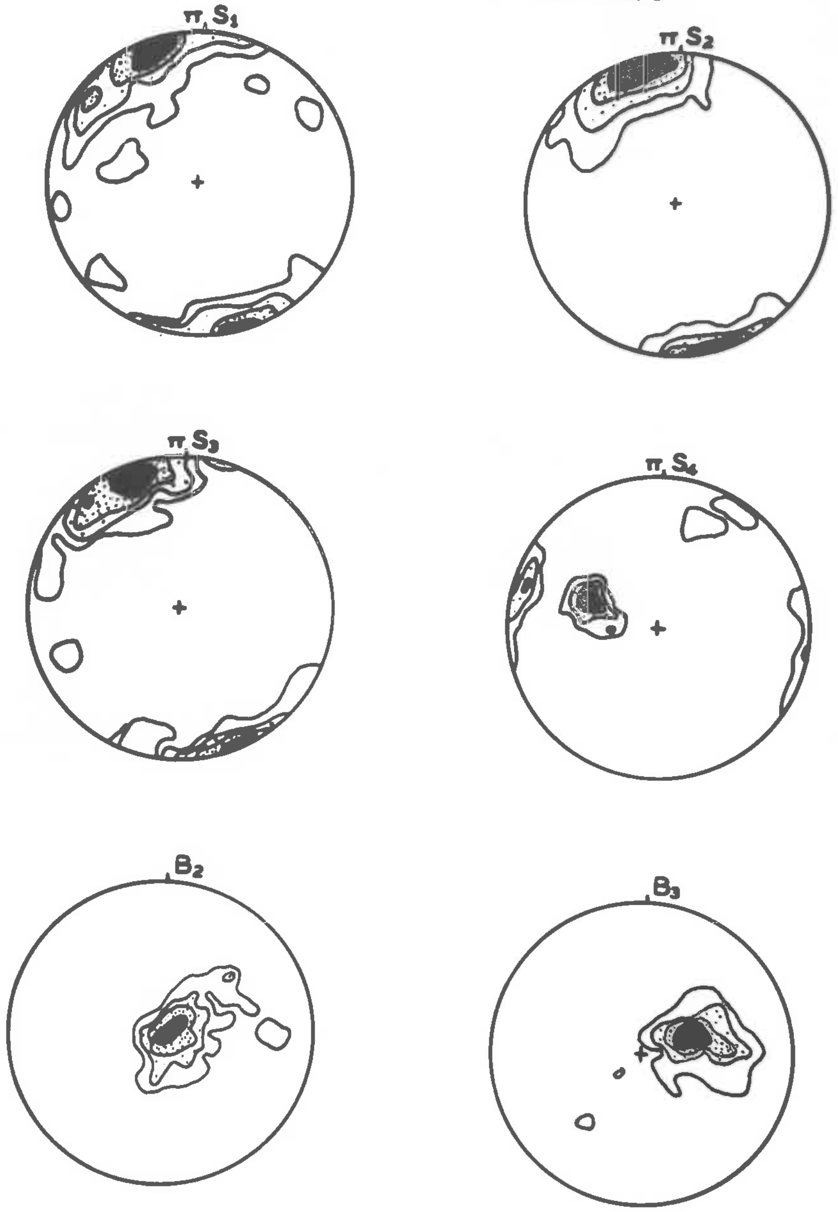


FIG 112A

Fig. 112B. Collective diagrams of structural elements of the Upper Precambrian.

πS 845 readings; 1-2-3-4% contours per 1% area
4% Max.

πS_2 217 readings; 1-5-10-15% contours per 1% area
16% Max.

B 152 readings; 1-3-5-9% contours per 1% area
15% Max.

L 28 readings.
(Pebble elongations)

πS_3 15 readings.

UPPER PRECAMBRIAN

COLLECTIVE DIAGRAMS

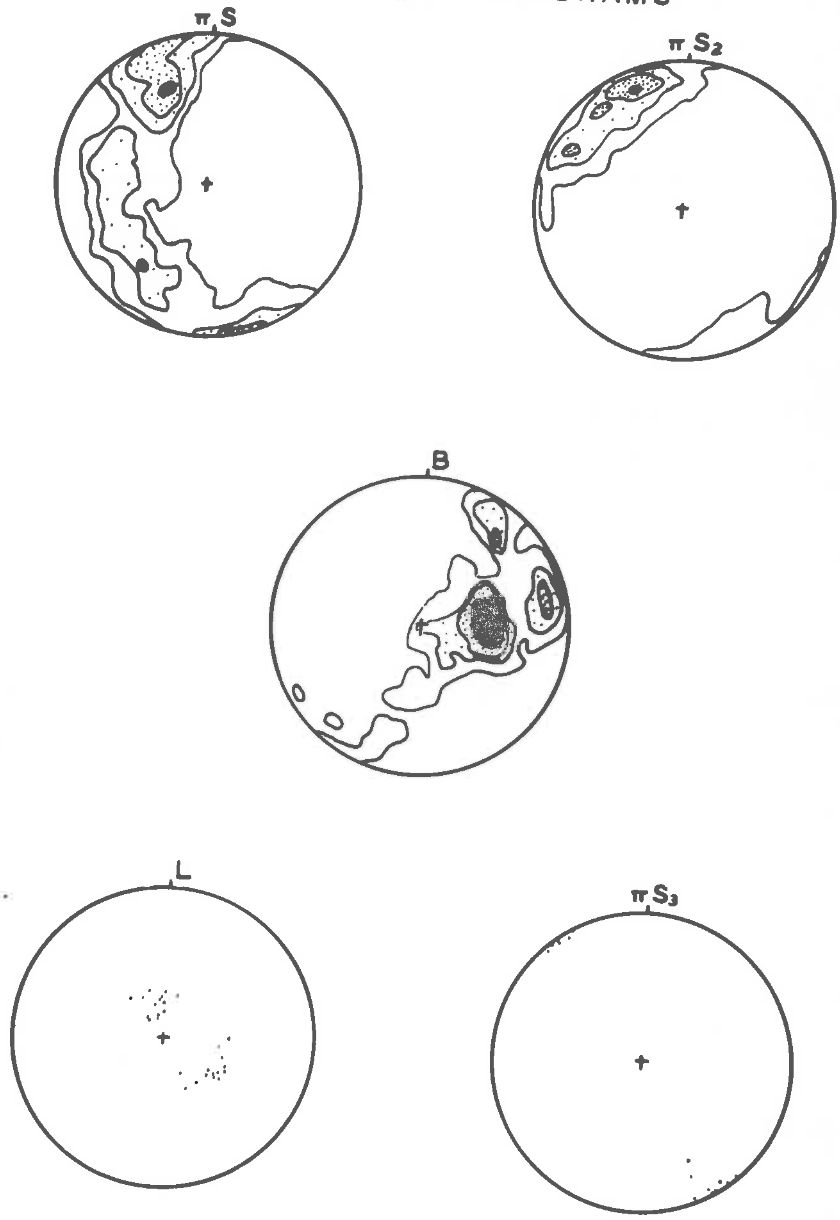


FIG 112 B

essentially parallel and B_2 and B_3 have a dominantly east plunge, although they vary considerably within the planes of S_2 and S_3.

The plots of the structural elements of the Upper Precambrian show a greater variation. Plots of Subareas A, B, C and D (Torrens Group) indicate moderate to steep easterly plunges of fold axes (Subarea D is rather irregular, due to the different plunge associated with the major fault on the east boundary of the area). The other subareas, all in the Sturt Group, show a much lower plunge and even reversals of plunge. This might be taken as evidence that the Torrens Group was deformed prior to the deposition of the Sturt Group. However, these differences can be explained in terms of the difference in proximity of the two groups to the Willyama Complex (see Chapter XIV).

The comparison of structural elements in the Willyama Complex with those in the Upper Precambrian shows certain similarities between the two suites. The foliation in both suites is essentially parallel and the attitude of fold axes in the Torrens Group are similar to those in the Willyama Complex.

Two subareas in the Upper Precambrian have not been plotted on Plate V. The first is the area in the far northeast (north of Subarea D) which shows the reversal in plunge of the axes in the Sturt Group. A collective plot of this subarea gives a false impression of the structure, and the variation in fold axes is in too small an area to be further subdivided on the scale of Plate V. Uncontoured diagrams of the limited data available from this area are shown in Fig. 113. These show the relatively inhomogeneous nature of this subarea.

The second subarea omitted from Plate V is that east of Subarea A and B. This subarea is extremely important but unfortunately lack of solid outcrop made it impossible to collect enough data in this area. The data collected (Fig. 114) indicate that in this subarea there is a very strong swing in foliation, layering and fold axes towards parallelism with the major fault direction. This swing, which is apparent in the foliation attitudes even in Subareas A and B, could be the result of a number of processes. It could represent overprinting of the east-west structure on a north-south structure or vice-versa, but no mesoscopic evidence was seen to support this hypothesis. Alternatively the structure could be an appressed basin structure similar to the "tillite" basin in the north. If so it is necessary to invoke some process which has rotated the cleavage into a more northerly attitude.

Movement picture (kinematics).

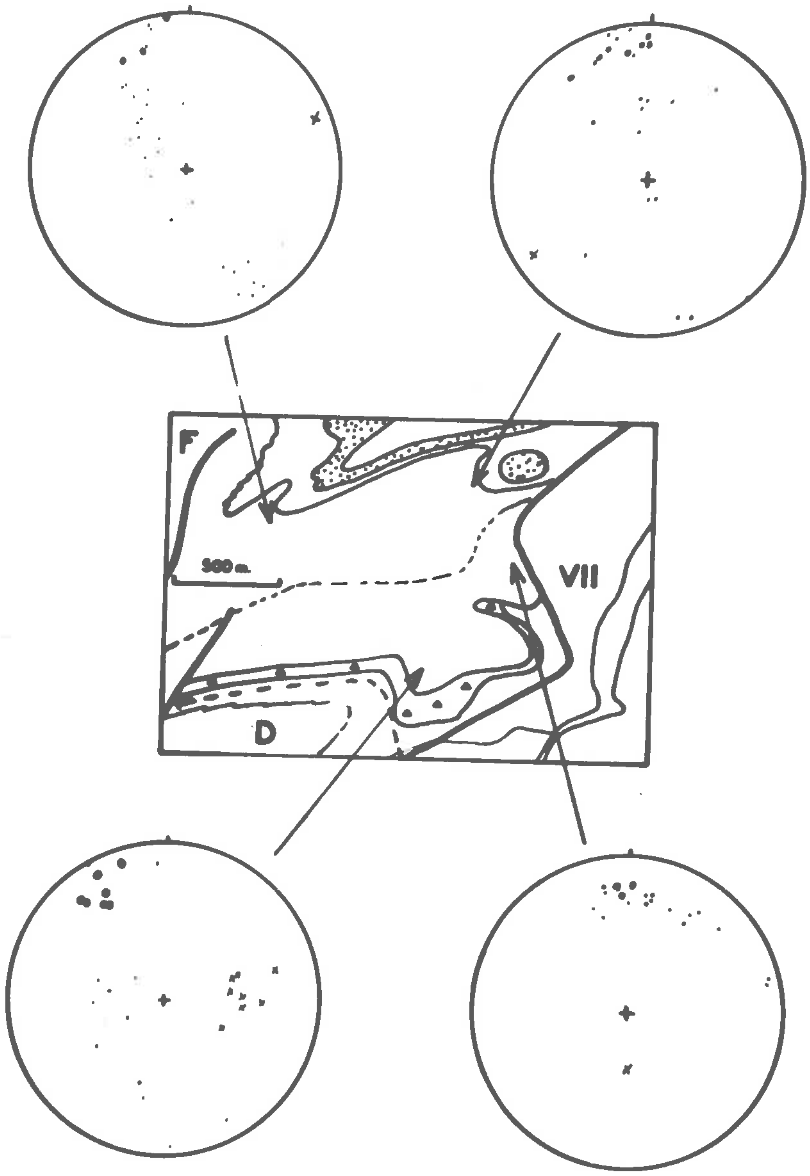
It has been demonstrated that there is a close geometric relationship between the structures in the various groups of rocks studied. It remains now to try to build a somewhat speculative movement picture relating the various structural

SUBAREA DIAGRAMS

NORTHEAST AREA

OF

PLATE IV



S • S • B ×

FIG. 113

COLLECTIVE DIAGRAMS
EAST OF SUBAREA
A

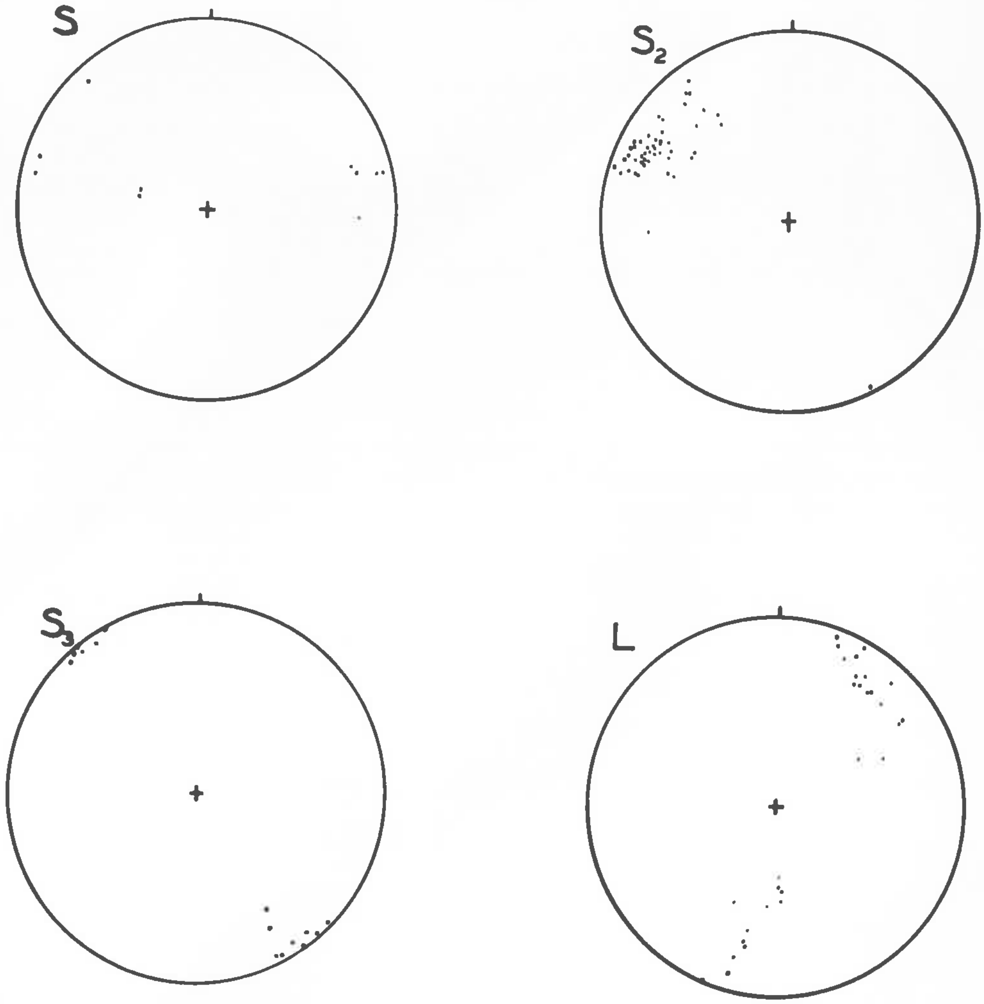


FIG. 114

elements observed.

Both mesoscopic and macroscopic evidence (e.g., the profile, Fig. 96) suggest that elements of both similar and concentric style of deformation are present in the final structures. To a large extent the deformation of the mica schists can be considered an S-passive deformation, the laminations playing only a small part in the deformation. The thicker layers may have exerted a constraining influence on the homogeneity of deformation. On the other hand the style of deformation in the thicker quartzite horizons is concentric and indicates an S-active deformation, with a considerable shortening perpendicular to the axial surface. As Carey (1962) points out this shortening can be purely a local effect and may not be found in the underlying strata, but in the absence of distinct planes of separation between the quartzites and mica schists it seems reasonable that the shortening in the quartzites is also present in the mica schists. This deformation could be accomplished by a plastic flattening perpendicular to the axial surface with extension perpendicular to andpossibly also parallel to the steep fold axes.

The flattening indicated by the deformed pebbles may not be applicable to the area as a whole. The amount of highly deformed conglomerate is small, but the orientation of the long axes is fairly uniform over the whole area and may indicate constant strain axes for the area. The axis of maximum (positive) strain is perpendicular to the foliation.

and the axis of maximum release is approximately vertical. It is very doubtful whether these strain axes can be correlated with movement axes. The symmetry of deformation of the pebbles (orthorhombic) is not the same as the symmetry of folding (monoclinic). One of the symmetry planes of the deformed pebbles corresponds to the foliation but the other symmetry planes are neither parallel nor perpendicular to the symmetry plane of the monoclinic folds. It is therefore not possible to relate the fabric axes of the pebbles to those of the folding.

The movement picture therefore has to be deduced for an area which has essentially triclinic symmetry but which on a large scale may be orthorhombic. The movements in the Torrens Group appear to be related to a single period of triaxial strain in which the axis of greatest shortening was perpendicular to the axial surface of the Torrens Group folds (and Willyama Complex foliation). A secondary axis of positive strain was oriented parallel to the foliation but was essentially horizontal. The bulk strain was accomplished by folding on east-northeast axes and possibly also by plastic flattening. The strain perpendicular to this direction appears to have been accomplished partly by cross-folding, resulting in variations in plunge in the Sturt Group and partly by faulting. In the Torrens Group the shortening was more localized and was manifested by intense deformation accompanied by faulting in the western sleeve and to a less extent in the eastern sleeve. The analogy of squeezing between the jaws of a vice (the two blocks of the Willyama Complex) is particularly striking in the western sleeve but may also be applied to all parts of the Upper Precambrian where adjacent to major faults.

These movements are consistent with the formation of variable plunges in the younger sequence and with the vertical axis of relief indicated by the deformed conglomerates. During this phase of deformation the Willyama Complex appears to have been plastically deformed without the formation of large folds. The greater ease of deformation of the mica schists has resulted in differential concentration of the strain in these rocks and it is believed that the mica schists have "flowed" and localized the more complex folds in the Torrens Group adjacent to them. The presence of more massive gneisses both to the west and the east may have resulted in an essentially vertical release in the central block of the Willyama Complex giving rise (in part) to the steeper fold axes in the Torrens Group.

CHAPTER XVI.

RELATIONSHIPS BETWEEN THE "ARCHAEAN" AND THE UPPER PRECAMBRIAN ROCKS.

In the previous chapters the evidence relating to the metamorphic and structural histories of two areas in South Australia has been discussed. In this final chapter some general conclusions are advanced concerning the relationships between metamorphic sequences of two different ages.

Metamorphic relationships.

In both areas examined it was found that the present metamorphic grade of the two sequences in contact are essentially similar. In the Paracombe-Mt. Gawler area the metamorphic grade of the two sequences is that of the lower greenschist facies. The Torrens Group rocks are characterized by assemblages of quartz-sericite-chlorite. Carbonates are also common. Albite and microcline may be solely relic sedimentary minerals but are stable under these conditions of metamorphism. The presence of microcline outgrowths on microcline could be an authigenic change or a higher grade metamorphic one (Baskin, 1956). These assemblages, together with the general absence of biotite, are consistent with the quartz-albite-muscovite-chlorite subfacies of the greenschist facies (chlorite zone) (Fyfe and others, 1958).

The adjacent Houghton Complex also contains assemblages consistent with the lower greenschist facies. The mineral assemblages are characterized by combinations of quartz, albite, muscovite, chlorite, epidote and actinolite. Biotite occurs rarely but may indicate a somewhat higher grade of metamorphism in parts of the Houghton Complex.

As well as these assemblages characteristic of the greenschist facies, relic assemblages of an earlier higher metamorphic grade are present in the Houghton Complex. Assemblages occurring as relics are microcline (perthite)-oligoclase (to andesine)-diopside; feldspar-hornblende; rocks with relic sillimanite, andalusite-sillimanite, garnet and biotite also occur. Epidote may have been part of the primary assemblage. These assemblages are consistent with the amphibolite facies (and also with the hornblende-granulite facies, but the absence of hypersthene and the possible presence of epidote is more consistent with the amphibolite facies). The rocks are thought to have been schists and gneisses with pegmatitic (or metasomatic) schlieren. The feldspar gneisses may have been igneous or metasomatic in origin.

The relationship of the period (or periods) of the retrograde metamorphism to the Torrens Group is not completely understood. The Houghton Complex must have undergone a period of retrograde metamorphism before the deposition of the Torrens Group, presumably in the greenschists facies. However, since the Torrens Group has also been metamorphosed to the greenschist facies, it is not known whether the main period of retrogression in the Houghton Complex was related to the earlier or later greenschist facies metamorphism. The only features which might be separated into two phases of retrogression are (1) the change andesine to oligoclase-albite and epidote, compared with the change oligoclase to sericite, and (2) the formation of zones of oriented sericite in rocks which showed a more uniform degree of alteration. These features could, however, have been produced in a single phase of metamorphism in the greenschist facies.

The grade of metamorphism in the area mapped (i.e., greenschist facies) is lower than in the surrounding areas. The grade increases sharply just to the east of the area mapped and the author has found sillimanite just north of Gumeracha. Conditions of higher grade metamorphism are more general about 5-10 miles east of the area mapped. Assemblages contain andalusite, staurolite, kyanite, garnet and sillimanite (Kleeman and Skinner, 1959).

Higher grade rocks (containing sillimanite) are also reported on the Gawler military sheet north of the area mapped (Campana and others, 1955; Alderman, 1942). Rocks of the biotite grade occur to the south of Adelaide (Campana and others, 1953). Kleeman and White (1956) state that the highest grades of metamorphism in the Mt. Lofty Ranges occur to the east of the "Archaean" masses (near Tanunda) and that the grade of metamorphism decreases rapidly out from this high grade region. The metamorphic evidence in the area mapped fits into this pattern of increasing metamorphism to the east. Most of the area mapped can be considered to be in the chlorite zone but a diffuse zone running northwest from Gumeracha, and including parts of the Houghton Complex defines an approximate biotite isograd. The presence of the "Archaean" core does not seem to have affected the distribution of metamorphic zones.

In the Whey Whey Creek area the rocks of the Willyama Complex and overlying Upper Precambrian are also similar in metamorphic grade. The characteristic mineral assemblages in the Willyama Complex are quartz-microcline-albite-muscovite-biotite; albite-actinolite-epidote; quartz-muscovite-biotite; quartz-muscovite-biotite-garnet; actinolite-albite-garnet-sphene. One outcrop containing staurolite, chloritoid and sillimanite may be a relic from an overall higher metamorphic grade. Knots of muscovite observed in the schists may be after andalusite and Mr. K. J. Mills reports andalusite and abundant garnet to the east of the area mapped. However, most of the assemblages observed in the Whey Whey Creek area are consistent with the biotite and garnet zones of the greenschist facies of metamorphism, and if the rocks are retrograde from a higher grade of metamorphism most of the evidence for this higher grade has been destroyed. It is perhaps suggestive that the area contains non-layered granites, massive pegmatites and numerous migmatites. Although these rocks could have formed in the greenschist facies they are more consistent with the amphibolite facies

(Turner's Fig. 107 in Fyfe and others, 1958) which the rare relics suggest. The present mineralogy of the rocks would then reflect a peiod of recrystallization in the upper greenschist facies.

The metamorphic grade of the Upper Precambrian rocks is also of the greenschist facies. The characteristic assemblages are quartz-muscovite-biotite; quartz-feldspar-biotite-actinolite; carbonate-tremolite. These assemblages are characteristic of the middle greenschist facies (biotite zone) and as noted before, the only equilibrium mineral found in the Willyama Complex but lacking in these assemblages is garnet. The presence of the rare garnet in the Willyama Complex may indicate an upper greenschist facies compared with the middle greenschist facies of the Torrens Group, but it is possible that the garnet is a relic mineral from the period of higher grade metamorphism.

The rocks of the two areas mapped show considerable differences in appearance. The widespread cataclasites of the Houghton Complex contrast strongly with the coarser-grained schists and gneisses of the Willyama Complex. Similarly the relatively abundant rocks with relic higher grade minerals in the Houghton Complex compares with their relative absence in the Willyama rocks in the Whey Whey Creek area. It is thought that these differences arise mainly through the differences in metamorphic grade between the two areas mapped. An increase in metamorphism from chlorite zone to biotite zone east of the Houghton Complex is accompanied by a large increase in grain size and obliteration of the original textures (but not larger scale structures such as bedding). Similarly it is expected that retrogression at the biotite zone will have resulted in more complete breakdown of the original minerals than retrogression at the chlorite zone. The lower grade of retrogression at Houghton appears to have resulted in dominantly cataclastic effects. In the Willyama Complex on the other hand the higher grade of metamorphism has resulted in schists and gneisses with a more normal appearance, with equilibrium grain growth fabrics and only rare plastic deformation fabrics.

Structural Relationships.

Areas in which two different sequences of rocks are deformed together are fairly common in many parts of the world. Several types of structural relationships between the two sequences are reported in the literature. Examples of relatively unmetamorphosed rocks of a younger sequence resting on a crystalline basement are recorded by Hudson (1955), Foose and others (1961) and Butcher (1962). In these examples the younger rocks appear to drape over an older sequence which has been either faulted, sheared or folded.

At a somewhat higher grade of metamorphism and deformation, parallel foliations in the two sequences are developed, and commonly the contact zones are highly sheared and any structures in the underlying sequence are rotated into parallelism with the contact. Such relationships have been reported by Oriel (1951), Brace (1958), Sutton and Watson (1958), Choquette (1960) and Bryant and Read (1962). Ross (1962) has attempted to analyse the effect of the folding in the upper sequence on the structures of the lower sequence. His conclusions are based on the supposition that the structures in the older sequence played a generally passive part in the deformation, or have been folded. Neither of these conditions are important in the areas studied by the author.

At even higher grades of metamorphism both the younger and older rocks may become mobilized and appear to act as a single unit. Under this heading are the mantled gneiss domes, the relationships of which were reviewed by Eskola (1949). However these conditions of metamorphism appear quite different from the South Australian examples, although some of Eskola's examples show broad features in common with those studied by the author.

In all of these studies cited the effects of pre-existing structures on the deformation of the younger sequence have apparently been unimportant. Even in regions where the strike of layering in the younger series cuts across the dominant strike of the foliation in the older sequence, the foliation trends swing parallel to the contact.

In South Australia and the bordering Barrier Ranges of New South Wales, relationships different from those described elsewhere are important. Leslie and White (1955), in a study of the unconformity between the Torrowangee series and the Willyama Complex, pointed out that the structure in the older sequence is much simpler than the structure observed in the Torrowangee rocks. They offered several alternatives in explanation. The main alternatives were (1) that the younger sequence was folded independently of the basement, (2) that faulting in the "Archaean" is reflected in folding in the Torrowangee rocks and (3) that differences in plasticity between the two sequences has given rise to differences in folding.

In the Whey Whey Creek area none of these alternatives will explain completely the relationships observed. The contacts between the Willyama Complex and the Torrens Group are most commonly unsheared, and only in the large highly appressed synclines can any significant movement or detachment have taken place. The correlation of mesoscopic and macroscopic folds in the Torrens Group with differences in rock type in the Willyama Complex suggests that differences in bulk strain in the different rocks of the older sequence have localized these folds. There is no evidence that any large scale folds in the Willyama Complex correspond to the scale of folds in the Torrens Group, and the small scale folds in the Willyama Complex are not important in determining the structures in the overlying beds. It has been noted that even close to large folds in the Torrens Group the foliation in the Willyama Complex maintains its planer preferred orientation and is not folded into conformity with the contact.

Rather the layering in the Torrens Group is seen to swing parallel to the foliation in the Willyama Complex in the cores of appressed synclines. It appears therefore that during the deposition of the Torrens Group the gross layering and foliation in the Willyama Complex were vertical. This perpendicular relationship must have profoundly influenced the deformation of the Torrens Group. The prior existence of a well-developed planar structure in the Willyama Complex has resulted in the concentration of deformation in zones parallel to the foliation, which has been modified in part by a crenulation cleavage. Although the strain in the Willyama Complex is probably reflected in the Upper Precambrian rocks, the deformation in the latter is to a large extent localized by the influence of the basement.

The contact of Upper Precambrian and basement rocks in the northwest part of the Whey Whey Creek area differs from the southern contact in that it is faulted over much of its length. The differences between these contacts probably resulted from the more massive nature of the "Archaean" in the north.

In the Paracombe-Mt. Gawler area the contact relationships are not so well exposed as at Whey Whey Creek, but the relationships appear to be essentially similar. Over much of the Houghton Complex a planar preferred orientation of the foliation is the dominant structural feature. This foliation is largely a secondary one, and its age relation to the deformation of the overlying Torrens Group has not been firmly established. However, as at Weekeroo, this foliation appears to have been active in the deformation. In the southern part of the area, in the"nose" of the regional anticline, rare contacts between the two sequences show a swing in the foliation into near parallelism with the contact. This relationship is similar to most of the examples cited in the literature. The swing of the foliation appears to result from shearing along the contact.

The structures in the Torrens Group have a style which conforms with the large scale deformation. Moderately dipping east limbs with faulted steep west limbs, indicate tectonic transport in the westerly direction.. The Houghton Complex appears to have acted as a simple core structure and has adjusted to the folding in the overlying Torrens Group by a predominantly S-passive deformation.

These relationships deduced from a study of two small areas in South Australia appear to have a broader application to other areas in the State, and an intimate relationship between the older rocks and the stratigraphy and structure of the Upper Precambrain rocks may prove to be of more regional importance than hitherto realized.

APPENDIX I.

ANALYSES OF HOUGHTON COMPLEX ROCKS.

Analysis	Rock type	Locality	Reference
A	Diorite	Houghton Complex	Benson, 1909.
B	Diorite	"	Alderman in Mawson, 1926.
C	Adamellite	"	Spry, 1951.
D	Syenite	Yankalilla	England, 1935.
E	Syenite	"	"
F	Diorite	"	"
G	Syenite	Mt. Compass	"
H	Dolerite	"	"
I	Diorite	"	"
J	Dolerite	"	"
K	Diorite	Aldgate	"
L	Augen gneiss	Humbug Scrub	Alderman, 1938.
M	Augen gneiss	"	"
N	Augen gneiss	"	"
O	"Soda hybrid"	"	"
P	Banded gneiss	"	"
Q	Schist	"	"
R	Schist	"	"

	A	B	C	D	E	F	G	H	I	J	K
SiO_2	56.85	58.19	69.25	53.31	52.08	54.95	60.62	48.66	51.78	49.99	61.09
Al_2O_3	14.76	15.28	13.40	15.79	15.67	15.88	14.94	17.99	15.53	15.54	15.75
Fe_2O_3	4.48	1.58	2.69	3.69	5.38	0.68	8.12	2.13	4.47	3.07	2.94
FeO	1.21	1.23	1.47	1.37	1.76	5.85	1.07	8.44	6.91	8.44	1.77
MgO	3.84	3.85	1.53	1.73	2.62	5.06	1.96	6.72	5.42	6.61	4.55
CaO	7.91	8.72	2.47	13.81	13.48	10.09	2.60	10.81	8.42	11.15	6.43
Na_2O	5.34	4.76	3.03	1.04	0.99	3.76	3.40	2.23	2.82	1.74	4.69
K_2O	1.91	3.02	4.66	4.15	4.31	0.76	6.16	0.30	0.72	0.43	0.73
H_2O+	0.12	0.16	0.49	0.51	0.53	0.41	0.33	0.37	0.69	0.89	0.51
H_2O-	0.08	0.29	0.16	0.21	0.18	0.22	0.13	0.19	0.20	0.14	0.08
P_2O_5	0.51	0.48	0.13	1.20	0.40	0.24	0.44	0.33	0.51	0.45	0.33
MnO	0.12	0.05	0.01	0.45	0.27	0.04	0.01	0.20	0.23	0.19	0.21
TiO_2	3.11	2.65	0.91	0.70	0.55	1.80	0.76	2.18	1.94	1.50	0.87
CO_2	-	-	-	2.01	2.37	-	-	Trace	-	-	n.d.
BaO	-	-	Trace	0.06	0.06	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.
	100.24	100.26	100.16	100.03	100.65	99.74	100.54	100.55	99.94	100.14	99.95

	L	M	N	O	P	Q	R
SiO_2	66.89	71.19	69.69	59.26	68.31	61.53	60.02
TiO_2	0.80	0.50	0.90	0.49	0.82	0.75	n.d.
Al_2O_3	14.96	15.26	15.51	22.94	15.12	20.70	21.84
Fe_2O_3	2.53	1.66	2.21	2.41	2.52	2.96	4.67
FeO	1.73	1.05	1.29	1.49	1.46	2.35	as Fe_2O_3
MnO	0.01	-	-	n.d.	n.d.	-	n.d.
MgO	1.57	0.78	1.10	1.00	1.22	1.69	2.11
CaO	1.59	0.56	0.44	0.63	0.91	0.26	0.43
Na_2O	2.13	2.64	2.84	7.13	2.41	0.71	1.09
K_2O	5.54	5.87	4.91	3.42	5.33	6.24	6.25
P_2O_5	0.17	-	-	n.d.	n.d.	n.d.	n.d.
H_2O+	1.23	0.96	1.12	1.33	1.11	2.87	n.d.
H_2O-	0.22	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.
CO_2	0.89	-	-	n.d.	0.55	n.d.	n.d.
	100.26	100.47	100.02	100.10	99.76	100.06	(partial analysis)

APPENDIX II.

PETROFABRIC STUDIES.

A limited number of quartz orientation diagrams were prepared to determine the relationship between the microfabric and the mesoscopic structures.

In the Paracombe-Mt. Gawler area, nine diagrams were prepared from the retrograde rocks of the Houghton Complex, and one from a rock showing granulitic fabric. Quartz orientations in three Torrens Group rocks were also studied.

In the Whey Whey Creek area, six diagrams were prepared from pebbles in the basal conglomerates of the Torrens Group and two from a foliated quartzite.

Statistically significant diagrams are shown in Plate VI. Contours are 1-2-3-etc % per 1% area.

SPECIMENS STUDIED.

Houghton Complex.

157-1375. Quartz-rich gneiss. The rock consists of an aggregate of coarse-grained quartz with minor feldspar and sericite. The quartz grains show undulatory extinction but are equidimensional. No significant pattern was seen after measuring 255 grains.

157-1374. A more highly deformed variant of 157-1375. The quartz grains are elongate parallel to a weak foliation. The quartz diagram shows monoclinic symmetry with a diffuse girdle perpendicular to a mesoscopic lineation (Fig. A, 259 grains).

157-283. Augen schist with zones of fine-grained quartz and sericite separating relic zones rich in quartz. The large grains show undulatory extinction but show no significant preferred orientation. Small quartz grains in a cross-cutting vein of quartz show a weak girdle perpendicular to the lineation and a maximum parallel to the lineation (Fig. B, 230 grains).

Small quartz grains in one of the retrograde zones show an incomplete girdle perpendicular to the lineation (Fig. C, 212 grains). However the maxima in diagrams B and C do not coincide.

157-71. Augen schist with a strong retrograde foliation. Diagrams prepared from both the large relic grains and the small polygonal grains show no significant preferred orientation.

157-172. Augen schist with rare quartz relics. The polygonal quartz grains show no significant preferred orientation. Two hundred and twenty polygonal grains in a single large quartz grain were measured separately from the other quartz grains. The preferred orientation is poor but the c-axes tend to be concentrated in the vicinity of the orientation of the c-axis (H in Fig. D) of the host grain (Fig. D, 220 grains). (Diagram prepared from field around that of text Fig. 41).

157-994A. Quartz-feldspar gneiss. The quartz occurs in highly elongate spindles, imparting a strong lineation to the rock. The diagram shows a strong incomplete girdle perpendicular to the lineation (Fig. E, 230 grains).

Torrens Group. Paracombe- Mt. Gawler area.

A specimen of a quartz-rich mica schist from north of Gumeracha

(157-1350, 267 grains) and two specimens of quartzites (157-34 and 62F, 302 grains in each) showed no significant preferred orientations.

Whey Whey Creek area.

Quartzite.

Four hundred grains were measured in each of two foliated quartzites. The diagrams are similar, and show two girdles intersecting in a line which is not perpendicular to the local fold axis. The axis perpendicular to this intersection is about 30° steeper than the fold axis; that is, it plunges about 75° to the northeast. Fig. F, 231-45; Fig. G, 231-46.

Conglomerates.

Five quartz diagrams from six pebbles in the basal conglomerate show similar features (Figs. H, J, K, M and N). Two girdles are more or less strongly developed symmetrically to the foliation and are broken in the region parallel to the foliation (particularly Fig. H). The best preferred orientation (Fig. N) is form a highly deformed conglomerate; the other diagrams are from weakly deformed conglomerates.

The apparently aberrant diagram (Fig. L) shows only one of the girdles shown in Figs. K and M (from the same specimen). In all three diagrams the northwest maximum is only weakly developed.

Fig. H, 231-24B, 412 grains; Fig. J, 231-6, 410 grains. Figs. K, L and M are from separate pebbles in the same specimen (231-24c). Fig. K, 355 grains; Fig. L, 259 grains; Fig. M, 300 grains. Fig. N, 231-104, 353 grains.

Discussion.

Petrofabric studies in the Paracombe-Mt. Gawler region confirm the findings of Kleeman (1954) who reported no significant preferred orientations from the retrograde schists of the Houghton Complex. However, a few of the diagrams show girdles perpendicular to a weak lineation. The granulitic gneiss shows a stronger concentration of points but with the same general pattern. Diagrams A, B, C and E show monoclinic symmetry with the lineation perpendicular to the symmetry. The girdles are therefore ac (fabric) girdles, and the lineation is parallel to a b fabric axis. However, the relationship of the fabric axes to possible kinematic axes cannot be deduced with confidence from the data available.

In the Whey Whey Creek area the patterns of preferred orientation of quartz are very similar to those published in the literature for deformed pebbles (Brace, 1955; Strand, 1944 and Flinn, 1956). Balk (1952) described similar features from quartzites in the neighbourhood of thrust faults. It is of note that these authors interpret the lineations as parallel to an a axis, and the girdles are therefore bc girdles. However, their choice of axes was based on considerations of other fabric elements and so may not be valid.

The fabric in the quartzites is of interest in indicating a b fabric axis which is steeper than the local fold axis. This relationship is similar to the mesoscopic evidence that the long axes of pebbles are steeper than the bedding-cleavage intersections.

Flinn (1956) reports a similar situation in the deformed Funzie conglomerate. He demonstrated that the long axes of the pebbles remained relatively constant over the whole area, although the bedding-cleavage intersections vary considerably. Such a situation, which is also present in the Whey Whey Creek area, indicates the difficulty in correlating fabric axes of one scale of homogeneity with those of another scale of homogeneity.

REFERENCES

Agron, S.L. 1950. Structure and petrology of the Peach Bottom Slates, Pennsylvania and Maryland, and its environment. Bull.Geol.Soc.Amer., 61, 1256-1306.

Alderman, A.R. 1938. Augen-gneisses in the Humbug Scrub area, South Australia. Trans.Roy.Soc.S.Aust., 62, 168-191.

——— 1942. Sillimanite, kyanite and clay deposits near Williamstown, South Australia. Trans.Roy.Soc.S.Aust., 66, 3-14.

Anonymous 1958. Australian Code of Stratigraphic Nomenclature (third edition). Jour.Geol.Soc.Aust., 6, 63-70.

Bailey, S.W., Bell, R.A. and Peng, C.J. 1958. Plastic deformation of quartz in nature. Bull.Geol.Soc.Amer., 69, 1443-1466.

Balk, R. 1952. Fabric of quartzites near thrust faults. Jour.Geol., 60, 415-435.

Baskin, Y. 1956. The study of authigenic feldspar. Jour.Geol., 64, 132.

Battey, M.H. 1955. Alkali metasomatism and the petrology of some keratophyres. Geol.Mag., 92, 104-126.

Benson, W.N. 1909. Petrographical notes on certain Pre-Cambrian rocks of the Mt Lofty Ranges with special reference to the geology of the Houghton district. Trans.Roy.Soc.S.Aust., 33, 101-140.

Bowes, D.R. and Jones, K.A.
1958. Sedimentary features and tectonics in the Dalradian of Western Perthshire. Trans.Edin.Geol.Soc., 17, 133-140.

Brace, W.F. 1955. Quartzite pebble deformation in central Vermont. Amer.Jour.Sci., 253, 129-145.

——. 1958. Interaction of basement and mantle during folding near Rutland, Vermont. Amer.Jour.Sci., 259, 241-256.

Brown, H.Y.L. 1886. Geological map of South Australia exclusive of the Northern Territory. Govt. Printing Office, South Aust.

—— and Woodward, H.P.
1885. Geological map of Barossa and Para Wirra. Parl. Paper, S.Aust., 178.

Bryant, B. and Reed, J.C.
1962. Structural and metamorphic history of the Grandfather Mountain area, North Carolina; A Preliminary Report. Amer.Jour.Sci., 260, 161-180.

Butcher, N.E. 1962. The tectonic structure of the Malvern Hills. Proc.Geol.Assoc., 73, 103-124.

Campana, B. 1955. The structure of the eastern South Australian Ranges: the Mt Lofty-Olary arc. Jour.Geol.Soc.Aust., 2, 47-61.

—— 1956. Geological atlas of South Australia. Plumbago 1 mile sheet. Geol.Surv.S.Aust.

Campana, B. and King, D.
1958. Regional geology and mineral resources of the Olary Province.
Bull.Geol.Surv.S.Aust., 34, 133 pp.

—— Wilson, R.B. and Whittle, A.W.G.
1953. The geology of the Jervis and Yankalilla military sheets.
Dept. of Mines Rept. of Investigations, No.3.

Carey, S.W. 1962. Folding.
Jour.Alberta Soc.Petrol.Geol., 10, 95-144.

Choquette, P.W. 1960. Petrology and structure of the Cockeysville Formation (Pre-Silurian) near Baltimore, Maryland.
Bull.Geol.Soc.Amer., 71, 1027-1052.

Christie, J.M. 1960. Mylonitic rocks of the Moine Thrust-Zone in the Assynt region, North-West Scotland.
Trans.Roy.Soc.Edin., 18, 74-93.

Clarke, F.W. 1924. Data of geochemistry.
Bull. U.S. Geol. Surv., 770.

Cummings, W.A.J. and Shackelton, R.M.
1955. The Ben Lui recumbent syncline (South-West Highlands).
Geol.Mag., 92, 353-363.

Daily, B. (in press). The fossiliferous Cambrian succession on Fleurieu Peninsula, South Australia.
Records S.A.Museum, 14, No.3.

David, T.W.E. 1922. Occurrence of remains of small crustacea in the Proterozoic(?) or Lower Cambrian (?) rocks of Reynella, near Adelaide.
Trans.Roy.Soc.S.Aust., 46, 6-8.

David, T.W.E. 1927. Notes on the geological horizon of the Archaeocyathinae. Trans.Roy.Soc.S.Aust., 51, 410-413.

——— 1928. Nullagine - Adelaide Series. Rep.Aust.Ass.Advance.Sci., 19, 231-234.

de Sitter, L.U. 1956. Structural Geology. McGraw Hill, 552 pp.

Dickinson, S.B., Sprigg, R.C., King, D., Wade, M.L., Webb, B.P. and Whittle, A.W.G. 1954. Uranium deposits in South Australia. Bull.Geol.Surv.S.Aust., 30.

Dietrich, R.V. 1960a. Banded gneisses. Jour.Pet., 1, 99-120.

——— 1960b. Banded gneisses of the Randesund area, Southern Norway. Norsk Geol. Tidds., 40, 13-63.

Dunroth, E. Verformung und Großbau in Südwestlichen Fichtelgebirge. Neues Jahr.für Geol. und Pal., 115, No.1, 121-135.

Emmons, R.C. 1943. The Universal Stage. Mem.Geol.Soc.Amer., 8, 205 pp.

England, H.N. 1935. Petrographic notes on intrusions of the Houghton magma in the Mt Lofty Ranges. Trans.Roy.Soc.S.Aust., 59, 1-15.

Eskola, P. 1948. The problem of mantled gneiss domes. Quart.Jour.Geol.Soc.London, 104, 461-476.

Evans, B.W. and Leake, B.E. 1960. The composition and origin of the striped amphibolites of Connemara, Ireland. Jour.Pet., 1, 337-362.

xxx

Flinn, D. 1956. On the deformation of the Funzie conglomerate, Fetlar, Shetland. Jour.Geol., 64, 480-505.

——— 1958. On the nappe structure of North-East Shetland. Quart.Jour.Geol.Soc.London, 114, 107-136.

Foose, Richard M., Wise, D.U., Garbarini, G.S. 1961. Structural geology of the Beartooth Mountains, Montana and Wyoming. Bull.Geol.Soc.Amer., 72, 1143-1172.

Forbes, B.G. 1960. Magnesite of the Adelaide System: Petrography and descriptive stratigraphy. Trans.Roy.Soc.S.Aust., 83, 1-9.

Fyfe, W.S., Turner, F.J. and Verhoogen, J. 1958. Metamorphic reactions and metamorphic facies. Mem.Geol.Soc.Amer., 73, 251 pp.

Gates, R.M. 1953. Petrogenic significance of Perthite. in Mem.Geol.Soc.Amer., 52. (Ed. Emmons).

Gillott, J.E. 1956. Structural geology of the Manx slates. Geol.Mag., 93, 301-313.

Gilluly, J. 1933. Replacement origin of the albite granite near Sparta, Oregon. U.S.Geol.Surv. Prof.Paper, 175-C, 63-81.

Glaessner, M.F. and Parkin, L.W. (Eds.) 1958. The Geology of South Australia. Jour.Geol.Soc.Aust., 5, pt.2, 1-163.

Goldich, S.S. and Kinser, J.H. 1939. Perthite from Tory Hill, Ontario. Am.Min., 24, 407-427.

Griggs, D. and Handin, J. (Eds.)
1960. Rock deformation.
Mem.Geol.Soc.Amer., 79, 382 pp.

Hamilton, W. 1960. Geology of northeasternmost Tennessee.
U.S.Geol.Surv. Prof.Paper, 311, 13-27.

Harland, W.B. and Bayly, M.B.
1958. Tectonic Regimes.
Geol.Mag., 95, 89-104.

Harry, W.T. and Emeleus, C.H.
1960. Mineral layering in some granite intrusions of South-West Greenland.
Int.Geol.Congr.Rep., 21, pt.14, 172-181.

Hietanen, Anna. 1938. Petrology of the Finnish quartzites.
Comm.Geol.Finlande Bull., 122, 1-118.

Hoeppener, R. 1956. Zum Problem der Bruchbildung, Schieferung und Faltung.
Geol.Rundschau, 45, 247-283.

Hossfeld, P.S. 1935. The geology of part of the north Mt Lofty Ranges.
Trans.Roy.Soc.S.Aust., 59, 16-67.

Howchin, W. 1897. On the occurrence of Lower Cambrian fossils in the Mt Lofty Ranges.
Trans.Roy.Soc.S.Aust., 21 (2), 74-86.

——— 1904. The geology of the Mt Lofty Ranges. Part I.
Trans.Roy.Soc.S.Aust., 28, 253-280.

——— 1906. The geology of the Mt Lofty Ranges. Part II.
Trans.Roy.Soc.S.Aust., 30, 227-262.

Howchin, W. 1926. The geology of the Barossa ranges and neighbourhood in relation to the geological axis of the country. Trans.Roy.Soc.S.Aust., 50, 1-16.

——— 1929. The geology of South Australia. 320 pp.

Hudson, F.S. 1955. Folding of unmetamorphosed strata superjacent to massive basement rocks. Bull.Am.Ass.Pet.Geol., 39, 2037-2052.

Jackson, E.D. 1961. Primary textures and mineral associations in the ultramafic zone of the Stillwater Complex, Montana. U.S.Geol.Surv.Prof.Paper No.358, 106 pp.

Jones, J.B., Talbot, J.L. and McBriar, E.M. 1962. A suite of volcanic rocks with spilitic affinities from the Archaean of South Australia. Aust.Jour.Sci., 29, 356.

Kleeman, A.W. 1954. A reconnaissance of the structural petrology of the Mt Lofty Ranges. Unpublished Ph.D. thesis, 37 pp.

——— and Skinner, B.J. 1959. The Kanmantoo Group in the Strathalbyn-Harrogate region, South Australia. Trans.Roy.Soc.S.Aust., 82, 61-71.

——— and White, A.J.R. 1956. The structural petrology of portion of the eastern Mt Lofty Ranges. Jour.Geol.Soc.Aust., 3, 17-31.

Leslie, R.B. and White, A.J.R.

1955. The "Grand" unconformity between the Archaean (Willyama Complex) and Proterozoic (Torrowangee Series) north of Broken Hill, New South Wales.
Trans.Roy.Soc.S.Aust., 78, 121-133.

Knill, J.L. 1959. The tectonic pattern in the Dalradian of the Craignish-Kilmeffort district, Argyllshire.
Quart.Jour.Geol.Soc.London, 140, 339-365.

——— 1960. A classification of cleavages, with special references to the Craignish district of the Scottish Highlands.
Int.Geol.Congr.Rep., 21, pt.13, 317-325.

——— 1961. Joint-drags in Mid-Argyllshire.
Proc.Geol.Assoc., 72, 13-19.

Knopf, E. 1931. Retrogressive metamorphism and phyllonitization.
Amer.Jour.Sci., 21, 1-27.

McBirney, A.R. and Best, M.G.

1961. Experimental deformation of viscous layers in oblique stress fields.
Bull.Geol.Soc.Amer., 72, 495-498.

Mawson, D. 1912. Geological investigations in the Broken Hill area.
Mem.Roy.Soc.S.Aust., 2, pt.4, 211-319.

——— 1926. A brief resume of present knowledge relating to the igneous rocks of South Australia.
Rep.Aust.Ass.Advanc.Sci., 18, 230-274.

Mawson, D. 1948. Sturtian tillite of Mount Jacob and Mount Warren Hastings, North Flinders Ranges.
Trans.Roy.Soc.S.Aust., 72, 244.

—— and Sprigg, R.C.
1950. Subdivision of the Adelaide System.
Aust.Jour.Sci., 13, 69-72.

O'Driscoll, E.S. 1962. Experimental patterns in superposed similar folding.
Jour.Alberta Soc.Pet.Geol., 10, 145-167.

Oriel, S.A. 1951. Structure of the Hot Springs Window, Madison County, North Carolina.
Amer.Jour.Sci., 249, 1-30.

Paterson, M.S. and Weiss, L.E.
1962. Experimental folding in rocks.
Nature, 195, 1046-1048.

Peach, B.N., Kynaston, B.A. and Muff, H.B.
1909. The geology of the Seaboard of Mid Argyll.
Mem.Geol.Surv.Scotland, No.36, 121 pp.

Pettijohn, F.J. 1957. Sedimentary Rocks, 2nd Ed.
Harper, 718 pp.

Ramberg, H. 1962. Intergranular precipitation of albite by unmixing of alkali feldspar.
Neues.Jahr.für Miner., 98, pt.1, 14-34.

Ramsey, J.G. 1960. The deformation of early linear structures in areas of repeated folding.
Jour.Geol., 68, 75-93.

—— 1962. The geometry and mechanics of formation of "similar" type folds.
Jour.Geol., 70, 309-327.

Rast, N. 1958. The tectonics of the Schichallion Complex.
Quart.Jour.Geol.Soc.London, 114, 25-46.

—— and Platt, J.I.
1957. Crossfolds.
Geol.Mag., 94, 159-167.

Rogers, J.J.W. 1961. The origin of albite in granite rocks.
Amer.Jour.Sci., 259, 186-193.

Rickard, M.J. 1961. A note on cleavages in crenulated rocks.
Geol.Mag., 98, 324-332.

Ross, J.V. 1962. The folding of angular unconformable sequences.
Jour.Geol., 70, 294-308.

Sprigg, R.C. 1945. Some aspects of the geomorphology of portion of the Mt Lofty Ranges.
Trans.Roy.Soc.S.Aust., 69, (2), 277-303.

—— Whittle, A.W.G., and Campana, B.
1951. Geological atlas of South Australia. 1" = 1 mile series, Adelaide Sheet.

Spry, A.H. 1951. The Archaean Complex at Houghton, South Australia.
Trans.Roy.Soc.S.Aust., 74 (1), 115-134.

Starkey, J. 1959. Chess-board albite from New Brunswick, Canada.
Geol.Mag., 96, 141-145.

Strand, T. 1944. Structural petrology of the Bygdin conglomerate.
Norsk Geol. Tiddskr., 24, 14-21.

Sutton, J. and Watson, J.
1958. Structures in the Caledonides between Loch Duich and Glenelg, North-West Highlands.
Quart.Jour.Geol.Soc.London, 114, 231-258.

Thomas, R.G. 1924. A monazite-bearing pegmatite from Normanville.
Trans.Roy.Soc.S.Aust., 48, 258.

Tröger, W.E. 1956. Optische Bestimmung der gesteinsbildenden Minerale.
Stuttgart, 147 pp.

Turner, F.J. 1947. Determination of plagioclase with the four-axis universal stage.
Amer.Miner., 32, 389-410.

Tuttle, O.F. and Bowen, N.L.
1958. Origin of granite in the light of experimental studies in the system $NaAlSi_3O_8 - KAlSi_3O_8 - SiO_2 - H_2O$.
Geol.Soc.Amer.Mem.74, 153 pp.

Upton, B.G.J. 1960. The alkaline igneous complex of Kungnat Fjeld, South Greenland.
Meddel.om Gronland, 123, No.4, 145 pp.

van der Kaaden, G. 1951. Optical studies on natural plagioclase feldspar with high-temperature and low-temperature optics.
Dissert.Univ.Utrecht, 105 pp.

Voll, G. 1960. New work on petrofabrics.
Liv.Man.Jour.Geol., 2, 503-567.

Walker, K.R., Joplin, G.A., Lovering, J.F. and Green, R.
1960. Metamorphic and metasomatic convergence of basic igneous rocks and lime-magnesian sediments of the Precambrian of North-western Queensland.
Jour.Geol.Soc.Aust., 6, 149-178.

Webb, B.P. 1953? Structure of the Archaean Complex of Mount Lofty Ranges. (undated) M.Sc. thesis, Univ. of Adelaide.

Weiss, L.E. 1957. Structural analysis of the Basement System at Turoka, Kenya. Overseas Geol. and Min.Resources, 7, 3-35 and 123-153.

—— 1959. Geometry of superposed folding. Bull.Geol.Soc.Amer., 70, 91-106.

—— and McIntyre, D.B. 1957. Structural geometry of Dalradian rocks at Loch Leven, Scottish Highlands. Jour.Geol., 65, 575-602.

Whitten, E.H.T. 1957. The Gola Granite and its regional setting. Proc.Roy.Irish Acad., 58, Sect.B, No.12, 245-292.

White, W.S. 1949. Cleavage in east-central Vermont. Trans.Amer.Geophys.Un., 30, 587-594.

Williams, H., Turner, F.J. and Gilbert, C.M. 1955. Petrography. Freeman, 406 pp.

Wilson, A.F., Compston, W., Jeffery, P.M. and Riley, G.H. 1960. Radioactive ages from the Precambrian rocks in Australia. Jour.Geol.Soc.Aust., 6, 179-196.

Woolnough, W.G. 1908. Notes on the geology of the Mount Lofty Ranges, chiefly the portion east of the Onkaparinga River. Trans.Roy.Soc.S.Aust., 32, 121-137.