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GRANITIC AND MIGMATITIC ROCKS OF THE COOKE HILL AREA, SOUTH AUSTRALIA AND THEIR STRUCTURAL SETTING

by

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SUMMARY

The area studied lies within the region of high grade metamorphic rocks of the Mt. Lofty Ranges. The rocks exposed include the Strangway Hill and Inman Hill Formations of the Lower Cambrian Kanmantoo Group. They have experienced two phases of ductile deformation and a third phase of brittle deformation.

A large number of small plutonic bodies of granodioritic and tonalitic composition were intruded towards the end of the second deformation.

The first generation structures are small folds with well developed axial plane schistosity (S_1). The S_1 schistosity is almost universally parallel to compositional layering (S_0) and a mineral lineation is parallel to axes of F_1 folds. During the second phase of deformation, two major structures, a synform and an antiform, were produced. The major F_2 structures are accompanied by small scale folds which have a penetrative crenulation cleavage or secondary schistosity (S_2) parallel to their axial planes. The third phase of deformation offsets the major structures along a number of small to large scale faults. During this period, the Strangway Hill Formation, now forming the eastern part of the area, was thrust over the Inman Hill Formation.

Migmatization of the quartzo-feldspathic, semi-pelitic and pelitic rocks commenced during the first phase of deformation and reached a peak during the second phase of deformation. Fibrolitic sillimanite is the only stable alumino-silicate polymorph and is generally associated with relicts of primary muscovite throughout the area. In places, fibrolite has been replaced at a later stage by skeletal muscovite.

The petrochemistry of the migmatites suggests that they have originated by metamorphic differentiation combined with some soda-metasomatism. The major and trace element chemistry of the

leucocratic veins of the migmatites is not consistent with their having been formed solely by differential anatexis of country rocks, nor with their having been formed by injection of granitic fluids from magmatic rocks such as the associated Cooke Hill intrusives, or other intrusives outside the present area.

The field relationships and chemistry of the Cooke Hill granitic intrusives suggest that they are magmatic rocks derived by anatexis of sediments of greywacke composition. The trace and major element chemistry of the adjacent country rocks indicate that the source materials for the anatectic Cooke Hill magma occur outside the present area, probably at greater depth.

Mineralogical and chemical studies of the calc-schists (actinolite/tremolite schists) and the calc-gneisses (diopside gneisses) show that they have been derived from the dolomitic greywackes. These contrast with the calc-silicates associated with the Milendella Marbles which are derived from argillaceous limestones and calcareous shales.

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, it contains no material previously written or published by another person except where due reference is made in the text.

Signed:

Syed Abul Fazlil Abbas

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CHAPTER 1

1.1. INTRODUCTION

This thesis presents a petrological, structural and geochemical study of an area of rocks covering some 26 square miles on the Mannum Sheet, north of Palmer township, and approximately 31 miles north-east of Adelaide, South Australia (Figs. 1 & 2). Geological mapping was carried out on a scale 3.7 inches to 1 mile using aerial photographs.

The rocks of the area investigated form part of the Kanmantoo Group of Cambrian age and comprise highly metamorphosed and strongly deformed sediments. Migmatization of high grade gneisses together with intrusion of a number of granitic bodies indicates a long and complex geological history during the Early Palaeozoic. Generally metamorphism reached the sillimanite grade.

1.1.1. Previous work in the Cooke Hill area

Brief reports on a few mining activities in the past are to be found in publications of the S.A. Mines Department principally in the Record of Mines of South Australia (1908), but the first geological investigation in the Cooke Hill area was carried out by White (1956) during the regional mapping of the Mannum Sheet, covering an area of about 120 square miles. His studies were mainly concerned with the petrology of the Palmer Granite and associated metasediments, as a part of his Ph.D. project. He gave a comprehensive description of some of the metasedimentary rocks including the Milendella Marbles and calc-silicates (White, 1956, 1959). However most of the rocks in the Cooke Hill area, in particular the diopside gneisses and actinolite-tremolite schists (calc-schists and calc-gneisses) were mentioned only briefly; these occupy a major part in the western portion of the area studied. Since then, no contribution has been made to the detailed petrological study of these rocks. Recently the Mines Department of the Geological Survey of South Australia has published the Adelaide Sheet on a scale of 1 inch to 4 miles (Thomson, 1969a). This map portrays the geographical distribution of the Strangway Hill Formation and the Inman Hill Formation, which constitutes two distinctive units in the area studied.

1.1.2. Purpose of investigation and nature of problem

The migmatites of the Mt. Lofty Ranges are associated with high grade metamorphics of amphibolite facies but studies on these rocks from different areas have led several workers to propose different hypotheses on their genesis (Chinner, 1955; White, 1956 and 1966a; Mills, 1964 and Offler, 1966). Generally, two different views are held among these petrologists, one suggesting that migmatites are the result of metamorphic differentiation and the other proposing the partial melting of country rock as the cause of migmatization.

Because of this controversy, it was felt desirable to carry out a detailed investigation involving structural, petrological, chemical and mineralogical studies on the migmatites and associated granitic rocks of the present area. The main concern of this thesis has been to elucidate the process by which the granitic component of migmatites has come into being. For this purpose, a detailed study has been carried out on the field relationships and petrochemistry of the leucocratic veins of migmatites together with host rocks and associated granitic intrusives.

An attempt has been made to establish as definitely as possible the geometry of the major structure and the sequence of events in the area. This was necessary to provide control of hypotheses of migmatite generation in space and time but it has also made a contribution to knowledge of the regional structure of the Kanmantoo rocks. In particular, superposition of two generations of folds on the macroscopic scale has been demonstrated for the first time in the Kanmantoo.

The petrology of the important rock groups mapped in the area has also been described. Again, in some cases, this was necessary in relation to the genesis of the migmatites. In addition however the investigation of the nature and origin of the calc-silicate rocks (actinolite-tremolite schists and diopside gneisses) has been studied. These rocks have not been described before and their stratigraphic position in the Kanmantoo Group is not entirely clear. In view of the structural geometry that was established during the present work, it was necessary to consider the possible correlation of the calc-silicate rocks on the west of the area with those to the east of the Cooke Hill Fault (the Milendella Marble and associated calc-silicate rocks).

These studies have allowed the migmatites and associated granitic intrusions to be examined in a structural and stratigraphic framework that can be related to the regional setting.

1.2. REGIONAL SETTING

The area under investigation forms a part of the eastern Mt. Lofty Ranges which consist of metasedimentary rocks that can be divided into three groups of Precambrian to Lower Palaeozoic age (Fig. 2). In Early Palaeozoic time, folding gave rise to the mountainous chain of the Mt. Lofty Ranges. Campana (1955) regarded the Mt. Lofty Ranges as a faulted anticlinorium in the core of which the crystalline basement forms elongated inliers. The oldest rocks, schists and gneisses commonly referred to as "Archaean", but probably better regarded as Lower Precambrian in age, occur as the Houghton-Barossa inlier, the Crafers-Aldgate inlier and the Mt. Compass-Yankalilla inlier. These are unconformably overlain by a vast thickness of Adelaide Geosynclinal sediments, ranging in age from Upper Precambrian to Lower Cambrian. The Precambrian sequence is known as the Adelaide Supergroup (Daily, 1963) and the overlying Cambrian consists of limestones and clastics of which the Kanmantoo Group forms the uppermost part. The Kanmantoo Group, consisting of a great thickness of mainly clastic rocks, was deposited in the Kanmantoo Trough. These sediments form much of the eastern part of the Mt. Lofty Ranges and in this area they are folded and regionally metamorphosed up to sillimanite grade. Covering these folded rocks in places there are glacial and post-glacial deposits of Permian age, including boulder tillites and cross-bedded sands; remnants of these are to be found especially in the southern Mt. Lofty Ranges, e.g. at Hallett Cove and Inman Valley. The Cooke Hill area is bounded on the east by the Murray Plain where Tertiary marine sediments abut against the metamorphosed rocks of the Kanmantoo Group. A brief description of the Precambrian basement rocks, the Adelaide Supergroup and the Kanmantoo Group is given below in order that the particular problems of the Cooke Hill area may be appreciated.

1.2.1. The Precambrian basement rocks

The basement rocks generally consist of metamorphosed sediments and igneous rocks, ranging in character from schists to gneisses.

The crystalline rocks, which are strongly metamorphosed, were regarded as igneous by Benson (1909). He studied the basement rocks near Houghton and named them "Houghton Diorite". Howchin (1904, 1906) studied the western part of the Mt. Lofty Ranges and regarded the Houghton rocks as an Archaean basement east of his unmetamorphosed Mt. Lofty Series (now the Adelaide Supergroup). Woolnough (1908) examined a section across the eastern part of the anges and proposed the name "Barossa Series" or "Barossian" for the metamorphic rocks which are now placed in the Adelaide Supergroup and in the younger Kanmantoo Group. Howchin (1926) recognized that the rocks described by Woolnough were largely the equivalent of his Mt. Lofty Series and therefore referred to the underlying Houghton basement rocks as the "Fundamental Complex or Houghtonian". However Hossfeld (1935) considered that Woolnough's term had priority for the Houghton rocks and therefore he referred to the basement inlier as the Barossian Complex. This term has since been used by Sprigg, Whittle and Campana (1951), Campana (1955) and Spry (1951)etc. Talbot (1962) thoroughly re-investigated the basement rocks of the Houghton area and essentially returned to Howchin's terminology by suggesting the name Houghton Complex for the rocks of the basement inliers.

1.2.2. The Adelaide Supergroup

The basement complex is unconformably overlain by the Adelaide Supergroup. Clarke (1938) was the first to use the term the Adelaide System. Later, it was used by Mawson (1948) and David (1950); Mawson and Sprigg (1950) formally defined the term. In type sections near Adelaide, Mawson and Sprigg (op.cit.) subdivided the system into three series known as the Torrensian Series, the Sturtian Series and the Marinoan Series. Since then, these terms have been widely used by South Australian geologists. Recently, Daily (1963) reviewed the terminology and suggested abandoning the Adelaide System and its three series, due to unwarranted time significance, and replacing them with the Adelaide Supergroup and the Torrens Group etc., thus following the recommendation of the American Code of Stratigraphic Nomenclature (1961). This suggestion is accepted here although it has not been adopted by the Geological Survey of South Australia (Parkin, 1969).

More than two thirds of the Mt. Lofty Ranges are occupied by the Adelaide Supergroup, covering an area of 12,000 square miles (Parkin, 1969). In most instances these rocks are relatively unmetamorphosed, with the exception of those in the eastern and southern Mt. Lofty Ranges. The thickness of the Adelaide Supergroup near Adelaide is estimated to be 28,000 feet (Mawson and Sprigg, 1950).

These rocks in the western Mt. Lofty Ranges were studied by Howchin (1926, 1929), Sprigg (1946, 1952), Mawson and Sprigg (1950), Sprigg, Whittle and Campana (1951) and Sprigg and Wilson (1954). Generally, the Torrens Group, Sturt Group and Marino Group are exposed in the western Mt. Lofty Ranges. Further south, in the Fleurieu Peninsula, Daily (1956, 1963, 1969), and Thomson and Horwitz (1961) showed that the Marino Group is overlain by Lower Cambrian sediments. On the Milang Sheet, purple and green shales have been mapped on the top of the Marino Group by Horwitz and Thomson (1960) who have proposed a slight unconformity between the Marino Group and the overlying sediments. Daily (1963) suggested that the Mt. Terrible Formation rested conformably on the Marine Group in the Stockyard Creek and Sellick Hill areas. He placed the Precambrian-Cambrian boundary within the Mt. Terrible Formation a short distance below the Wangkonda Formation which contains a fauna including Hyolithes. However, later Daily (1969) indicated in his Table 1 that the Mt. Terrible Formation is transgressive and unconformable on the Adelaide Supergroup. The Brighton Limestone near Adelaide is placed at the base of the Marino Group by Thomson and Horwitz (1962). The Delamere Marbles contain Lower Cambrian fossils (Daily, 1963) and consequently have been equated with Lower Cambrian rocks of the Sellick Hill-Normanville area. These occur immediately below the base of the Kanmantoo Group. A brief summary of the rocks of the Adelaide Supergroup is given below:

Torrens Group

The Torrens Group consists of shallow water to deltaic deposits. Cross-bedded and heavy-mineral banded conglomerates, sandstones, mudstones and dolomites are typical rock types. Transition from sedimentary magnesites into dolomitic beds is a characteristic feature of this Group. Thickness varies considerably from a few feet to

3,000 feet. It unconformably overlies the "Archaean" basement in the Mt. Lofty Ranges although the contact between the two is mostly a faulted one.

Sturt Group

The sediments in this Group are highly variable in nature. Common rock types are slates, phyllites, quartzites, limestones and tillites. In the type section of Adelaide, the Belair Slates and Quartzites together with the Mitcham Quartzites are estimated to be up to 1,000 feet thick. They constitute the lowest beds in the succession. The Sturt Tillite, about 1,000 feet thick, conformably overlies these rocks. The Tapley Hill Slate consists of calcareous, well laminated, bluish coloured rocks and conformably overlies the Sturt Tillite. Blue or buff coloured Brighton Limestone, oolitic in places, forms the top of this sequence. In the Stockyard Creek section, black laminated phyllites have been equated with the Tapley Hill Slate by Daily (1963). The phyllites are calcareous but pass gradually to carbonate rock which is correlated by Daily (1963) with the Brighton Limestone. The majority of workers believe that the Sturtian conformably overlies the Torrens Group in the Mt. Lofty Ranges. Unconformity between the two groups in the Flinders Ranges appears to be widespread.

Marino Group

The Marino Group conformably overlies the Sturt Group and consists of shallow water deposits. Typical rocks are purple, grey and greenish shales and slates, with siltstones, quartzitic sandstones and intercalated dolomites.

Unmetamorphosed sediments gradually become metamorphosed in the eastern Mt. Lofty Ranges (Fig. 3). Metasediments are represented as mica schists at Williamstown, and slates and dolomites in the Mt. Crawford area. Elsewhere, carbonates have been changed to marbles, marly shales and calcareous sandstones to calc-silicate rocks: and epidote hornfelses; and argillites to garnet schists. The Sellick

Hill-Normanville fossiliferous sequence (Abele & McGowran, 1959; Daily, 1963) lies between the Adelaide Supergroup and the Kanmantoo Group.

1.2.3. Kanmantoo Group

Most of the eastern Mt. Lofty Ranges and a major portion of Kangaroo Island is underlain by Kanmantoo Group sediments. The term Kanmantoo Group was first used by Sprigg and Campana (1953) who described its age and facies in the eastern Mt. Lofty Ranges. The Group consists of a thick sequence of non-fossiliferous sediments, generally greywackes, arkoses, pelites and calcareous rocks (Sprigg and Campana, op.cit.; Thomson, 1960). The Kanmantoo Trough is believed to have been formed during the tectonic movements when the southern and the eastern sea floor of the Adelaide Geosyncline collapsed along submarine faults. This tectonic episode has been called the Waitpinga Subsidence and took place during early Cambrian time (Thomson, 1969b). An enormous volume of sediments was deposited in the Trough during the subsidence (Fig. 3). Rocks of this Group form an arcuate belt stretching from Kangaroo Island to the eastern Mt. Lofty Ranges. The apparent thickness of the Kanmantoo Group sediments has been given as about 60,000 feet, estimated from surface bedding dips (Thomson, op. cit.), although this thickness appears to be overestimated due to repetition of beds during polyphase folding. Sprigg and Campana (1953) suggested that the base of the Kanmantoo Group might be separated from the Adelaide System by the Nairne fault. Later Campana and Horwitz (1955) regarded the Nairne fault as a local movement and considered that the Kanmantoo Group unconformably overlies the Adelaide System.

Kanmantoo Group sediments suffered folding and regional metamorphism to give rise to a variety of schists, gneisses, migmatites, marbles and calc-schists-gneisses. In places, syn- and post-tectonic acidic rocks and minor basic rocks have been intruded into the metasediments. Generally, the grade of metamorphism ranges from chlorite grade to sillimanite grade. The peak of metamorphism culminates in the formation of migmatites in sillimanite zone rocks (Fig. 4). Four metamorphic zones have been distinguished in the eastern Mt. Lofty Ranges by detailed studies of White (1956, 1966a), Mills (1964), Offler (1966), Offler and Fleming (1968), and Fleming (1971). They are the (1) Chlorite and Biotite Zone, (2) Andalusite and Staurolite Zone, (3) Kyanite Zone and (4) Sillimanite Zone. Garnet occurs sporadically in the biotite zone and therefore a separate zone cannot be assigned to this index mineral.

Following the system adopted by the Geological Survey of South Australia (Adelaide Sheet S 1 54-9, compiled and revised by B.P. Thomson, 1969a) the Kanmantoo Group can be subdivided into three units in the eastern Mt. Lofty Ranges, viz.

Top	Brukunga Formation
	Inman Hill Formation
Base	Strangway Hill Formation

The Strangway Hill Formation

This is the lowest unit in the Kanmantoo Group succession. The typical rocks of this Formation are phyllites, schists, gneisses, marbles and calc-silicates. In the eastern Mt. Lofty Ranges, between the Barossa Valley and Kanmantoo, high grade metamorphic rocks of sillimanite grade are commonly found in this unit. Grey to pink coloured marbles with associated calc-silicates are characteristic. These calcareous rocks have been called the Milendella Marble Member (Thomson, 1969b). In the Delamere area, the Carrickalinga Head Formation of Lower Cambrian age is the base of the Kanmantoo Group (Daily, 1963,1969). The Macclesfield and Paris Creek marbles which lie about 1,100 feet below the Nairne Pyrite Formation (Kleeman & Skinner, 1959) may be correlated with the Milendella Marble of the Strangway Hill Formation. The thickness of this Formation varies considerably mostly ranging from 2,000 to 4,000 feet. The base of this Formation is not present in the area mapped by the author.

The Inman Hill Formation

This formation conformably overlies the Strangway Hill Formation. Intertonguing with the underlying Formation is a characteristic feature of this unit (see Fig. 3). In its type area in Inman Valley, the apparent thickness of the metasediments is about 15,000 feet. The original sediments were generally composed of greywacke to arkosic type material with negligible carbonate. Sedimentary structures such as ripple marks, cross-bedding, slump folds and scour and fill structure are preserved.

The Inman Hill Formation probably attains the greatest thickness in the eastern Mt. Lofty Ranges but appears to be thinning in the southern Ranges. The well crystallized sediments are generally represented by quartzo-felspathic schists, gneisses, migmatites and calc-schists and calc-gneisses. The per-aluminous shale recrystallizes to mica schists containing andalusite, staurolite, kyanite, garnet and sillimanite.

The Brukunga Formation

The youngest unit of the Kanmantoo Group is the Brukunga Formation which lies directly above the Inman Hill Formation. The basal marker bed of this unit is the Nairne Pyrite. This formation is represented by medium grade metamorphic rocks with the maximum thickness, about 24,000 feet, in the eastern Mt. Lofty Ranges. The common rocks are phyllites, schists and minor carbonate near Tungkillo. A lenticular body of the Rathjen Gneiss probably forms part of this unit grading into surrounding country rocks of the Brukunga Formation (see White, 1966b). The age of the Brukunga Formation is considered to be Middle Cambrian (Thomson, 1969a,b).

The age of the Kanmantoo metasediments has been debated among geologists due to lack of fossils and the crystalline nature of the rocks. Sprigg and Campana (1953) first differentiated the Kanmantoo Group from the Adelaide Geosynclinal sediments and gave it a Cambrian to Ordovician age. The base of the Kanmantoo Group in the Southern Mt. Lofty Ranges is the Carrickalinga Head Formation which contains Lower Cambrian fossils (Daily, 1963, 1969). Crystalline sediments of the eastern Mt. Lofty Ranges and of the Victor Harbour area enclose abundant intrusive granites. The radiometric age datings of these granites ranges from 420 m.y. to 490 m.y. Potassium argon dates on biotite from Victor Harbour Granite range from 420 m.y. to 440 m.y. (Evernden & Richards, 1962). Lead-alpha dates on zircons from Victor Harbour Granite give 460 m.y. (Fander, 1961). Rb/Sr dates on the Palmer Granite give 490 ± 15 m.y. (White, Compston & Kleeman, 1967).

1.3. METAMORPHISM AND OROGENESIS

The older basement rocks of the Mt. Lofty Ranges are exposed as isolated inliers in the Adelaide Geosynclinal sediments (England, 1935; Alderman, 1938; Spry, 1951 etc.). They show strong regional metamorphic features. Rock types of amphibolite facies sometimes retrograded to chlorite facies are recognized. Thus a complex geological history involving folding, faulting and metamorphism took place before the deposition of the Adelaidean sediments.

The younger sediments which are lying on the basement rocks with marked unconformity show a pronounced change in their metamorphism and tectonism. The tectonic movements which gave rise to the mountainous chains of the Mt. Lofty Ranges are called the Delamerian Orogeny (Thomson, 1969b). This orogeny probably started in Upper Cambrian time and folded and regionally metamorphosed the Adelaidean and the Kanmantoo Trough sediments. The climax of metamorphism was reached in the eastern portion of the Ranges where high grade rocks with large scale migmatization developed. Structural events of the Mt. Lofty Ranges have been summarized by Offler and Fleming (1968) and Fleming (1971). They showed that there are at least four successive phases of deformation forming a complex fold and fault pattern. During the Tertiary period, block faulting uplifted the Mt. Lofty Ranges during its final tectonic movements. There is evidence that some of the Pre-existing faults of Palaeozoic time were active during this period (Sprigg, 1946; Glaessner, 1953; Webb, 1958; Walpole, 1962).

Gentle doming and upward arching of the Ranges resulting from movement along ancient fault planes in Tertiary to Recent time may be responsible for most of the present elevation.

CHAPTER 2

STRUCTURE AND STRATIGRAPHY

2.1. DISTRIBUTION OF MAIN ROCK UNITS AND SEDIMENTARY STRUCTURES

Almost all of the metasediments exposed in the area studied belong to the Kanmantoo Group. According the the stratigraphic scheme outlined in section 1. 2. 3. they can be allocated to the Strangway Hill Formation and the Inman Hill Formation (Fig. 3). The detailed stratigraphy of the area is discussed in section 2.11 after the structure has been established.

A north-south trending escarpment rising abruptly from the Murray Plain forms the eastern limit of the area. This significant physiographic feature of the region is due to the Milendella Fault along which vertical movements have occurred during the Tertiary. The Strangway Hill Formation crops out in the eastern escarpment along a north-south trending thrust fault, here named the Cooke Hill Fault. The most common rock types are quartz feldspar schists, gneisses, impure marbles, calc-silicate rocks, semi-pelitic and pelitic schists, which do not show primary structures of undoubted sedimentary origin such as are sometimes seen in less metamorphosed and migmatised parts of the Kanmantoo Group. However the layering is here interpreted as bedding, preserved in more or less modified form (cf. Dietrich 1960a, b) rather than the product of metamorphic differentiation (Turner, 1941; 1948; Katz, 1970).

The quartzo-feldspathic, semi-pelitic and pelitic schists and gneisses all have compositional layers appearing as light and dark bands due to variation in biotite content (Plate 1A & B). The thickness of layers is variable, from a fraction of an inch to a few inches, but they are uniform and continuous for long distances. Generally, the imposed schistosity of the first deformation is parallel to compositional layering but it crosses the layering in the hinges of small folds. The second deformation produced a weak schistosity which is at a slight angle to layering. It does not seem possible therefore that the layering is produced by metamorphic differentiation parallel to either of the observed schistosities. Sometimes, heavy mineral concentrations occur in thin layers rich in biotite and grains have anhedral shape frequently with long axes generally parallel to the compositional layering of the rocks. These observations therefore strongly suggest that most of the layering in the metasediments arose from original stratification layering in sediments consisting of alternations of arkose and shale. This is also supported by the concordance of interbedded calcareous sediments represented by marbles and calc-silicates in the rocks of the Strangway Hill Formation.

There are numerous marbles intercalated with the schists. Their massive nature resists mechanical weathering so that they form prominent ridges. The layers of marble vary in thickness but rarely exceed two hundred feet. Calc-silicate and quartzo-feldspathic seams within the marbles apparently represent original intercalations of marly and quartzo-feldspathic material, though it is difficult to rule out some contribution by metamorphic differentiation. The marbles sometimes can be observed to grade along strike into less calcareous calc-silicate rocks within the enclosing quartz-feldspar schists and gneisses, which also contain isolated lenses of calc-silicate rocks representing original calcareous shale developments. White (1956) had named these rocks informally the "Milendella marbles and calc-silicates." The lateral variability of these rocks contrasts with the regularity and continuity of layering in the quartzo-feldspathic rocks. The author, therefore, concludes that although the layering in the latter appears to be based essentially on original bedding it may have been enhanced by deformation and metamorphic differentiation.

The area between the Palmer Fault zone and the north-south trending thrust plane is occupied by the Inman Hill Formation. The main rock types are quartzo-feldspathic schists-gneisses, migmatites, granite gneisses, semi-pelitic/pelitic schists, actinolite-tremolite schists, diopside gneisses and minor aluminous pelitic schists. Migmatization plays a significant role, forming abundant leucocratic veins in quartz feldspar schists and gneisses. The coarse-grained diopside gneisses occur in two lenticular bodies about three miles and five miles long respectively. An intertonguing relationship with the country rock has been observed with gradual changes in their mineral assemblages. The diopside gneisses grade laterally through a medium grained zone into adjacent actinolite schists and biotite gneisses which separate them from the normal quartzo-feldspathic schists (Map 1). Generally, the actinolite schists are friable rocks; therefore, mechanical weathering has deeply eroded them and in the valleys the schists are either covered under the soil or exposed in isolated outcrops (Plate 2). These schists are clearly directly associated with the diopside gneisses; however, the contact with the gneisses is sharp and concordant and can be traced out continuously along its trend-line. Actinolite schists are intercalated with the tremolite schists, ranging in thickness from a few inches to a couple of tens of feet. Compositional layering is well developed in the schists due to concentration of amphibole and quartz-feldspar in alternate layers. Actinolite-tremolite schists show a pronounced schistosity and lineation due to parallel preferred orientation of amphibole prisms. The gradational changes and interfingering relationships between the diopside gneisses and the actinolite-tremolite schists suggest a facies change in original sediments. On the basis of the field observations and their mineral assemblages, it is concluded that these rocks are recrystallized dolomitic sandstones and dolomitic shales.

Quartzo-feldspathic schists and gneisses in the Inman Hill Formation are well-banded rocks characterized by micaceous layering. Schistosity is generally parallel to layering. A few slump structures and some cross-bedding has been observed in the northern part of the area where the quartzo-feldspathic schists are less affected by strong metamorphism. Compositional layers are remarkably uniform and can be traced out in the field for several hundred yards along the strike. In the coarse grained gneisses, layering is widely spaced but in fine grained schists, it appears as thin laminations. The rock breaks along the micaceous layering. Occasionally, heavy mineral concentration in the micaceous layering provides evidence for sedimentary depositional features. In hand specimen, the compositional layering appears as dark and light bands of uniform thickness. This layering is again interpreted essentially as deformed bedding (Dietrich, 1960a, b; Tucker, 1960).

The granitic gneisses are well exposed near the Cooke Hill triangulation station. The road cuttings provide excellent exposures of these rocks. The gneisses outcrop in a lenticular body, less than half a mile long. They can be easily recognized in the field by the typical tor-like outcrops. Generally the texture is variable from coarse grained to medium grained and the rocks are usually pink in colour. The gneisses are strongly lineated due to parallel orientation of biotite and elongate coarse quartz grains. The contact is conformable with the country rock in the few exposed sections. Previously these gneisses were mapped as an intrusive body due to their granitic appearance and texture (White & Thatcher, 1957). The rocks greatly resemble the Rathjen Gneiss (White, 1966b) in hand specimen. The lack of xenoliths, conformable contacts with the country rocks and gradational features of the gneisses do not support the igneous origin. This field evidence rather suggests that these are re-crystallized sediments or tuffs, as suggested by White (1966b). The very strong linear fabric is a puzzling feature however and is further discussed in Chapter 4.

One of the notable features of the Inman Hill Formation lithological units is the abundance of leucocratic granitic veins in the gneisses, here collectively described as migmatites (Chapt. 5). The degree of migmatization increases from north-west to south-east parallel to the increasing grade of metamorphism indicated by the breakdown of the quartz-muscovite assemblage to sillimanite K-feldspar in the host rocks.* Generally, the leucocratic veins of the migmatites are conformable with the host rocks. The migmatites appear to be stratigraphically equivalent to meta-arkoses of the Cambrai area (see Map 3). The sedimentary structure is rarely visible in the migmatitic rocks due to increase of grain size during recrystallization of metasediments. The layering is well developed and characterized by abundance of oriented biotite and often by concentration of heavy minerals in the biotite rich lamination of the host rock. As pointed out before, regularity and continuity of the layering reflects original bedding.

2.2. STRUCTURAL STYLE AND HISTORY

INTRODUCTION

The structural investigation was carried out to analyse the phases of deformation and to establish the original stratigraphic succession of the metasediments. All linear and planar structures were measured in the field (Map 2) and have been plotted on equal area, lower hemisphere, stereographic projections. One major synform and one antiform have been recognized in the area and there is also a series of faults which resulted during the last phase of deformation.

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* In Maps 1 & 2 a boundary has been drawn across the quartzo-feldspathic rocks in the Inman Hill Formation dividing the Formation into two portions according to abundance of migmatite veins. Two different colours have been used on these maps to indicate the quartzo-feldspathic gneiss with plentiful migmatite veins (light violet) and quartzo-feldspathic schists with few migmatite veins (dark violet). The rocks on either side of the boundary are similar in composition and there is no break in the stratigraphic succession.

Two phases of ductile deformation have been recognized, an earlier phase (D_1) and a later phase (D_2). There are differences in the style of folding between first and second generation structures. The mesoscopic first generation folds (F_1) are isoclinal to asymmetric similar folds with a strong penetrative axial plane schistosity (S_1) (Plates 3-7). The second generation folds (F_2) are generally open with crenulation cleavage or secondary schistosity (S_2) parallel to their axial planes. The S_2 is weakly developed on a regional scale (Plates 8-10). The orientation of F_1 and F_2 fold axes and associated lineations are not significantly different from each other. Similar observations have been recorded in the Cambrai area by Mills (1964) and the Dawesley-Kanmantoo area by Fleming (1971). The relative paucity of the earlier folds in the present area suggests that they may have been extensively destroyed or transposed by later generation folding. In places, refolded structures have been mapped (Fig. 7; Plates 12 and 13). The last phase of tectonic events, D_3, is a brittle deformation which gave rise to a series of faults. The major structures were affected during the third phase and cut in places by faults. The closure of the main synform was sliced into two parts, now appearing as two synformal structures separated by a narrow fault zone (Fig. 9; Maps 1, 2 & 3). Previously, due to lack of detailed mapping, the fault was not detected and these structures were interpreted as two separate synclines. Generally, outcrops are poor in the vicinity of faults, and they could not be revealed by the mapping on the regional scale by White (1956) and White and Thatcher (1957). However, recent mapping of the Geological Survey of South Australia has also confirmed the presence of these faults (Thomson, 1969a).

2.3. FOLDING

INTRODUCTION

During the course of regional mapping and petrological study, White (1956) gave a brief account of the structure of the present area. He collected some data for minor fold axes and mineral lineations in the field. The apparent parallelism of these structures on stereographic projections led him to conclude that there was only one period of deformation (see also Kleeman & White, 1956).

After the detailed structural studies of the Kanmantoo Group metasediments in the eastern Mt. Lofty Ranges, it became obvious there was more than one period of deformation. Important contributions in this regard were made by Mills (1964), Fleming (1965), Offler (1966), Offler and Fleming (1968) and Fleming (1971). In the Cambrai region, just north of the present area, Mills (op.cit.) has described three periods of deformation, although minor structures belonging to these three episodes were not separated on his map. In the Pewsey Vale area, Offler (op.cit.) indicated three generations of folds though evidence for distinction between F_2 and F_3 folds was not convincing. Fleming (1971) has demonstrated three groups of folds in the Dawesley-Kanmantoo area. His first generation folds have axial plane schistosity representing the first period of deformation, while two other groups, F_2 and F_3, have crenulation cleavages with two different orientations and different micro fabrics; Fleming showed that one overprints the others and that F_2 and F_3 therefore represent another two periods of deformation. None of these authors was able to demonstrate superposition of folds on the macroscopic scale but this has been possible in the present area.

2.3.1. Nomenclature

Description of structural features such as structural surfaces, fold axes, planar and linear structures etc. follows the practice of Ramsay (1967), Turner and Weiss (1963), Fleuty (1964) and Means (1962, 1966), Weiss and McIntyre (1957) as outlined below.

(a) Structural surfaces

S_0 lithological layering (relict bedding)

S_1 axial plane schistosity of F_1 defined by preferred orientation of mica and amphibole

S_2 crenulation cleavage or secondary schistosity associated with F_2

(b) Form surfaces

Any recognizable surfaces, i.e. stratigraphic, transposed bedding etc., defining the gross form of a fold (**Turner & Weiss,** 1963).

(c) Fold groups and generations

F_1 folds in which the form surface is S_0 with axial plane schistosity as S_1. Away from fold hinges S_1 and S_0 are sensibly parallel

F_2 folding of schistosity S_1 with crenulation cleavage or secondary schistosity, S_2 as axial plane. Where an axial plane schistosity is developed it is usually limited to the hinge zones (fig. 6D).

(d) Scale

Micro - thin section

Meso - hand specimen or outcrop

Macro - map scale

The observed relationships are summarised in the following table.

Period of deformation	Phases of folding	S - surfaces as penetrative structures	L - Lineations
D_1 : ductile deformation	F_1 : folds in which the form surface is S_0	S_1 : axial plane schistosity to F_1 folds	L_1 : lineation defined by intersection of S_0/S_1 surfaces
D_2 : ductile deformation	F_2 : folds in which the form surface is S_1 (and S_0)	S_2: crenulation cleavage or secondary schistosity as axial plane to F_2 folds	L_2 : lineation defined by intersection of S_1/S_2 surfaces
D_3: brittle deformation (faulting)	No folding	No penetrative structures	No linear structures

2.3.2. Evidence for polyphase deformation

As indicated above the superposition of S_2 on S_1 provides the main basis for allocating the two morphological fold groups to two generations.

It is recognised that deformation may in fact have been continuous from F_1 to F_2 but it is believed, on the basis of the following criteria, that F_2 structures are generally younger than F_1.

(1) Form surface of F_1 is bedding (S_0) only, for F_2 is first schistosity (S_1) and transposed bedding (S_0) (Fig. 5; Plates 3-7).

(2) The presence of axial plane schistosity (S_1) for F_1 and secondary schistosity or crenulation cleavage (S_2) superposed on S_1 for F_2 (Figs. 5 & 6; Plates 3A & B, 4C & D, 9A & C, 10A, B & C).

(3) F_1 structures such as axial plane schistosity (S_1) and axes of folds, rodding and mineral lineations are refolded by F_2 structures (Fig. 7; Plates 12, 13 and 17A).

(4) F_1 folds are generally moderately to tightly appressed, F_2 folds are of more open type (Fig. 6; Plates 8 & 9).

(5) Orientation of axial plane of F_1 is irregular due to refolding, but for F_2 is more constant.

These criteria provide unambiguous distinction between F_1 and F_2 folds in many cases, but since a new schistosity can be developed in F_2 folds it is often not possible to distinguish S_1 from S_2 where minor folds are absent. It is believed, however, that since S_2 schistosity tends to be limited to fold hinges, the regional schistosity is essentially S_1.

In the sections that follow the styles of the three main deformation episodes are described first.

2.4 FIRST PHASE DEFORMATION D_1

INTRODUCTION

F_1 folds are uncommon and restricted in their occurrence due to a strong overprinting effect of the later deformation. Most commonly, they are found as microscopic to mesoscopic scale folds especially in the limbs of the major F_2 synform. No macro structure of first generation folding has been certainly recognized. One of the most striking features of the first deformation is the development of a strong regional schistosity (S_1) which is visible in almost every outcrop in the field. This indicates that the earliest phase of folding was essentially ductile, and planar and linear structures were well developed by structural readjustment and mineral reconstitution by an intense tectonic deformation.

2.4.1. Microscopic structures

Minor F_1 folds are common on the microscopic scale in micaceous rocks. The style of folding varies from intrafolial to slightly asymmetric type (Fig. 5). In marbles and calc-silicate rocks, folds are less common and their general shape is isoclinal to similar type. The form surfaces of micro folds are light and dark laminations characteristically rich in calcite and diopside-amphibole contents respectively. Micro folds are less common in actinolite-tremolite schists and rare in diopside gneisses.

S_1 schistosity is the most strongly developed planar surface in all types of rocks. It is defined as axial plane schistosity of F_1 folds by preferred orientation of platy minerals and dimensional orientation of quartz and plagioclase. In thin sections, schistosity passes through the hinges of folds essentially parallel to axial surfaces. Sometimes, if the folded rock is composed of incompetent and competent layers, i.e. pelitic and arkosic layers, schistosity is strictly parallel to the axial plane in the pelitic layer but slightly refracted in the arkosic layer. Similar relations have been described by Poole (1969, p. 13) from south of the Kanmantoo area. In some micaceous layers, elongate quartz grains and plagioclase are oriented parallel to the axial planes of folds. Undulatory extinction in quartz is a common feature of all deformed rocks. In calcareous rocks flattening or elongation of calcite grains marks the schistosity plane S_1.

2.4.2. Mesoscopic structures

Generally, mesoscopic folds are more common in micaceous rocks, wavelengths ranging from a few inches to a few feet. The fold style varies from isoclinal to asymmetrical similar type (Plates 3-7). Generally the folds are moderately to tightly appressed.*

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* It is a common observation that most of F_1 folds have thickened hinges and thinned limbs (Plates 3-7). Sometimes the thinning of one limb is greater than the other limbs (Plate 5A). Frequently, quartz veins in schists and gneisses have been deformed during F_1 (Plate 6B).

The most prominent structural feature is the penetrative schistosity S_1. Schistosity is defined by strong dimensional orientation of biotite, amphiboles, quartz and feldspar. Because of the dominant isoclinal style of the folds in metasediments, the lithological layering is generally parallel to the S_1 schistosity. Thus, though it is believed that most of S_1 is developed as an axial plane schistosity, some may be partly inherited from original bedding plane schistosity. Strong allignment of platy biotite flakes or of amphibole prisms within the schistosity plane makes up the prominent lineation (L_1). Sometimes quartz veins and strong siliceous lithological units rupture along the hinges during tight isoclinal folding and form elongated rods parallel to the mineral lineation L_1 (Wilson, 1952, 1953). In coarse grained gneisses (e.g. diopside gneiss) which lack platy minerals the S_1 plane is not well defined but dimensional orientation of quartz-feldspar aggregates defines L_1. There are lineations in marbles caused by flattening of calcite and dimensional orientation of silicate minerals. Due to the virtually identical orientation of F_1 and F_2 folds, the unambiguous distinction of L_1 and L_2 lineations is not possible (cf. Hossack, 1969). Most of the lineations measured have therefore been placed in an undifferentiated group (L) but it is believed that they are mainly of F_1 age and are associated with the dominant S_1 (cf. Hooper & Gronow, 1970; Hossack, 1968).

An interesting feature of the first deformation is the structural control of migmatization. There are evidences of the mobility of granitic material which is emplaced along the axial surfaces of F_1 folds (Plates 4A, 5C & 6A). The migmatites consisting of granitic veins and gneissic host show general concordance with the S_1 schistosity.

2.4.3. Macroscopic structures

The recognition of any major F_1 folds is clearly crucial to the interpretation of the stratigraphy of the Kanmantoo Group. A major F_1 syncline and complementary anticline have been postulated by Mills in the less metamorphosed Cambrai area, north of the present area (Fig. 9). It thus appears that the present area is on the overturned western limb of an F_1 anticline so that the beds should young to the west. This agrees with Thomson's interpretation of the area west of the Palmer Fault as a major F_1 syncline containing the Brukunga Formation (although it is regarded as F_2 by Fleming and Offler, 1968). It therefore seems unlikely that the present area contains a major F_1 structure.

It should be noted, however, that the northern termination of the main outcrop of calc-schists and calc-gneisses in the Inman Hill Formation appears to be the hinge of a tight fold with northerly plunge and with axial plane schistosity. It is not certain that this schistosity is S_1 rather than S_2 although the tightness of the fold suggests this, but in any case the succession cannot be regarded as one younging simply to the west.

It is also possible that the major structure west of the present area is of F_2 age rather than F_1 and even that the major syncline recognized by Mills (1964) is F_2 rather than F_1. The structural and stratigraphic implications of this are discussed further in a later section.

2.5. SECOND PHASE DEFORMATION - D_2

INTRODUCTION

The most conspicuous folds occurring in the present area are F_2 folds. Both mesoscopic and microscopic folds occur and there is no clear break in scale between them. The hinge zone of the main synform, here called the Cooke Hill Synform, is made up of a number of intermediate scale folds whose enveloping surface describes a very open structure.

The hinge zones of the major structures and most of the minor folds are confined to the quartzo-feldspathic and pelitic rocks which have behaved in a ductile manner. The diopside-gneisses, marbles, calc-silicate rocks etc. are confined to the limb of the major structures in this area, and only a few minor folds are found in the calcareous rocks. It appears therefore that the calcareous rocks may have behaved in a relatively competent manner during the second folding. (This assumes that the fold in the main calc-schists-gneiss outcrop is of F_1 age as suggested on p. 24).

The form surface of F_2 folds is S_0 - S_1 with crenulation cleavage in pelitic rocks and secondary schistosity S_2 commonly developed in quartzo-feldspathic rocks. In the marbles, calc-silicate rocks and granitic gneisses etc., small-scale folds are less common and schistosity S_2 is rarely developed. No fracture cleavage occurs in association with the S_2 schistosity and crenulation cleavage, indicating that the more competent rocks were still ductile during this phase of deformation.

2.5.1. Microscopic structures

In thin section, the planar surface is characterized by the presence of two types of cleavages associated with minor folds. The first type is crenulation cleavage*, which is superposed on the S_1 schistosity. This cleavage is not commonly found on the regional scale but is strongly developed in pelitic rocks (Plates 10 & 16A). In quartzo-feldspathic gneisses, coarse aggregates of quartz and feldspar restrict the development of penetrative cleavage, and long flaky biotites form a crude crenulation by simple bending (Plate 17B). In some gneisses, if stumpy mica porphyroblasts are present, kinking occasionally develops as narrow kink planes at a high angle to pre-existing S_1 schistosity (Plate 18A). The presence or absence of this penetrative cleavage therefore appears to be mainly controlled by differences in lithology and texture of the deformed rocks.

The second type of cleavage is secondary schistosity S_2 which is parallel to the axial plane of F_2 folds (Plates 14A, 15A, B & 16C). The form surfaces of F_2 folds is S_0 containing an older schistosity S_1. The planar schistosity S_2 is marked by preferred orientation of biotite, more strongly developed in the hinges than the limbs of folds (Plates 15 & 16C). There is evidence that S_2 cuts early folds at a high angle to the axial surfaces (Fig. 7 A, B; Plate 17A). The secondary schistosity varies considerably in its development even on the microscopic scale, depending upon the composition of lithological units (mainly rich in quartz and feldspar) and texture of rocks. It has been observed that crenulation cleavage often grades into secondary schistosity in anisotropic rocks consisting of alternating pelitic and arkosic bands. Similar features have been reported from other parts of the world (White, 1949; Knill, 1960; Talbot, 1964). S_2 schistosity in the marbles and calc-silicate rocks is not definitely distinguishable from S_1 due to general lack of F_2 folds but it seems unlikely that the S_1 schistosity has been substantially modified in these rocks. However twinning and kink structures are very common in calcite. The kink bands are at moderate to high angle with the visible S_1 schistosity. Sometimes, micro folding has been observed by deformed twinning planes of calcite and diopside in calcareous rocks (Plate 19A, B). No evidence is available to demonstrate that such minor structures in calcite and diopside have developed during the second deformation. It is also possible that such structures may have been formed during the brittle deformation of D_3 (See Griggs et al., 1953; Turner et al., 1954; Raleigh & Talbot, 1967).

2.5.2. Mesoscopic structures

Mesoscopic F_2 folds are more common in arkose-gr ywacke metasediments (Plates 8 & 10) due to the abundance of biotite which facilitates the deformation. Generally, folds are open to moderately appressed with axial plane schistosity S_2 mostly dipping towards the west. The dominant type of folds is similar in style. In general, thickening takes place in hinges while limbs are more or less uniform. Less common types of folds are concentric to disharmonic in the migmatitic zone rocks. The plunge of F_2 folds varies considerably, ranging from less than 5° to 70°, but most of them range from 15° to 35° trending north-west to south-east (Fig. 10F).

Lineations L_2 are defined by dimensional orientation of biotite in the S_2 plane (Plate 11B) or by micro-crenulations parallel to F_2 axes (Plate 16A). Generally, due to the low angle between the S_1 and S_2 planes, the mineral lineation L_1 is difficult to distinguish from L_2.

L_2 can only be distinguished from L_1 where minor folds do occur. Over much of the ground layering and schistosity are virtually parallel. It is assumed that this schistosity is S_1, but in many cases it could be modified by superimposition of the second generation structure. In such cases, therefore, the lineations have been mapped as undifferentiated L.

The general trend of L_2 is parallel to the axis of F_2 folds. Most of the L_2 lineations plotted on stereographic projection (Fig. 10E) trend towards north-north-west or almost north. However, there is less variation in the amount of plunge of the mineral lineation, compared with the considerable variation of F_2 fold axes. Mullions, rodding and boudinage also occur occasionally (Plate 11A). Usually, boudinage structures are formed by leucocratic pods and lenses of granitic and quartz materials. Pinch and swell structures also develop along the axial surfaces of F_2 folds.

2.5.3. Macroscopic structures

One major synform has been recognised and part of a complementary major antiform (Map 2; Fig. 9). The hinge of the major synform particularly is made up of a number of subsidiary macroscopic antiforms and synforms so that the shape of the major structure is open and asymmetrical in style. Away from the hinge area, both limbs of the synform have steep dips. The hinges of macroscopic folds are thickened and the limbs are thinned. The axial planes of the folds dip towards the west at moderate to steep angles. The trace of the axial surface of the main synform, here called the Cooke Hill Syncline, is not readily traced in detail because of the subsidiary folds but the generalised trend indicates that it has been displaced at three places by late faulting of the third deformation (Map 2)*. A major fault, here called the Cooke Hill Fault (see Map 2), is associated with the eastern limb of the main antiform which is locally overturned, especially south of the Cooke Hill triangulation station. Further south most of the antiform structure has been cut off by the fault.

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* The two faults concerned strike north-north-west. There is some variation in the plunge of mesoscopic folds (Map 2) across the more southerly of these faults but the major structure has a northerly plunge throughout.

2.6. MECHANISM OF FOLDING

2.6.1. F_1 Folds

F_1 folds show the thickened hinges and thinned limbs which are characteristic features of similar folds. Such folds are commonly well developed in the heterogeneous rocks of the present area consisting of competent and incompetent layers marked by alternate bands rich in quartz-feldspar and biotite content resepectively (Plates 3, 4 & 5). The axial plane schistosity S_1 is more strongly developed in the hinge zone than in the limbs. This fact may be explained by the concept of "flattening" which plays a significant role in fabrics of deformed rocks. The term flattening was originally proposed by Sander (1930) and later used by Knopf and Ingerson (1938), Cloos (1947), de Sitter (1956, p. 214-6) and Ramsay (1962a, b). Ramsay (1967), in dealing with mechanism of flexural folding, has pointed out that the "most important process that leads to development of similar folds is that of flattening, in particular, that of differential flattening in 'a-c' section." In the present area, the amount of flattening has been measured for a few specimens and was found to range from 10-25% using the method (Coward, 1969) based on Ramsay's dip isogons. The F_1 folds therefore appear as a combination of flexural-slip and flattening.

2.6.2. F_2 folds

Like the F_1 folds, most of the F_2 folds are 'similar' in geometrical style, although 'concentric' types are not uncommon. Thus, the hinges of folds are generally thickened, but where interlayering of psammitic and pelitic materials is well developed the layers retain almost constant thickness throughout the folds. In migmatitic rocks the leucocratic granitic veins behave in a distinctly less ductile way than the psammitic and pelitic hosts and the fold styles are variable, but thickening in the hinges is not uncommon (Fig. 6 B). On the regional scale, the F_2 folds show weaker development of S_2 cleavage and schistosity indicating comparatively less strong deformation than the F_1. The style therefore suggests that flexural-slip was relatively more important and flattening relatively less important during the second deformation.

During F_2 , flattening resulted in considerable variation in similar folds even on hand specimen scale. The amount of flattening measured for a few mesoscopic folds ranges from 5-50%, which was mainly controlled by the relative ductility of the various components of the deformed rocks. This has been demonstrated by an F_2 fold consisting of multi-layered quartzo-feldspathic and pelitic rocks, illustrating a discontinuity in dip isogons (Fig. 8).

2.7. FAULTING

INTRODUCTION

The last phase of tectonism is an episode of brittle deformation. There are a number of major and minor faults mapped in the present area. Owing to complex histories, these have been classified into three groups. The first group of faults is related to the third phase of deformation. At the dying stage of second deformation, the falling temperature and pressure developed brittleness in the metasediments and resulted in a series of late faults. The other two groups of faults are complex in nature and their extension for large distances beyond the present area precluded full study of their history. These faults are related to parallel north-south scarp ridges on the eastern and western boundaries of the studied area. The western escarpment is produced by the Palmer Fault and the eastern escarpment controlled by the Milendella Fault.

Generally the faults have been recognized in the field by zones with cataclastic textures and breccias. Breccias contain rock fragments frequently cemented together by very fine grained rock flour, secondary carbonates and iron oxides. The strong shearing and crushing movement resulted in the formation of closely spaced uncemented tension fractures in fault zone rocks. It has been noticed that some of the pre-existing planar structures were not completely destroyed but their orientation has been disturbed. Commonly the fault zones are strongly weathered and some secondary mineralization of iron and copper of no economic importance is also seen. Retrograde metamorphism is seen as chloritization of biotite, sericitization and kaolinization of feldspar and less commonly alteration of diopside to amphibole. Rarely, kinking has developed in minor folds apparently due to fault movement.

2.7.1. Cooke Hill Fault with associated smaller faults

One major fault, here called the "Cooke Hill Fault", trends north-south and generally separates the Strangway Hill Formation from the Inman Hill Formation. In the north-east of the mapped area (Maps 1, 2 & 3) a branch of the fault diverges into the Strangway Hill Formation. The continuation of this Fault in the north has been mapped by Mills (1964), Rowley (1951) and Harms (1951). The Fault is characterized by a crush zone of a few tens of feet to a few hundred feet wide. In the crushed zone breccias are angular rock fragments, often cemented by finely ground rock flour or secondary calcites. The fine grained matrix shows bent twin lamellae in plagioclase and Boehm lamellae or very strong straining in inclusions of fractured quartz.

Unfortunately, due to poor outcrop and absence of marker beds, the displacement of the stratigraphic sequence could not be measured. Numerous water courses eroding deep down in the crushed zone indicate that the Fault dips towards the east with moderate to high angle. North of the present area where outcrops are well exposed, Mills (1964) has observed the displacement of metamorphic isograds on a fault which appears to be part of the same system. He suggests that the eastern side moved about 2½ miles northwards.

There are a few smaller late faults. Two important ones run roughly parallel to each other in north-west to south-east directions. These faults have narrow crushed zones with highly altered rocks. They displace the stratigraphic sequence and disturb the major structures. In the middle part of the area, one of these faults cuts obliquely across the nose of the major syncline. The western side shows an apparent offset of the axial trace of the major synform of about one mile to the north-west. The dip of the fault could not be determined due to poor exposures. Another fault in the north has been detected by a discontinuous narrow zone of mylonitized rocks with abrupt changes in dip and strike of the country rocks on either side of the fault. A good section of this fault is exposed in Cooke Creek where it is dipping almost vertically. These two faults therefore have apparent horizontal displacements consistent with their being conjugate to the Cooke Hill Fault. It seems unlikely however that the displacements are real. More probably the F_2 synforms originally developed in an en echelon pattern and the faults have later developed between the major elements of the pattern.

2.7.2. Palmer Fault

The western escarpment has a prominent physiographic feature parallel to the Palmer Fault zone. Originally the name Palmer Fault was proposed by White (1956, 1957) during the regional mapping of the Mannum Sheet. A small portion of the Fault is exposed in the studied area. The Fault can easily be detected on aerial photographs and is characterized in the field by a wide crushed zone. Brecciation, albitization and retrograde metamorphism of country rocks is a notable feature related to faulting. There are numerous undeformed acidic and basic dykes intruded in the fault zone which apparently post-date the faulting. The fault dips west at a high angle. There are no marker beds which allow the throw to be determined, but possibly the scarp is a Tertiary fault scarp rather than a fault line scarp.

Fenner (1930) has suggested that the Palmer Fault was initiated during the Tertiary movements as one of a series of meridional faults which gave rise to the "Mount Lofty Horst". However albitization and retrograde metamorphism along the Fault and the presence of the post-tectonic dykes suggest that the faulting was initiated in the Delamerian Orogeny although it may have been reactivated in the Tertiary period.

2.7.3. Milendella Fault

The abrupt rising up of the eastern escarpment from the Murray Plain was suspected to be due to a fault by White (1956) although he could not find any supporting evidence. Later Wills' (1964) detailed mapping north of the present area confirmed the presence of this Fault and suggested continuation of this Fault into the present area. The length of the eastern escarpment from north of the Truro Sheet through the Cambrai region to the present area is about 40 miles and the escarpment finally disappears south-east of Milendella in the Murray Plain. The Milendella Fault is coincident with this escarpment and was considered to be a high angle meridional Fault by Sprigg (1946). The Fault has been observed in the present area in only a few exposed sections, mainly south of Cooke Creek. The fault zone has been recognized by brecciated rocks composed of fragmented grains of quartz and plagioclase with bleached biotite. Unfortunately, most of the fault zone is highly weathered and a blanket of soil covered with thick bushy vegetation conceals most of the outcrops. The fault dips west at a high angle (70-80°). The dip of the Milendella Fault on the Truro Sheet (Coats & Thomson, 1959) is 60-80° west and 70-80° west in the Cambrai region (Mills, 1964). Mills has found evidence that faulting was initiated in (?) Palaeozoic time with displacement of many thousand feet. During the Tertiary period, the Fault was again active and has a net displacement of about 850 feet from early Tertiary to Miocene and about 200-300 feet from Miocene to Recent time (see Mills, 1964, Table 16).

2.8. STRUCTURAL ANALYSIS

The evidence from style and superposition presented in the previous section has been supplemented by statistical analysis of structural data as set out below. The collective diagrams showing the linear and planar structures do not show much complexity as a consequence of the polyphase deformation and structural displacement by late faulting. However, the area has been subdivided into 6 units (Fig. 11) to obtain a clearer picture of the degree of structural homogeneity and of the style of the major structures. There is considerable homogeneity of various structural elements when plotted separately in different sub-areas.

2.8.1. Schistosity S_1

Poles to schistosity for the whole area are shown in Figure 10. There are two principal maxima corresponding to the orientations of the limbs of the main synform and confirming a northerly plunge for the macroscopic structure. The great circle defined by these maxima, however, indicates a lower plunge for the S_1 structure than the general spread of the S_1 poles. Analysis of the various sub-areas shows that his is due to variation of plunge from sub-area to sub-area. The S_1 diagrams for the sub-areas 1-6 are plotted in Figure 11 and show that the sub-areas are considerably more homogeneous than the area as a whole. Figure 12 is a composite plot of the π S_1 girdles and β maxima from the sub-areas. It shows that, with the exception of sub-area 6, the π S_1 girdles vary in strike from 245° to 254° and that the plunge of the β axes which they define varies from less than 5° to 30°. The trend and plunge of girdle axes for sub-areas west of the Cooke Hill Fault vary systematically, being nearly horizontal in sub-areas 5 and 4 and steepening progressively through sub-areas 3, 2 and 1. The direction and amount of plunge corresponds reasonably well with the regional variation of small F_2 folds and L_2 lineations (Fig. 11). Sub-area 6, which lies to the east of the Fault, is exceptional in that the girdle axis has an almost vertical plunge and this is the only sub-area in which steep plunging fold axes and lineations were observed. The spread of the β axes in Figure 12 also corresponds closely with the average position of the axial plane of F_2 structures, and the intersections of the various sub-area S_1 girdles thus correspond to the concentration of the poles of F_2 axial planes. The statistical analysis therefore demonstrates convincingly that the F_2 folds are essentially cylindrical in individual sub-areas but show a systematic variation in plunge from sub-area to sub-area within a constant axial plane.

Sub-areas 3, 2 and 1 represent the west limb, the hinge zone and the east limb respectively of the main synform. This suggests that the major synform is not perfectly cylindrical with respect to the constant axial plane. It can be shown in this area that the F_2 axial plane is superposed slightly obliquely to F_1 and L_1 structures, which plunge gently south in the west limb and gently north in the east limb. Presumably this oblique superposition on earlier structures may be responsible for the plunge variation of F_2. Alternatively, the variation may be related to the presence of the Cooke Hill Fault on the east.

2.8.2. Layering S_0

The S_1 diagram is thus fully explained in terms of F_2 folding. Since S_0 and S_1 are usually sensibly parallel, S_0 was not usually measured separately so that Figure 10A can be regarded as a plot of S_1 and S_0. Figure 10B is a plot of S_0 measurements in places where S_0 could be seen to be folded around mesoscopic F_2 folds. The measurements therefore came mainly from sub-area 2 and can be compared with S_1 measurements from sub-area 2 in Figure 11. The plots therefore demonstrate that S_0 and S_1 are statistically parallel. This data thus confirms that the distribution of poles to S_0 and S_1 is due to the F_2 folding, which as Figure 10D and F show was essentially co-axial with F_1. The two maxima of poles to S_0 in Figure 10B define two average attitudes of layering in the two limbs. They show that both limbs are steep, with average dips of about 70° and suggest that the major structure is an upright, symmetrical fold with gentle north-north-west plunge. These conclusions are confirmed by the sub-area plots of S_1 for sub-areas 1, 2 and 3 (Fig. 11). However, in sub-area 4 the eastern limb of the synform has only a moderate average dip to the west while in sub-area 5 the western limb of the synform is overturned and has a steep westerly dip. Thus in sub-areas 4 and 5 the fold is clearly asymmetric with a westerly dipping axial plane.

2.8.3. Lineation L

As mentioned above it can be shown locally that the F_2 major structure has been superposed obliquely to L lineations. However, the measurements of L were not numerous and show considerable spread in orientation so that the statistical analysis does not allow any distinction in the orientation of L from F_2 in the area as a whole. The two maxima representing L and F_2 are virtually coincident and lie across the average position of the F_2 axial plane.

2.8.4. Fold axis (F_1)

The attitude of the F_1 folds on the collective diagram (Fig. 10D) shows parallelism with the L lineation (Fig. 10C). In the sub-areas, the trend of F_1 is variable, including a range of northerly and southerly plunges (Fig. 11, 1-6). Southerly plunges are especially common in the west limb of the Cooke Hill Synform in sub-area 3. Statistically, the first generation fold axes have the same orientation as second generation fold axes and linear structures again indicating general parallelism between the F_1 and F_2 deformations and a possible control of F_2 orientations by pre-existing F_1 fabrics.

2.8.5. Lineation (L_2)

Mineral lineation L_2 shows variable trend and plunge but has a well defined maximum indicating that the general orientation of L_2 is towards north-west (Fig. 10E). The general spread of L_2 lineations on the stereogram is similar to that F_2 axes (Fig. 10F) and β maxima all of which fall close to the average position of the F_2 axial plane (Fig. 10H).

2.8.6. F_2 axis

The collective diagram of mesoscopic F_2 fold axes for the whole area indicates a strong north-north-west plunging maximum which confirms the overall northerly plunge of the macroscopic Cooke

Hill Synform as indicated by the trend lines on Map 2. This indicates that the F_2 minor folds are congruent with the major structure in accordance with Pumpelly's rule (1894). In detail there is some plunge variation between the sub-areas (Fig. 11) but these are also consistent with the β axes derived from plots of poles to S_1 (Fig. 12). Sub-areas 3, 4 and 5 from the west limb of the Cooke Hill Synform exhibit nearly horizontal β axes and the F_2 mesoscopic folds show gentle plunges both to north-north-west and south-south-east. Sub-areas 1, and 2 from the east limb of the Cooke Hill Synform show definite north-north-west plunging β axes and the F_2 mesoscopic folds also show definite north-north-west plunging maxima.

Thus, in summary, the major fold plunges gently north. Mesoscopic data from the eastern limb confirms this plunge but mesoscopic data from the western limb indicates a nearly horizontal plunge. The consistency of this F_2 mesoscopic data with F_1 mesoscopic data has already been noted. The scatter of plunges within individual sub-areas is caused by a number of factors such as inhomogeneous plastic deformation in high grade metamorphism (cf. Ramsay, 1958) or superposition of F_2 folds on earlier F_1 structures (Offler, 1966; Hossack, 1968) or formation of disharmonic major F_2 structures resulting in fanning of S_2 planes (cf. Weiss & McIntyre, 1957; Hobbs, 1965). The first two factors are considered to be the major cause of the variation of F_2 fold axes.

2.8.7. Axial plane (S_2)

The collective diagram (Fig. 10G) shows a strong concentration of poles to axial planes of F_2 folds. As noted above, this concentration defines a virtually constant axial plane attitude in which all the linear structures of the area lie. This average axial plane (Fig. 10H) has a slightly more northerly trend than the north-north-west plunging F_2 fold axes but it nevertheless suggests that the main fold axial trace should trend west of north. The average trends sketched on map 2 however, trend about north to south. This suggests either, that the major fold does indeed have an axial plane slightly oblique to that of the mesoscopic fold maximum (but well within the range of mesoscopic orientations) or that the generalised fold trace in Map 2 is made up of smaller folds en echelon.

2.9. DISCUSSION OF STRUCTURE

The analysis shows that S_0 and S_1 are statistically parallel. If S_1 is an axial plane schistosity to F_1 folds as is indicated by the relation to mesoscopic folds, then the F_1 folds are isoclinal, and it follows that the present distribution of poles of S_0 and S_1 must be due to F_2 folding. As has been pointed out, the plunge variation of β axes occurs within a constant axial plane attitude of F_2. However, the relative paucity of F_1 minor folds and the lack of evidence for isoclinal recumbent folding in the Kanmantoo rocks generally (Offler and Fleming, 1968) suggests that S_1 may not be an axial plane schistosity to major F_1 isoclines but may have been initiated as a bedding plane schistosity perhaps developing from an earlier low grade bedding plane cleavage. If S_1 were partly a schistosity which was originally formed parallel to layering, the axial plane schistosity of F_1 folds would have to be regarded as a local modification of a pre-existing schistosity. Moreover, it would hardly be possible to distinguish between F_1 and F_2 folds as the cause of the distribution of S_0 and S_1 poles. The major Cooke Hill synform could then in fact be the composite result of F_1 and F_2 folding.

The distinction of major F_1 and F_2 folds therefore depends on the recognition of S_1 as an axial plane schistosity to F_1 major folds.

Mills (1964) in the Cambrai area, recognized two major folds with axial plane schistosities which he classed as S_1. His major syncline (here called the Evans Syncline) is a relatively open structure but his major anticline (here called the Somme Anticline) is relatively tight (Fig. 9). Consequently, the schistosity is virtually parallel to the layering on the overturned south-western limb of the SommeAnticline. The present area is apparently part of the overturned limb of that structure so that it is possible to regard S_1 as an axial plane schistosity on the macroscopic as well as the mesoscopic scale.

The tendency for F_1 structures, parallel to the hinge of the Somme anticline, to show southerly plunges in the west limb of the Cooke Hill Synform and northerly plunges in the east limb is to be expected from this superposition of macroscopic structures.

This discussion also indicates that there should be a major F_1 syncline complementary to the Somme Anticline in, or west of, the present area.

No such syncline has been recognized within the present area and if one occurred it should cause the Milendella marbles to reappear in its western limb. For this reason the possible stratigraphic correlation of the Milendella marbles with the calc-schists and the calc-gneisses in the west of the present area has been considered. The correlation has been rejected however, since although each rock association is fairly constant in character along strike, they are completely different across the suggested synform (see p. 24). Moreover, there is no evidence from the mapping of trend lines in the intervening quartz-feldspathic sediments that such a correlation could be achieved.

West of the Palmer Fault (see Map 3) a major synformal structure (here called the Burns Synform) occurs containing gneisses and psammites which have been mapped as Brukunga Formation on the Adelaide 4 mile Sheet. Insufficient work has been done on the relations of F_1 and F_2 in this region although it was regarded as an F_2 structure by Offler and Fleming (1968). It seems possible from the map relations that the Burns and Evans Synforms are of similar character (i.e. of the same fold generation), in which case there would be conflict between the interpretations of Mills (op. cit.) and Offler and Fleming (op. cit.).

This problem will be resolved only by further detailed observations and mapping outside the present area and is beyond the scope of the structural work in this thesis.

One other general point however deserves notice. The major F_2 folds (Fig. 9) form an en echelon arrangement adjacent to the Cooke Hill Fault and the fold trend roughly bisects the angle between the two main fault trends. It therefore seems possible that the

Cooke Hill Fault was developed in a later stage of the continuous deformation which first produced the F_2 folds. It is tempting to suggest on this basis that the Evans Synform is also essentially F_2. However, Mills finds that the axial plane schistosity of the Evans Syncline is the earliest tectonic structure recognized and therefore regards it as S_1 and the fold as F_1.

The structural analysis of the present area was undertaken primarily to provide a setting both in space and time for the study of migmatization and granite emplacement in the area. The strong linear fabric in the Rathjen Gneiss and similar rocks in the present area is not easily understood in terms of what is known of the F_1 deformation in the metasediments but it demonstrates that the Rathjen Gneiss is the earliest of the granitic rocks and was present throughout the deformation history. Migmatization apparently commenced during F_1 and leucocratic veins of granitic and pegmatitic composition are developed parallel to S_0 and to the axial planes of F_1 folds. These veins are folded by F_2 folds (Plates 8A, 10B, 11A) but migmatization also continued through the F_2 folding since leucocratic veins commonly occur along the axial surfaces (plate 8A) and in the hinge zones of F_2 folds.

During the later stages of the F_2 deformation, many small bodies of the Cooke Hill tonalite-granodiorite and other minor acidic intrusives were emplaced in the metasediments. The intrusives are mildly deformed with weak preferred orientation of platy minerals and common undulose extinction in quartz grains. The intrusions tend to be concentrated in the hinge zones of the major F_2 folds and elongate parallel to their axial traces (Map 1). They have been found to cut across F_2 structures. The field relations and petrological features of the intrusive rocks are described in more detail in Chapter 7.

These intrusions apparently occurred when the rocks were becoming less ductile presumably due to lower temperature and pressure, prior to the D_3 faulting. There is another event of acidic intrusion characterized by undeformed granitic rocks probably emplaced during and soon after the D_3 brittle deformation. The most important point to emphasise, however, is that the Cooke Hill intrusives can be regarded as a subsequent event to the main migmatization of the area.

2.10. STRATIGRAPHY

2.10.1. Introduction

Despite the absence of persistent marker beds and lack of sedimentary facings in high grade rocks, the general stratigraphic succession has been established by the structural studies. The metasedimentary rocks of the present area are similar to those in adjacent areas in the north, although the present sequence is inverted in the downward-facing Cooke Hill synform and antiform. The established succession can be seen in the geological sections drawn across the mapped area (Fig. 13) and is presented in Table 1. The stratigraphy of the present area can be divided into two major units on the basis of broad lithological associations. These units correspond to the Strangway Hill and Inman Hill Formations of Thomson (1969a) and are separated in the present area by the Cooke Hill Fault, except on the north-eastern part of the area.

The group of rocks west of the Cooke Hill Fault has been shown as Inman Hill Formation on the Adelaide 4 mile Sheet (Thomson, 1969a) and it appears that this group can be traced continuously to the type area on the Barker Sheet. It has been demonstrated by Mills (1964) that this group lies in the overturned limb of the Somme Anticline so that the rocks of the Somme Anticline are stratigraphically older and have been correlated with the Strangway Hill Formation. It is generally considered that the base of the Kanmantoo Group metasediments includes numerous lens shaped bodies of limestone (Sprigg & Campana, 1953; Horwitz et al., 1959). The correlation with the Strangway Hill Formation on lithological grounds is therefore reasonable although it must be noted that the nearest undoubted Strangway Hill Formation on the Barker Sheet is some 60 miles distant.

According to Geological Survey mapping recorded on the Adelaide Sheet (Thomson, 1969a) the group of rocks east of the Cooke Hill Fault can be mapped continuously into the Strangway Hill Formation of the Somme Anticline (Fig. 9). However, the writer's mapping shows that the Cooke Hill Fault continues northwards thus isolating some calcareous rocks east of the Evans syncline. Nevertheless, the writer accepts the correlation of these rocks with the similar calcareous rocks west of the syncline in the Somme Anticline as postulated by Mills (op. cit.). It therefore seems reasonable to apply the term Strangway Hill Formation to these rocks since they contain marbles and calc-silicate rocks.

The metasedimentary rocks exposed on the east of the Cooke Hill Fault contain abundant marble and calc-silicate beds intercalated with quartz-feldspar schists and gneisses (Map 1; Chapter 3, Sections 1 & 2). The Milendella marbles and interbedded schists and gneisses are consistently dipping towards the west, and in the geological sections A-B, C-D, E-F and G-H (Fig. 13) the marbles do not re-appear following the limbs of the major F_2 structure. The continuity of these rocks with those mapped by Mills east of the Evans Syncline suggests that the rocks are in normal stratigraphic succession and young to the west. Three lithological units have been distinguished on map 1 and have therefore been numbered 1, 2 and 3 (Table 1) from east to west.

On the west of the Cooke Hill Fault, the major synform and antiform are essentially composed of the Inman Hill Formation and are downward-facing structures. Therefore, the younging in the west limb of the main synform is towards the west and the core of the synform is occupied by quartzo-feldspathic schists and gneisses. The schists and gneisses on the west of the Cooke Hill Fault do not show much difference in modal composition from those on the east but the composition of biotite does appear to be different. A characteristic variation in colour of biotites in thin section gives some evidence that the schists and gneisses of the eastern and the western sides of the Cooke Hill Fault have a different source material.

TABLE 1

STRATIGRAPHIC SUCCESSION

YOUNGEST

INMAN HILL FORMATION	7.	Quartzo-feldspathic schists and gneisses
		Palmer Fault
	6.	Diopside gneisses, actinolite-tremolite schists and quartzo-feldspathic gneisses
	5.	Quartzo-feldspathic gneisses, migmatites, minor semi-pelitic/pelitic gneisses and aluminous pelitic schists
	4.	Quartzo-feldspathic gneisses, mica schists, granitic gneisses, minor aluminous pelitic schists
		Cooke Hill Fault
STRANGWAY HILL FORMATION	3.	Quartzo-feldspathic gneisses, semi-pelitic/pelitic gneisses, minor aluminous pelitic schists and calc-silicate rocks
	2.	Impure marbles and associated calc-silicate rocks and interbedded quartzo-feldspathic gneisses
	1.	Quartzo-feldspathic gneisses, semi-pelitic/pelitic gneisses and calc-silicate rocks
		Milendella Fault

OLDEST

In the western part of the area, two lenticular bodies of calc-gneisses (diopside gneisses) enveloped by calc-schists (actinolite-tremolite schists) are exposed as a part of the Inman Hill Formation. Similar types of rocks have been mapped on the north-west of the present area by the Geological Survey of South Australia (Map 3). These rocks are well exposed in a major syncline, about 3 miles north of Burns. The core of the syncline is occupied by the Brukunga Formation which overlies the Inman Hill Formation (Thomson, 1969a). A cross-section across the syncline shows that calc-schists and the calc-gneisses are in the top of the Inman Hill Formation succession although as noted on p. 24 the significance of fold hinges recognized in these rocks is not clear.

The stratigraphic succession of the Inman Hill Formation of the area studied is presented in Table 1, in the sequence from unit 4 to unit 7. This succession is based on the younging of beds in the major F_2 structures. The lowest unit of the Inman Hill Formation is quartz-feldspathic schists, mica schists, granitic gneisses and minor aluminous pelitic schists (Table 1, unit 4). This is overlain by the quartzo-feldspathic gneisses and intercalated semi-pelitic and pelitic gneisses in quartzo-feldspathic gneisses and veined gneisses (migmatites) with minor aluminous pelitic schists (unit 5). The diopside gneisses and actinolite-tremolite schists are included in the upper part of the Inman Hill Formation (unit 6). The quartzo-feldspathic schists and gneisses (unit 7) to the west of the Palmer Fault occupy only a small portion of the area studied. On Map 3 which is taken from the Adelaide 4 mile Sheet (Thomson, 1969a), these rocks are shown as the Brukunga Formation. However, the rocks seen immediately west of the Palmer Fault seem to be identical with those east of the Fault and were initially mapped as the Inman Hill Formation (Maps 1 & 2). None of the diagnostic marker beds - Pebble beds or Pyritic beds, as shown by Thomson (1969b, Fig. 43) are found in the area mapped. Moreover, White and Thatcher (1957) do not show the Brukunga Formation in this area. For this reason, the quartzo-feldspathic rocks west of the Palmer Fault have been shown as the highest unit of the Inman Hill Formation. It is possible they may belong to the Brukunga Formation, but whether they do or not, does not affect any arguments developed in this thesis.

The stratigraphic thickness of the Strangway Hill Formation and the Inman Hill Formation was measured in the cross-sections A-B and G-H (Map 1). The maximum thickness of the Strangway Hill Formation (Table 1, units 1-3) is estimated to be about 4,000 feet in the G-H section, while the Inman Hill Formation (Table 1, units 4-6) is about 10,000 feet in the A-B section. These thicknesses are probably overestimated due to polyphase deformation. The tectonic movements have caused the repetition of rock sequences during F_1 and F_2 folding. Generally, the beds are thinned in the limbs and thickened in the hinges of mesoscopic and macroscopic folds, indicating a considerable variation in the thickness of stratigraphic units. Therefore, the above thicknesses measured in the geological sections must be used with caution although no large scale repetitions have been detected.

CHAPTER 3

PETROLOGY OF THE COUNTRY ROCKS

I - MILENDELLA MARBLES AND CALC-SILICATES

3.1. INTRODUCTION

The Strangway Hill Formation occurs along the eastern edge of the area. It is separated from the Inman Hill Formation to the west by the Cooke Hill Fault. In this area the most conspicuous rocks are the marbles and associated calc-silicates which outcrop along the eastern escarpment. The marbles are lenticular beds of somewhat irregular shaped locally thickening in the middle part and tapering towards the end. They are interbedded with quartzo-feldspathic schists and gneisses. Generally the marbles grade laterally into less calcareous rocks and then in turn to quartzo-feldspathic schists and gneisses. The contact of any individual marble bed with country rock is relatively sharp; the gradational zone is composed of calc-silicates. Frequently less calcareous rocks and quartzo-feldspathic bands are intercalated as small lenses in the marbles. A typical outcrop of marble shows flat weathered surfaces frequently, partially or totally covered by soil and bushy vegetation. Sometimes surface weathering produces subrounded blocks of marble with a general tor-like appearance (Plate 34A). The compositional layering in the marbles and calc-silicates is marked by concentrations of silicate minerals, and laminations about a centimetre in thickness are continuous for as much as 10 metres in length. The concentration of silicate minerals in layers is interpreted as reflecting thin marly bands in the original limestones. These bands are parallel to the marble beds and intercalated quartz feldspar metasediments, and consequently the macro-layering is taken to represent the original bedding (see Map 1).

There is practically no pure marble, there being at least 10% silicate minerals in all samples. The rocks are described in two categories: the impure marbles and the calc-silicates. The distinction is based on field observations. Those beds which are mostly calcite with some minor patches rich in silicates were mapped as marbles and are described in this section and those beds which are predominantly calc-silicates are described in Section 3.3.

3.2. MARBLES

The marbles contain 10-30% silicate minerals and the colour of the marble ranges from light to dark grey depending upon the amount of silicate minerals present in it. In general, it has a coarse grained texture. No prominent mineral lineation is visible except for flattened calcite grains and rare mica flakes which define the weak schistosity plane in the crystalline limestone. Foliation develops due to enrichment of silicate minerals in alternate dark bands. Amphiboles and clinopyroxene show up as green equidimensional grains in the marble. Brown coloured sphene and black iron ores are also visible by naked eye. Occasionally, flakes of biotite (phlogopite) are seen in hand specimen. The common minerals in the marbles are calcite, diopside, scapolite, quartz, plagioclase, potash feldspar, biotite, hornblende, and phlogopite. One assemblage contains spinel, clinohumite and forsterite. The accessory minerals are sphene, tourmaline, muscovite, rutile, actinolite and calcite. The secondary paragenesis in the marble probably results from retrograde metamorphism of high grade mineral assemblages at the end of a tectonic episode, due to decrease in temperature and increase in P_{H_2O}, or both changes operating together. Besides the mineralogical alteration some textural changes have been observed in marbles: coarse grained granular aggregates of calcite have been altered to fine grained polygonal calcite producing mortar texture. This is believed to have occurred when the original coarse grained rock recrystallized by an annealing process during retrograde metamorphism.

Garnet is found in those marbles and calc-silicate rocks which are in direct contact with, or near granitic bodies. Commonly garnet appears as reaction rims around cloudy scapolite and pyroxene. This observation leads to the conclusion that the formation of garnet in the calcareous rocks here is due to metasomatism by the residual fluids from the granitic rocks. Thus rocks containing garnet are regarded as true skarn rocks and have been described in section 3.4.

The following mineral assemblages have been observed in the marbles:

(1) Quartz-calcite-diopside-scapolite-plagioclase-potash feldspar
(2) Quartz-calcite-diopside-scapolite-amphibole-potash feldspar
(3) Quartz-calcite-diopside-scapolite
(4) Quartz-calcite-diopside-scapolite-potash feldspar-biotite-phlogopite (chlorite)
(5) Quartz-calcite-diopside-scapolite-plagioclase-potash feldspar-epidote
(6) Calcite-forsterite-clinohumite-spinel.

All contain mica, sphene, tourmaline, rutile, pyrite and iron ores as accessory minerals.

The mineral assemblages recorded in the marbles show that an appreciable amount of argillaceous material contaminated the original carbonate rocks. The presence of quartz in all assemblages, except in silica-deficient marble (assemblage 6), and excess of potash feldspar and plagioclase makes it difficult to represent all the minerals present on ACKF diagrams. Mills (1964) attempted to present most of the assemblages of high grade marbles of the Cambrai area on an ACKF tetrahedron, and the present author finds his representation quite adequate for most of the Milendella Marbles (see Mills, 1964, Fig. 37). The silica-deficient marble has also been represented on a tetrahedral diagram modified from that used by Burnham (1959) for contact metamorphic marbles at Crestmore to illustrate the calcite-forsterite-clinohumite-spinel assemblage of the present area (Fig. 17). Other assemblages seen by Burnham do not occur here where the amount of original siliceous and argillaceous impurity was probably much greater.

3.2.1. Petrography

Generally, the marbles have a granoblastic texture with some flattened calcite grains making up 70% of more of the total. Most of the remaining 30% is an association of the common minerals, viz. diopside, scapolite, quartz and feldspar. The amount of mica minerals is variable but quite small. In general the scarcity of mica in the hig grade marbles of the Milendella area indicates that muscovite and phlogopite found at lower grades, became unstable in the presence of quartz and calcite and reacted to form potash feldspar, scapolite and diopside. This has been demonstrated in the north of the present area by Mills (1964) where mica is plentiful in the low grade rocks but starts diminishing with the first appearance of scapolite and diopside in marbles when the sillimanite-muscovite zone is reached. The marbles from the Mills' area extend into the present area (Map 3).

The petrographic description of the major component minerals is described below:

Calcite ranges in shape from large granoblastic grains to a mosaic of fine grains with polygonal shape. Twins, well developed in the crystalloblastic mineral, are commonly highly deformed and show kink bands (Plate 34B). The development of a mosaic of polygonal shaped grains is attributed to annealing recrystallization (Plate 35A). Small inclusions of quartz, feldspar and scapolite are often seen in the crystalloblastic calcite.

Quartz is present in most of the mineral assemblages. It appears as either tiny blebs or as subrounded grains interspersed in the calcite matrix. The grain size ranges from .01-.5 mm and averages about .2 mm. The amount of quartz is fairly variable, averaging approximately 5%.

Potash feldspar occurs invariably as zenomorphic grains filling interstices between the calcite grains. Twinning is uncommon but in a few grains cross-hatch twinning is seen. The feldspar is recognized in thin section by its low relief and is confirmed by the yellow stain when tested with cobaltinitrite. The amount of potash feldspar is variable; about 5% is average.

Plagioclase is found in many shapes and sizes, sometimes as large crystalloblastic grains, and also as inclusions in other minerals. The grain size ranges from .01-1.5 mm., the larger crystals showing irregular outlines. Albite twinning is commonly seen, either well developed or in patches, but untwinned feldspars are not uncommon. Deformation features such as bent twinning and brittle fracturing displacing the twin lamellae are frequently seen. Secondary calcite commonly occurs as fracture filling. Alteration of plagioclase to sericite and clouding with a fine dust is usual. The plagioclase ranges in composition from oligoclase to labradorite, the maximum anorthite content being An_{58}. Plagioclase is always less in amount than potash feldspar.

Clinopyroxene forms large xenoblasts, irregularly intergrown with scapolite. The grain size of clinopyroxene ranges from less than 1 mm to 5 mm. Large crystals up to 3-4 mm are common. In thin section, the mineral is colourless to green with faint pleochroism. White (1959) measured the optical properties of a clinopyroxene from the Mi endella Marbles as $\beta = 1.702$, $2V_Z = 60^{\circ} \pm 1$, indicating a composition of about 60% diopside and 40% hedenbergite. Lamellar twinning and kink bands are commonly found in the diopside (Plate 35B); some twins show deformation features similar to the experimentally deformed twin lamellae of diopside described by Raleigh and Talbot (1967). Many diopside grains show fractures and regular partings. These fractures are always filled with secondary calcite indicating the post-tectonic tension fracturing of brittle minerals. The average amount of diopside is about 15%.

Scapolite occurs as large ragged xenoblasts or as aggregates of equant crystals in the rock matrix. The growth of scapolite is contemporaneous with diopside as the two minerals are often intimately associated with each other. Very often the rapidly growing scapolite engulfs diopside or vice versa. Tension fractures, as parallel partings which are sometimes filled with secondary calcite, are frequently found in the scapolite. White (1959) showed the composition of scapolite in the Milendella Marbles to be about 60-65% meionite. Small inclusions of sphene diopside and feldspar are common. Quantitatively, it is the next abundant mineral after diopside.

Clinohumite (X = pale yellow, Y = Z = golden yellow). This mineral has been recorded by the author for the first time in the Milendella-Cambrai Marbles. It was detected in the thin section and confirmed by its X-ray powder pattern. One specimen (A285/635) collected from a marble layer contains clinohumite together with calcite forsterite and spinel (assemblage 6). In thin section, it appears as skeletal shaped grains intimately associated with forsterite (Plate 36A). Textural evidence indicates that it has crystallized later than forsterite.

Forsterite occurs as irregular aggregates, ranging in grain size from .1-.5mm in diameter. No cleavage is visible but numerous irregular fractures are common. It is intimately associated with clinohumite along curved boundaries. It commonly alters to antigorite along the fracture planes.

Antigorite is found as a secondary mineral replacing forsterite or clinohumite along the fracture planes. Sometimes, the small granules of both minerals are completely replaced by antigorite. This mineral has been identified by its X-ray powder pattern.

Spinel is green in colour with well developed subhedral crystals or small equant grains (Plate 36B). The grain size is variable and ranges from less than .1 mm to 1 mm. Euhedral crystals are common (average .5 mm) with typical octahedral cleavage. There is no doubt that it is a primary mineral crystallized in equilibrium with clinohumite and forsterite.

Amphibole is present in small amount. Small crystals of hornblende with X = yellowish green, Y = dark green, Z = dark bluish green are found in some marbles. In thin section, it appears as a primary mineral without any indication of being a secondary product of pyroxene.

Actinolite, X = pale green, Y = dark green, Z = dark bluish green, also occurs. Its common association with diopside as poorly crystallized irregular shaped aggregates suggests that it is a retrograde product of secondary origin.

Epidote is found in various shapes and sizes as disseminated masses or skeletal grains. Textural evidence indicates that its occurrence as a primary mineral is doubtful. Mills (1964) has also described epidote in high grade marbles and pointed out its absence in low grade marbles of the Cambrai area which supports the contention that it is a secondary mineral. The most common variety occurs as X = pale yellow, Y = greenish yellow, Z = yellowish green. Besides this, clinozoisite (colourless) and piemontite (pink to amethyst with moderate pleochroism) have also been found in some marbles.

Mica is frequently seen as small flakes, rarely exceeding more than 2% of the total rock. Biotite, phlogopite and muscovite occur as primary minerals either interspersed in calcite-diopside-scapolite matrix or as small inclusions in xenoblastic grains of feldspar and scapolite. Occasoonally, crenulated flakes of biotite are also seen in some thin sections. Chlorite is found as a secondary mineral replacing biotite. Minor accessories: sphene is the most abundant mineral amongst the accessories. It appears as lozenge-shaped crystals or anhedral granules embedded in the calcite matrix. Very frequently, it is found as an inclusion in granoblastic grains of scapolite, diopside and calcite. Subrounded grains of apatite, reddish brown acicular crystals of rutile and disseminated granular masses of pyrite and iron ores are other common accessories. Less commonly, tourmaline as a colourless to pale yellow mineral is seen in some marbles.

3.2.2. Metamorphic reactions for paragenesis of primary minerals

The most common mineral assemblages recorded in the marbles are calcite, diopside, scapolite, potash feldspar, plagioclase and quartz. This clearly shows that the original limestones contained enough argillaceous material to give rise to the sillicate minerals. Calcite is the only carbonate mineral found in the crystalline limestones, with the exception of one sample (A285/635) which comes from a magnesian-carbonate rock near the marginal part of a calcic marble bed contacting with quartzo-feldspathic rock. This may have been a dolomitic rich band which gave rise to forsterite and clinohumite by contact metamorphism caused by pegmatitic bodies. The actual specimen collected from the marble bed does not show any direct contact with any igneous body, but such pegmatites are common in adjacent rocks.

The diopside in the high grade marbles of the present area is considered to be formed by the destruction of biotite by the following reaction.

$$\text{biotite} + 3\ \text{calcite} + 6\ \text{quartz} \rightleftharpoons 3\ \text{diopside} + \text{potash feldspar} + 3CO_2 + H_2O \quad \ldots\ldots\ldots\ldots\ldots\ldots\ldots (1)$$

The presence of K-feldspar and the absence of dolomite in the marbles of the Cooke Hill-Cambrai area shows that the following reactions for genesis of diopside are unlikely:

$$5\ \text{dolomite} + 8\ \text{quartz} + H_2O \rightleftharpoons \text{tremolite} + 3\ \text{calcite} + 7CO_2 \ldots\ldots\ldots (2)$$

$$\text{tremolite} + 3\ \text{calcite} + 2\ \text{quartz} \rightleftharpoons 5\ \text{diopside} + 3CO_2 + H_2O \ldots\ldots\ldots (3)$$

The absence of primary tremolite in low grade marbles of the Cambrai area rules out the possibility that the above reactions (2) and (3) have contributed to the formation of diopside in the high grade marbles during progressive regional metamorphism. On the other hand, it has been noticed that biotite is abundant in low grade marbles but starts diminishing in favour of clinopyroxene with increasing grade of metamorphism. On the basis of these observations, Mills (1964) advanced the idea that destruction of biotite is mainly responsible for the genesis of clinopyroxene in the marbles of the Cambrai area. He devised the following reaction by which the original biotite decomposes to form a more magnesium rich biotite in equilibrium with diopside and potash feldspar.

$$\text{biotite I} + \text{calcite} + \text{quartz} \rightleftharpoons \text{biotite II} + \text{diopside} + \text{potash feldspar} + CO_2 + H_2O \quad \ldots\ldots\ldots\ldots\ldots (4)$$

He demonstrates that the above reaction starts from the sillimanite-muscovite zone in which some diopside was formed in coexistence with magnesium rich biotite. With further increase in metamorphic grade, the biotite disappeared gradually, until it was missing in the highest grade. This mechanism for the formation of diopside from biotite is similar to that proposed by Ramberg (1952) for the destruction of phlogopite to give rise to diopside in high grade marbles of amphibolite facies. He gave the following reaction:

$$\text{phlogopite} + 3\ \text{calcite} + 6\ \text{quartz} \rightleftharpoons \text{potash feldspar} + 3\ \text{diopside} + 3CO_2 + H_2O \quad \ldots\ldots\ldots\ldots\ldots (5)$$

Mills' reaction (4) produces diopside at the cost of biotite rather than phlogopite (biotite being the much more abundant mineral in the low to middle grade rocks of the Cambrai area). But a progressive enrichment of magnesium content in the biotites, with increasing grade of metamorphism, resulted in the formation of a biotite which was close to phlogopitic composition (see Mills, 1964, R.I. of biotites in Appendix III; Fig. 39). After attaining such a composition, a phlogopitic biotite will become unstable and will form diopside according to reaction (5).

In the present area, biotite is uncommon, a fact which appears to support Mills' hypothesis for the genesis of diopside. The potash feldspar which is a product of the biotite in equilibrium with diopside is found in most of the marbles. A characteristic feature of the potash feldspar is the lack of twinning, a feature which is also observed in the marbles of the Cambrai area.

Scapolite: the formation of scapolites may have resulted from destruction of muscovite in the high grade marbles. The presence of muscovite in low grade rocks of the Cambrai marbles and its rarity in the high grade marbles indicates that it has contributed to the genesis of scapolite. Mills suggested the following reaction:

$$\text{muscovite} + 2\ \text{calcite} + 4\ \text{quartz} \rightleftharpoons \text{orthoclase} + \text{anorthite} + 2CO_2 \quad \ldots.(6)$$

The anorthite released from the above reaction will be used up to form scapolite in the presence of Cl and CO_2 in the fluid phase. Some anorthite may go to the coexisting plagioclase increasing its calcium content.

The stability of scapolite in equilibrium with potash feldspar and the absence of primary muscovite leaves no doubt that Mills' reaction (6) is responsible for its genesis in the Milendella Marbles.

Clinohumite appears to be formed by metasomatism in silica-deficient dolomitic marble by pegmatitic intrusives. Burnham (1959) has described the assemblage calcite-clinohumite-forsterite-spinel formed by metasomatic effect of quartz-monzonite intrusion into dolomitic marble. The aqueous solution released from the intrusive added silica and alumina to the marbles and also increased P_{H_2O} which resulted in dilution of P_{CO_2} to a few bars. Under these conditions, clinohumite was metasomatically formed by brucite reacting with silica derived from an aqueous fluid.

$$9\ \text{brucite} + 4\ \text{silica} \quad 1\ \text{clinohumite} + 8H_2O \quad \ldots\ldots\ldots\ldots\ldots\ldots\ldots\ldots(7)$$

Texturally, there is no evidence that brucite was stable in the marble of the present area. Therefore, the clinohumite is considered to be formed by the following reaction:

$$\underset{\text{dolomite}}{7CaMg(CO_3)_2} + \underset{\text{silica}}{3SiO_2} + H_2O \rightleftharpoons \underset{\text{clinohumite}}{Mg(OH)_2\ 3Mg_2(SiO_4)} + \underset{\text{calcite}}{7CaCO_3} + 7CO_2 \ldots.(8)$$

It is proposed that the silica was metasomatically introduced in the system through the aqueous fluid released from the nearby pegmatite bodies, which also maintained a low level of P_{CO_2} due to abundance of the hydrous phase.

Forsterite can be formed by the following reaction as suggested by Turner (1968) based on an equilibrium curve for decarbonation reaction.

diopside + 3 dolomite $\rightleftharpoons$ 4 calcite + 2 forsterite + $2CO_2$ ....(9)

The forsterite bearing mineral assemblage does not contain any diopside. Therefore, the above reaction, which usually takes place in high grade metamorphic conditions (cf. Metz, 1967), may have operated in the present area due to contact metamorphism of igneous bodies. Burnham (1959) gave the following reaction for genesis of forsterite in the Crestmore marble.

2 clinohumite + 1 silica $\rightleftharpoons$ 9 forsterite + $2H_2O$ ....(10)

Textural relationships of forsterite and clinohumite in the marble of the present area do not support the above reaction. The petrographic studies indicate that the forsterite is earlier than clinohumite as the forsterite is extensively replaced by clinohumite. Therefore, the forsterite seems to be formed by reaction (9).

Spinel has been formed by the following reaction when the dolomite is metasomatically destroyed in the presence of a hydrous phase.

$$\underset{\text{dolomite}}{CaMg(CO_3)_2} + Al_2O_3 \rightleftharpoons \underset{\text{spinel}}{MgAl_2O_4} + \underset{\text{calcite}}{CaCO_3} + CO_2 \text{ ....(11)}$$

Al_2O_3 was added to the system by aqueous solution discharged from the pegmatites.

3.3. CALC-SILICATE ROCKS

INTRODUCTION

The calc-silicate rocks are distinct lithological units of the Strangway Hill Formation and serve as good marker beds on a local scale (Plate 37A). They occur as lenses or thin bands in the marbles and quartzo-feldspathic rocks, ranging in thickness from about a centimetre to as much as 10 metres. These rocks are extremely variable in texture and in mineralogical composition; consequently, alternate light and dark bands appear, due to variation in the amount of silicate minerals. The regional schistosity of the quartzo-feldspathic schists and gneisses is parallel to compositional bands of the calc-silicate rocks.

The mineral assemblages of the calc-silicate rocks show a great degree of variation from one bed to another. Such features have also been recorded by White (1956, 1959), Mills (1964) and Offler (1966) for the calc-silicate rocks of the Palmer, Cambrai and Pewsey Vale areas respectively. The most common primary mineral assemblages observed in the present area are listed below.

(1) Clinopyroxene-amphibole-scapolite ± calcite
(2) Clinopyroxene-amphibole ± scapolite
(3) Clinopyroxene-scapolite ± amphibole
(4) Amphibole-scapolite ± clinopyroxene

In all these mineral assemblages, one or more of quartz, plagioclase, potash feldspar and biotite may be present as an important constituent. Sphene, apatite, zircon and iron ores are the common accessories.

3.3.1. Petrography

The calc-silicate rocks are fine to medium in grain size with an average of .1-.5 mm. The general variation in texture is seen on various scales from mesoscopic scale to thin section scale, consisting of alternate fine to medium grained layers. These layers also show distinct variation in mineralogical composition, i.e. one is rich in diopside, another is rich in amphibole etc. This preferential crystallization of clinopyroxene and amphibole in different layers may be related to the initial CO_2/H_2O ratio of the original primary calcareous layers during metamorphism (cf. Hietanen, 1963, 1971). Hietanen considers that diopside crystallizes at high P_{CO_2} and actinolite at high P_{H_2O}. A brief account of individual minerals of the calc-silicate rocks is described below.

Clinopyroxene is usually found as large ragged porphyroblasts up to 2-5 mm in diameter but some may exceed 10 mm in size. The colour is light green non-pleochroic with patches of small inclusions of strongly pleochroic, blue-green amphiboles (Plate 37B). The optical properties are indicative of the diopside-hedenbergite series. The deformational twinning, as closely spaced fine lamellae, is frequently seen in clinopyroxene (e.g. Raleigh, 1965; Carter & Raleigh, 1969). Some grains are strongly deformed and form kink structures. The clinopyroxene alters along the cleavage planes and around the edges to amphiboles.

Amphiboles vary considerably in their shape and size, occurring either as strongly oriented prismatic crystals or as ragged masses (Plate 38A). Generally, there are two types of amphibole; one type is hornblende with X = olive green to pale green, Y = green to dark green, Z = dark bluish green; the other is actinolite, X = colourless to pale green, Y = green, Z = dark green. Hornblende is much more abundant, and is a common constituent of the calc-silicate rocks. As in the marbles, actinolite is found as an alteration product of the clinopyroxene with which it is usually associated. The amphibole ranges in size from .1 mm. to elongated plates up to 15 mm. in length. The common average size is 3-5 mm.

Scapolite is the most abundant mineral in the felsic bands or layers. It also fills the interstices between interlocking aggregates of amphibole and clinopyroxene. Some of the well crystallized scapolites appear as large granoblasts with numerous roughly parallel fracture planes (Plate 38B). These fracture planes are filled with secondary calcite and formed during the last phase of brittle deformation. The small inclusions of sphene and feldspar are commonly seen. It is the next most abundant mineral after clinopyroxene and amphibole.

Calcite is extremely variable in its occurrence. It may be an important constituent in some calc-silicate rocks or may be completely absent. Whenever it is present, it shows well spaced fine polysynthetic twinning (Plate 39A). Most of the calcite occurring in the presence of abundant quartz in the original sediment has disappeared under high grade metamorphic conditions to form calc-silicate minerals. Parras (1958) has described a similar type of high grade calc-silicate rock lacking in calcite.

Quartz and feldspar form the smaller proportion of the rocks. In some felsic bands they may be as abundant as scapolite. Generally, twinning is poorly developed in plagioclase, but where it occurs it has simple albite or a combination of both Carlsbad and albite law twins. The composition of plagioclase ranges from oligoclase to labradorite.

K-feldspar also shows a lack of twinning in general. Biotite is less common, and is present in a small amount in some calc-silicate rocks. Epidote is a secondary mineral and forms as an alteration product of scapolite and plagioclase, appearing as skeletal shaped aggregates most commonly along the cleavage planes or near the margin of scapolite. In some rocks epidote has such abundant and well crystallized idioblasts, that its secondary origin is doubtful (Plate 39B). Generally, epidote is a moderately pleochroic mineral, X = colourless, Y = Z = pale yellow, and has strong birefringence. Sphene is found as euhedral lozenge shaped crystals to subrounded granular aggregates. It is much more abundant than other accessory minerals, occasionally becoming an important constituent of the rock. Granular masses of iron ores, apatite and subrounded zircon may be other common accessories in the calc-silicate rocks.

3.3.2. Conclusions

The close association of the calc-silicate rocks with marbles and features such as less calcareous beds grading into more calcareous marble beds indicates that these rocks have been formed by metamorphism of calcareous shale. The presence of regular banding parallel to marble beds shows the sedimentary character. Those calc-silicate rocks which are intercalated with the quartzo-feldspathic schists and gneisses have similar mineral assemblages, and, therefore, are considered to be derived by a similar process. The progressive mineralogical changes are the same as have been described for the Milendella Marbles. The abundance of scapolite over plagioclase in the calc-silicate rocks may be due to strong control of $\mu\ Cl_2$ and $\mu\ CO_2$ (Shaw, et al., 1963; Shaw, et al., 1965).

3.4. SKARN ROCKS

Introduction

Garnet bearing skarn rocks have been found in the marbles and calc-silicate rocks. Such rocks are much more common in the southern part of the Milendella Marbles and calc-silicate horizon, where swarms of intrusive pegmatite bodies occur. The skarn rock appears as lenses or pods directly associated with or near the pegmatites. Generally, they are about a metre in thickness and up to 10 metres in length. But one exceptionally large lens of about 10 m wide and about a thousand metres in length is the most prominent among all of them. It occurs in the calc-silicate rock about 1.2 km west of Milendella township. The primary nature of garnet in the marbles and calc-silicate rocks of the Cambrai-Palmer area has been doubted by Mills (1964) and White (1956, 1959) and a similar view is held by the present author due to its localized occurrence in some marbles and the associated calc-silicate rocks. The dominant mineral in the skarn rock is garnet, identified as grossularite by X-ray powder pattern (Sample Nos. 285/100 & 285/102). In thin section, garnet is seen forming coronas around clinopyroxene, scapolite and hornblende (Plate 43A,B). The scapolite is generally cloudy, but fresh and unaltered scapolite grains are also seen. The clinopyroxene may be light green in some cases and in others, it is dark green and slightly pleochroic. Optically, it belongs to the diopside-hedenbergite series. Among the amphibole minerals, hornblende is the most common. It is a light green to dark green or bluish green strongly pleochroic mineral forming xenoblastic grains or small patchy inclusions in clinopyroxene. A colourless amphibole with the fibrous appearance of tremolite, has been seen in some thin sections (e.g. 285/286). Epidote is a ubiquitous mineral in most of the skarn rocks. It usually occurs as granular aggregates or well crystallized grains (Plate 44A). Some secondary epidote is also found filling fractures in scapolite (Plate 44B). Calcite is found in small amounts as a primary mineral with well developed twinning; secondary calcite as an alteration product of scapolite and clinopyroxene is not uncommon. Most of the skarn rocks contain quartz in variable amount, either as unstrained sub-rounded grains in the matrix or granoblastic inclusions in epidote, scapolite and clinopyroxene. A notable feature is the scarcity of plagioclase in the skarn rock, whereas it is a common mineral in the marble and calc-silicate host rock. Sphene, biotite and iron ore form the accessory minerals.

3.4.1. Origin of Skarn Rocks

The field and petrotraphic observations clearly show that the skarn rocks have been produced by the metasomatic action associated with the formation of pegmatites. The garnet and epidote which forms a significant proportion of such rocks are usually absent in the regionally metamorphosed calcareous rocks. Therefore their characteristic occurrence near pegmatite bodies suggests that residual liquids released from the intrusives have played a significant role in the genesis of skarn rocks. The formation of garnet and epidote has been considered by White (1956, 1959) as a metasomatic process, which is expressed by the following reactions:

$$\underset{\text{meionite}}{\mathrm{3(CaAl_2Si_2O_8).CaCO_3}} + \underset{\text{calcite}}{\mathrm{5CaCO_3}} + \underset{\text{quartz}}{\mathrm{3SiO_2}} \longrightarrow \underset{\text{grossularite}}{\mathrm{Ca_3Al_2Si_3O_{12}}} + \mathrm{6CO_2} \quad \ldots\ldots.(1)$$

$$\underset{\text{meionite}}{\mathrm{3(CaAl_2Si_2O_8).CaCO_3}} + \underset{\text{clinopyroxene}}{\mathrm{3Ca(Mg\ Fe)Si_2O_6}} \longrightarrow \underset{\substack{\text{grossularite with some}\\ \text{almandine and pyrope}}}{\mathrm{3Ca_2(Mg\ Fe)Al_2Si_3O_{12}}} + \underset{\text{calcite}}{\mathrm{CaCO_3}} + \underset{\text{quartz}}{\mathrm{3SiO_2}} \quad \ldots\ldots.(2)$$

$$\underset{\text{meionite}}{\mathrm{3(CaAl_2Si_2O_8).CaCO_3}} + \mathrm{H_2O} \longrightarrow \underset{\text{epidote}}{\mathrm{2Ca_2Al_3O_{12}.OH}} + \mathrm{CO_2} \quad \ldots\ldots.(3)$$

The present study of these rocks indicates that the formation of garnet and epidote is much more complex than suggested by the above reactions. A feature of notable interest is that epidote and garnet are so abundant in most of the skarn rocks that the formation of these minerals in such rocks can not be accounted alone to the destruction of scapolite. Moreover, unaltered scapolite is not uncommon in some rocks. Also, most of the plagioclase is extensively altered and clouded with a fine-grained "dusty" substance. Unlike the common plagioclase of marbles and calc-silicate rocks which is more calcic (andesine to labradorite), the plagioclase in the skarn rocks is more sodic (albite-oligoclase) in composition. All these facts indicate that the formation of epidote and garnet is not only related to destruction of scapolite and clinopyroxene as suggested by White (see also the comments of Fleming, 1972, p. 130) but the anorthite molecule released by the destruction of plagioclase may also have contributed to the formation of epidote and garnet as shown by the following reactions:

$$\underset{\text{anorthite}}{3CaAl_2Si_2O_8} + CaCO_3 + H_2O \rightleftharpoons \underset{\text{epidote}}{2Ca_2Al_3Si_3O_{12}.OH} + CO_2 \quad \ldots\ldots\ldots\ldots \quad (4)$$

$$\underset{\text{epidote}}{4Ca_2Al_3Si_3O_{12}.OH} + SiO_2 \rightleftharpoons \underset{\text{anorthite}}{5CaAl_2Si_2O_8} + \underset{\text{grossularite}}{Ca_3Al_2Si_3O_{12}} + 2H_2O \quad .. \quad (5)$$

Since the anorthite plagioclase was not stable in the skarn rocks, therefore the by-product of anorthite molecule in the reaction (5) again reacted with calcite to form epidote. The formation of epidote and garnet in the above reactions have been experimentally proved by Holdaway (1966) through the hydrothermal experiments.

The following textural relationships of garnet and epidote in the skarn rocks indicate they have been formed by a number of processes involving the destruction of scapolite and plagioclase:

1. the presence of reaction rims of garnet around epidote and altered scapolite
2. reaction rims of epidote around altered scapolite and garnet
3. some epidote occurs interstitially within aggregates of garnet
4. inclusions of unaltered clinopyroxene and scapolite in garnet, indicating their stability at the time of garnet formation
5. some scapolite and clinopyroxene have direct contacts without showing any disequilibrium to form garnet, in other cases a reaction rim of garnet separates the two minerals
6. the occurrence of reaction rims of garnet around calcite.

In the light of the above facts the formation of garnet and epidote in the skarn rocks may be related to destruction of scapolite and plagioclase in hydrothermal metasomatic condition developed by the intrusion of pegmatite bodies in calcareous rocks. The field relations clearly show that skarn rock formation commenced at the end of second ductile deformation when the host marbles and calc-silicate rocks were already recrystallized by regional metamorphism. The skarn rocks are non-deformed, massive looking bodies.

Since the proportion of major minerals present in the skarn rocks of marbles and calc-silicate rocks show considerable variations, it therefore appears a complex geochemical process involving different reactions for chemically and mineralogically different calcareous rocks have operated to produce skarn rocks of diverse mineral assemblages. To get a clear picture of contact metamorphism and metasomatism for the genesis of skarn rocks, a detailed chemistry of rocks and their individual minerals will be required to test the validity of the above hypothesis (as demonstrated by Floyd, 1965, for metasomatic skarn hornfelses of the Land's End aureole at Tater-du, Cornwall). Such a petrochemical and mineralogical study on skarn rocks is beyond the scope of the present investigation.

CHAPTER 4

PETROLOGY OF THE COUNTRY ROCKS

II. QUARTZO-FELDSPATHIC SCHISTS AND GNEISSES*

4.1. INTRODUCTION

The term quartzo-feldspathic schists and gneisses has been used by White (1956, 1966a), Mills (1964), Offler (1966) and Fleming (1971) to describe quartz feldspar rich rocks of the eastern Mt. Lofty Ranges. Similar types of rocks have been mapped in the present area; they show a wide variation in biotite content reflecting the variation in the original sediments. (Such variation due to concentration of biotite in the form of layers and lenses). Therefore, for descriptive purposes, these rocks have been divided into quartzo-feldspathic, semi-pelitic and pelitic schists and gneisses based on the relative proportion of biotite in their mineral assemblages. The rocks containing biotite up to 20% are described as quartzo-feldspathic, 20-50% as semi-pelitic and more than 50% as pelitic schists and gneisses (Fig. 14). This classification has previously been used by Mills (1964) and Fleming (1971) for quartz-feldspar-biotite rocks of the Cambrai and the Kanmantoo-Dawesley areas respectively.

The quartzo-feldspathic schists and gneisses are the most abundant rocks of the present area. High grade regional metamorphism has completely recrystallized the original clastic sedimentary grains. All small-scale primary textures and structures, such as

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* The term quartzo-feldspathic schist and gneiss has been used in this thesis following Williams, Turner and Gilbert (1954) for the high grade quartzo-feldspathic rocks. The term "schist" refers to those rocks which are generally finer grained (average grain size less than .5 mm) with few or no migmatite intercalations, whereas the term "gneiss" has been adopted for coarser grained rocks (average grain size .5-1 mm) with abundance of migmatites (see White, 1956, p. 86.)

cross-bedding, graded bedding, ripple-marks etc. have been destroyed. Migmatization had an important effect on the texture and structure of quartzo-feldspathic rocks, so that generally, there are textural differences between quartzo-feldspathic rocks inside and outside the migmatite zone.

Mineralogically,, the quartzo-feldspathic schists and gneisses have a similar appearance in the Strangway Hill Formation and the Inman Hill Formation in the mapped area, except that the colour of biotite is reddish-brown to brown in the Strangway Hill Formation while a dark-brownish colour is the more common in the Inman Hill Formation.

4.2. QUARTZO-FELDSPATHIC SCHISTS OUTSIDE THE MIGMATITE ZONE

Introduction

The quartzo-feldspathic schists are medium grained rocks with well developed laminations (relict bedding) (Plate 20A). They are light coloured, greyish rocks derived mainly from arkosic type sediments. The average grain size of the schists ranges from 0.2 to 0.5 mm, though some of the porphyroblasts range up to 2.5 mm. The laminations are characterized by an abundance of biotite and rarely by concentrations of heavy minerals in dark laminations, and quartz-feldspar enrichment in the light bands. It is believed that these laminations represent original compositional (sedimentary) layering. The quartzo-feldspathic schists have well developed planar schistosity characterized by moderate to strong preferred orientation of mica and flattened coarse quartz grains. Generally, schistosity is parallel to the laminations or less commonly at moderate angles to lamination.

4.2.1. Petrography

The common minerals in the quartzo-feldspathic schists are quartz, plagioclase, K-feldspar, biotite and lesser muscovite (Plate 20B). The minor accessories are apatite, tourmaline, zircon, sphene, rutile, sillimanite and opaques. The following mineral assemblages have been observed:

(1) Quartz-plagioclase-biotite-muscovite
(2) Quartz-plagioclase-potash feldspar-biotite-muscovite
(3) Quartz-plagioclase-biotite
(4) Quartz-plagioclase-biotite-potash feldspar (± muscovite ± sillimanite)

Micrometric analyses of quartzo-feldspathic schists done by White (1966a) indicate the following mineral variations in the schists: quartz 30-50%, plagioclase 25-60%, potash feldspar 0-25%, biotite 0-20%, and muscovite 0.-10%; the major accessories are apatite and magnetite.

Plagioclase ranges in composition from An_{28-32}* but in some rocks the plagioclase may be as sodic as An_{20} and in other rocks it may go up to An_{42}. Generally, the plagioclase is clear and fresh but some are cloudy, the alteration products being fine grained sericite and unidentified "dust". Twinning is well developed in some grains but untwinned grains are plentiful. Twin lamellae are usually broad and sharp, predominantly of albite law. Zoning is sometimes seen in plagioclase grains; commonly normal but weak oscillatory zoning is also present.

K-Feldspar usually has patchy cross-hatch twinning but some grains are untwinned. The crystals are almost invariably xenomorphic, filling the interstices between plagioclase and quartz: porphyroblasts are rare. In some grains inclusions of quartz, plagioclase and mica are seen.

Quartz occurs as subrounded to lobate grains. Strain is a common feature in quartz especially in the big grains. The quartz-quartz boundaries are generally curved. Small inclusions of plagioclase and mica are common in big grains. In strongly deformed rocks, flattened lenticles of rather coarse quartz are common in the biotite rich bands.

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* The composition of plagioclase reported in this thesis were determined on the extinction angle of albite twinning normal to (010).

Biotite is the only coloured mineral in the schists. Pleochroism is light pale brown to dark brown. Pleochroic halos are occasionally visible. Biotite is moderately to strongly oriented in S_1 or S_2 but the length of biotite flakes is distinctly different in S_1 and S_2: those in S_1 are generally coarse to very coarse ranging from 0.5 mm to 3.5 mm in length whereas those in S_2 are finer and range in length from 0.2 to 0.6 mm (Plate 22B). Alteration to chlorite is common.

Muscovite is generally fine to coarse with a preferred orientation in S_1 and S_2 planes. Sometimes, post-tectonic porphyroblasts are seen. Small flakes of white mica occur in altered cores of plagioclase.

Iron ores are usually a minor constituent in most of the schists. It is randomly distributed in the rocks but sometimes is concentrated in dark micaceous layers.

Sillimanite occurs as fibrolite, seen as discrete needles crowded in muscovite and quartz.

Accessory minerals: tourmaline is a common accessory in the schists. It occurs as pale green to very pale green grains. Zircon is found as subrounded grains. Rutile is rare but is found in some rocks. Epidote is present as a secondary mineral and usually occurs in the altered cores of plagioclase.

4.3. QUARTZO-FELDSPATHIC GNEISSES IN THE MIGMATITE ZONE

Introduction

The migmatite zone is defined by the abundance of leucocratic vains in the quartzo-feldspathic gneisses. A line dividing the quartzo-feldspathic schists from migmatites is drawn to indicate the abundance of migmatization and gradual increase in the grain size of the quartzo-feldspathic rocks (Map 1). The gneisses generally show a large-scale vein formation and a gradual increase in grain size from north to south. The mineral assemblages observed in the gneisses within and outside the migmatite zone are the same. Sillimanite (fibrolite) is much more common in this zone, occurring in the field as small white tufts several millimetres in length, parallel to their mesoscopic lineation. The metamorphic layering is much more pronounced and widely spaced in the gneisses after the notable increase in grain size of original laminations (Plate 21A). These quartzo-feldspathic rocks are interbedded in migmatites and can be differentiated from the migmatized gneisses only by the absence of veins, otherwise both rocks are very similar in appearance, texture and mineral assemblages. The regular colour bands are characteristic features of these rocks, marked by enrichment of biotite and quartz-feldspar in alternate dark and light layers. The layers may vary in width from a few millimetres to 10 cm and can be traced for distances up to a hundred metres.

4.3.1. Petrography

The major components of the gneisses are quartz, plagioclase, K-feldspar, biotite and muscovite. In a number of rocks, a small amount of fibrolitic sillimanite is also present. The following variations in mineral proportions of these rocks are found:

quartz - 30-55%; plagioclase - 25-60%;
potash feldspar - 0-25%; biotite - 5-20%
muscovite - 0-10%.

The gneisses are well crystallized rocks with excellent mica schistosity. The average grain size is more than 0.5 mm.

Plagioclase occurs as zenoblastic grains with irregular boundaries against other feldspar and quartz grains. Albite twinning is common but untwinned grains are equally abundant. Bent and broken twin lamellae indicative of post-crystallization deformation, are occasionally seen. Small inclusions of quartz and mica are frequently seen in bigger grains (Plate 21B). Weak zoning, generally normal type, may be found in some grains. The composition of plagioclase is generally oligoclase but in some rocks it may be acid andesine as in the quartzo-feldspathic schists outside the migmatite zone.

Quartz grains are highly strained and quartz-quartz boundaries are sutured. Inclusions of plagioclase and biotite are common in porphyroblasts of quartz. In places, quartz appears to replace plagioclase and mica.

Potash feldspar is less abundant than plagioclase, and ranges from almost nil in some rocks to as much as 20% of the total rock. This great variation in abundance has also been noted by White (1956) and Mills (1964). The cross-hatch twinning indicative of microcline is common but untwinned grains are also present.

Biotite is normally light yellowish brown to dark brown or reddish brown, occasionally with pleochroic halos. Biotite flakes usually form a matrix between quartz and plagioclase grains. In some rocks, however, the biotite grains are more idioblastic (Plate 22A). Generally, they have a strong preferred orientation in S_1 plane. Those rocks which exhibit secondary schistosity have fine biotites oriented in S_2 plane (Plate 22B).

Muscovite is less common than biotite. Some post-tectonic porphyroblasts have been seen in a few thin sections (Plate 23A). Second generation muscovites occur as ragged flakes or fine grained sericites in altered core of plagioclases.

Accessory minerals include iron ores, rutile, zircon, sphene, tourmaline and apatite. Epidote is found as a secondary mineral in the altered plagioclase. Chlorite is associated with some of the altered biotite.

4.4. SEMI-PELITIC/PELITIC SCHISTS AND GNEISSES

Introduction

The semi-pelitic and pelitic rocks are characterized by the abundance of biotite in their mineral assemblages. Unlike the quartzo-feldspathic rocks which are rich in quartz and feldspar, these rocks have a high proportion of mica ranging from about 20-50% in semi-pelitic and more than 50% in pelitic schists and gneisses (Fig. 14). Those rocks which are rich in sillimanite (fibrolite) are treated in the next section as aluminous pelitic schists, although fibrolite sporadically occurs as a minor constituent of the semi-pelitic and pelitic rocks.

The semi-pelitic and pelitic schists and gneisses are interbedded with quartzo-feldspathic schists and gneisses, individual layers ranging in thickness from a few centimetres to several metres. The gradational changes from quartzo-feldspathic rock to semi-pelitic and pelitic rock is marked in the field as gradual increase in relative proportion of biotite without any significant change in texture. When gradational changes are confined to a small scale, they appear as continuous bands or layers marked by light quartz feldspar rich bands and dark biotite rich bands. The field observations indicate that compositional changes from quartzo-feldspathic to semi-pelitic or pelitic rocks are due to variation in original composition of sediments. There is no evidence either of metasomatism or metamorphic differentiation which might have changed the gross chemistry of quartzo-feldspathic rocks to form semi-pelitic or pelitic rocks. Rather, uniformity in thickness for long distances and gradual changes along and across the strike lines suggest original compositional variation in sediments. The semi-pelitic and pelitic schists and gneisses form a small portion of quartzo-feldspathic rocks in which they are interlayered on a variable scale.

The semi-pelitic and pelitic schists often contain muscovite porphyroblasts, randomly distributed and transecting the planar schistosity of the rocks, indicating that these porphyroblasts are post-tectonic products. In the field, muscovite porphyroblasts appear as silvery white stumpy masses on the outcrops. As pointed out earlier, generally biotite is more abundant than muscovite. The well oriented flakes of biotite and muscovite impart a strong penetrative schistosity in schists and gneisses.

The following are major mineral assemblages of semi-pelitic and pelitic schists.

(1) Quartz-plagioclase-biotite
(2) Quartz-plagioclase-biotite-muscovite
(3) Quartz-plagioclase-potash feldspar-biotite-muscovite
(4) Quartz-plagioclase-potash feldspar ± biotite
(± muscovite ± sillimanite)

4.4.1. Petrography

The most common minerals in the semi-pelitic and pelitic schists are quartz, plagioclase (oligoclase-andesine), and biotite with less common potash feldspar and muscovite as major constituents. The accessory minerals are apatite, zircon, sphene, tourmaline, and

iron ores. The texture is granoblastic with average grain size ranging from 0.3 to 0.6 mm. In mica rich rocks, mica exhibits a strong preferred orientation. The presence of sillimanite appears to indicate breakdown of muscovite. Muscovite is either completely eliminated to give rise to sillimanite and potash feldspar or partially altered to form sillimanite and potash feldspar in association with skeletal muscovite. These petrographical observations indicate that destruction of muscovite in the presence of quartz was not only dependent upon the rising temperature and total pressure of metamorphism, but that some other factors (i.e. the activity of water and anorthite content of plagioclase) may have some bearing on the elimination of muscovite. Sillimanite most commonly occurs as bundles of fibrolite or trails of discrete needles in skeletal muscovite and quartz (Plate 23B). Often fibrolitic sillimanite has intergrown with biotite. These marginal fibrolites are frequently crenulated or contorted whereas there is no sign of deformation in biotite flakes. The close association of fibrolite with biotite often results from extensive replacement of biotite by fibrolite during the nucleation process. In such cases, biotite has been completely surrounded by rapidly growing needles of fibrolites piercing the biotite crystals from all directions.
Minute prisms of sillimanite are often found within biotite (Plate 24A).

Other petrographical features of semi-pelitic and pelitic schists and gneisses are described below.

Quartz appears as clear granoblastic grains to fine aggregates in the rock matrix. Granoblastic quartz varies both in abundance and grain size. It usually shows sutured boundaries at quartz-quartz and quartz-plagioclase junctions. It forms a mosaic texture between plagioclase and biotite in less micaceous rocks. Elongated grains of quartz are common in micaceous schists expressing dimensional orientation parallel to schistosity planes. The bigger grains of quartz contain small inclusions of mica and plagioclase.
Plagioclase is generally well twinned on albite law but untwinned or irregularly twinned grains are also common. Some grains

occasionally show bent and broken twin lamellae. A few of the large grains show zoning; it may be sharp normal zoning or irregularly developed patchy zoning. The composition of plagioclase in most rocks ranges from An_{28} to An_{34} but in some rocks it may be as high as An_{39} or as low as An_{20}. In some sillimanite bearing pelitic gneisses interbedded in quartzo-feldspathic rocks, the composition of plagioclase tends to be more calcic (An_{35-39}) than it is in rocks which lack sillimanite (An_{28-32}). This calcic tendency of plagioclase is particularly notable in those rocks in which perthitic K-feldspar is in equilibrium with sillimanite. Hence, the sillimanite which appears to be formed by a complex reaction of Guidotti (1963), the relative enrichment of anorthite content in coexisting plagioclase may have resulted due to loss of some albite component in the development of perthite in K-feldspar.

Potash feldspar has cross-hatch twinning, and occasionally shows fine hair perthites. It is usually fresh and unaltered with xenoblastic crystal form. It is a common mineral in sillimanite bearing schists indicating its genesis during the formation of sillimanite.

Biotite occurs in two different varieties, one dark brown and the other reddish brown. The reddish brown variety is the typical mineral associated with the Strangway Hill Formation metasediments, whereas the dark brown biotite is commonly found in the Inman Hill Formation metasediments. Flakes of biotite range from 0.3-3.5 mm in length with strong preferred orientation along two visible schistosity planes (S_1 and S_2). They seldom show pleochroic halos.

Muscovite Most muscovite occurs as large, clear flakes with strong preferred orientation. Intergrowths of muscovite with biotite are commonly seen. In addition large muscovite grains are randomly oriented as post-tectonic porphyroblasts containing inclusions of quartz, plagioclase and biotite. Some of the large porphyroblasts (3.5 mm) are considered to be formed by metasomatism, similar to those found in the mica schists described in Section 4.9. Secondary muscovite has been formed in altered cores of plagioclase grains during a static stage of metamorphism.

Sillimanite is found in three forms. It occurs as faserkiesel (bundles of fibres forming a mat structure), trails of discrete needles enclosed in partially altered muscovite or as numerous tiny needles crowded in quartz grains. The third type is less common and is found as small euhedral prismatic sillimanite crystals in biotite (see Sect. 4.6.).

Most common accessory minerals are apatite, as euhedral to subhedral crystals. Tourmaline is greenish in colour with weak pleochroism. Iron ores are another important minor constituent of semi-pelitic and pelitic schists. Usually they are concentrated as disseminated tabular crystals in biotite rich layers. Zircon, sphene and rutile are other common accessory minerals.

4.5. ALUMINOUS PELITIC SCHISTS

Introduction

The aluminous pelitic schists are characterized by an abundance of fibrolite which is the only stable aluminium silicate occurring in the present area. Biotite, which is plentiful in semi-pelitic (20-50%) and pelitic (over 50%) rocks, is generally less abundant (10-20%) in aluminous pelitic schists.

The aluminous pelitic schists occur as thin bands of a few millimetres to interbeds as much as 10 metres thick in quartzo-feldspathic rocks. In the field, they can be traced out to hundreds of metres along the strike line and then more or less abruptly die out in the country rock. The schist has a characteristic appearance, containing fibrolite as small tufts or faserkiesels, about 0.5 cm to 4 cm in length (Plate 24B,C). Some of these tufts and nodules are commonly elongated parallel to S_1 schistosity. This typical appearance of sillimanite in the schists assists its easy recognition during the mapping. Even on highly weathered outcrops, the white faserkiesels project as eye shaped nodules on the smooth surface of the rock. The schists contain a good schistosity defined by parallel orientation of micas, but compositional layering is generally less common than in other schists. The texture is even grained consisting of quartz, feldspar and mica. Occasionally, thin laminations may develop due to concentration of biotite along the schistosity plane. Such laminations are poorly defined and lack continuity.

The following variations in the proportions of the various minerals are quartz, 20-60%; plagioclase, 5-30%; biotite, 10-20%; muscovite, 5-25%; fibrolite, 5-25%. Secondary muscovites are common in most of the rocks probably as a result of a post-tectonic late stage muscovitization process in a relatively static environment. The abundance of water released from the dehydration process during regional metamorphism will be available to promote the potassium metasomatism to form late stage muscovite (cf. Eugster, 1970). Potash feldspar is a common mineral in most of the schists. The significance of its occasional absence in sillimanite bearing schists is discussed later in this chapter. The common accessory minerals are chlorite, apatite and iron ores, with less common tourmaline, rutile, sphene and zircon. The following mineral assemblages have been observed in the aluminous pelitic schists.

(1) Quartz-plagioclase-potash feldspar-biotite-fibrolite
(2) Quartz-plagioclase-potash feldspar-biotite-muscovite-fibrolite
(3) Quartz-plagioclase-biotite-fibrolite.

4.5.1. Petrography

Texturally, the schist is medium grained rock, with average grain size 0.2 to 0.4 mm. Biotite forms a good schistosity defining the S_1 plane. Primary muscovites of skeletal shape are not uncommon even in high grade metamorphic conditions, commonly associated with fibrolites, but some late stage muscovite also occurs. An interesting feature of petrological interest is the formation of sillimanite nodules, or faserkiesels, in the schist. Such nodules mainly consist of fibrolite and quartz with skeletal muscovite containing numerous discrete needles of sillimanite, but the surrounding matrix consists of K-feldspar, biotite and plagioclase. A striking thing about the constituent minerals of the nodules is the absence of K-feldspar in immediate contact with fibrolite mats. These fibrolite nodules are intimately associated with numerous quartz grains in which numerous tiny needles of fibrolite are projected in all directions or bundles of fibres making a felt structure growing on the quartz grains (Plate 25A). The petrographic evidence indicates that the growth of some nodules took place at the end of the first tectonic movement so that they transect the original schistosity S_1 defined by biotite orientation (Plate 25B). The situation is thus similar to that described by Losert (1968) who suggested that nodules consisting of sillimanite were formed after the main tectonic movements in a relatively static environment. Another interesting feature in the fibrolitic nodules of the present area is the appearance of skeletal muscovite filling the interstices between the quartz grains (Plate 26A). The appearance of such muscovite has also been reported by Watson (1948) in the migmatites of Kildon, Sutherland (see also Chinner, 1961 and Tozer, 1955).

Quartz is common as equigranular subrounded grains but less commonly is found as lobate grains elongated parallel to S_1 schistosity. Quartz-quartz and quartz-feldspar boundaries are usually curved. The quartz forms a mosaic structure between plagioclase and biotite and generally occurs as strain free grains.

Plagioclase is usually fresh and unaltered. Well developed twinning is seen and there are also discontinuous and patchy twin lamellae in some grains. Many grains are untwinned. The composition of plagioclase measured on twinned grains ranges from An_{28-35}.

Potash feldspar invariably occurs as xenomorphic granoblasts, filling interstices between quartz-plagioclase matrix. The cross-hatch twinning is well developed. Fine hair perthites are commonly seen. Inclusions of quartz and biotite with fine grains of opaques are not uncommon. Generally, potash feldspar is less common in country rocks (quartzo-feldspathic rock) and its abundance in aluminous pelitic schists is considered to be related to breakdown of primary muscovite to form K-feldspar and sillimanite. Kink structures are frequently seen in some potash feldspar grains.

Biotite (X = light brown, Y = Z = dark brown to reddish brown). Pleochroic halos are occasionally seen. Biotite is usually absent in the fibrolitic nodules, except in rare cases. Inclusions of rutile are often found in the biotite.

Muscovite is generally found as skeletal shaped crystals with ragged boundaries with numerous tiny needles or sheaves of fibrolites. Some stumpy porphyroblastic flakes are also found not containing any fibrolite. In general, small flakes of muscovite are associated with oriented biotite indicating its primary nature.

Fibrolite occurs in different forms and appearances. Swarms of fine, fibrous and felted fibrolite either grow on the quartz grains or enormous numbers of discrete needles are enclosed in skeletal muscovite. Usually, the porphyroblastic fibrolites are intimately associated with quartz aggregates riddled with clusters of minute needle shaped sillimanite crystals. Quite often, the sheaves of fibrolitic sillimanite cluster around the irregularly shaped iron ores (Plate 26B).

Among accessory minerals, iron ores are the most common in the schists. They are found as irregularly shaped grains or in subhedral to euhedral crystal forms. Small reddish brown crystals of rutile are usually associated with biotite or randomly distributed in rock matrix. Sphene is generally found as subhedral to anhedral form crystals. Tourmaline is frequently seen as an olive green weakly pleochroic mineral. Apatite is ubiquitous as a small anhedral shaped grain to euhedral prismatic crystal.

4.6. ORIGIN OF FIBROLITE

Fibrolitic sillimanite is an important constituent of the aluminous pelitic schists, but it is also found in variable amount in the quartzo-feldspathic rocks, semi-pelitic and pelitic schists and gneisses. Some pelitic rocks contain minute prismatic crystals of sillimanite included or associated with biotite. Textural evidence indicates that generally the muscovite is unstable in the presence of fibrolite although it did not completely disappear in some rock assemblages. Recent literature has tended to discount the first appearance of fibrolite as a potential indicator of the sillimanite zone because fibrolite is frequently found in the andalusite-staurolite zone. This problem has been discussed in detail by Fleming (1971) with reference to the metamorphic belt of the Mt. Lofty Ranges. In the Dawesley-Kanmantoo area, Fleming has suggested that the fibrolite has resulted by breakdown of andalusite. The evidence supported by him is the absence of potash feldspar in the rocks and any direct association of iron ores with fibrolite.

The fibrolites of the present area show no evidence that they have been formed by breakdown of andalusite (cf. Mills, 1964; Fleming, 1971). Since andalusite has not been recorded in the present area, it is unlikely to be formed by this reaction. The textural and mineralogical assemblages containing fibrolite appear to indicate that it has been formed by different processes in different types of rocks. In the quartzo-feldspathic rocks, muscovite is generally unstable and breaks down to form fibrolite with production of K-feldspar as a by-product. The plagioclase may or may not be directly involved during the destruction of muscovite in the presence of quartz and a hydrous phase. In some rocks, hair perthite is present in K-feldspar with closely associated fibrolite, but in others no such perthitic development was found showing no contribution of plagioclase in the genesis of fibrolite. The following two reactions are responsible for the formation of fibrolites in most of the rock assemblages.

$$\text{muscovite} + \text{quartz} \rightleftharpoons \text{K-feldspar} + \text{sillimanite} + H_2O \text{ .... } \quad (1)$$

(Turner and Verhoogen, 1960)

$$\text{muscovite} + \text{quartz} + \underset{\text{(component of plagioclase)}}{\text{albite}} \rightleftharpoons \underset{\text{(containing perthite)}}{\text{K-feldspar}} + \text{sillimanite} + H_2O \text{ ... } \quad (2)$$

(Evans and Guidotti, 1966)

Relicts of primary muscovite are often associated with fibrolite even in the highest grade of the migmatite zone. The stability limit of muscovite is shown in Figure 15A, B for variable P-T conditions. Evans' curve intersects the low temperature melting curve of granite at about 3.5 kb pressure and 650°C (Fig. 15B). As the migmatites of the present area have hardly reached the partial melting conditions (see Section 5.1), erratic stability of muscovite is probably due to variations in the water pressure (cf. Winkler, 1970, Table 5). The fibrolite in nodules has been formed by two different processes as exemplified in the following two cases:

(1) In thin section, the coarse mat of fibrolite is closely associated with quartz and/or skeletal muscovite, both of which contain a large number of small needles. The striking thing is the absence of potash feldspar from the faserkiesels, although it is abundant in the surrounding matrix along with quartz, plagioclase and biotite. The potash feldspar occurs in variable shapes, in grains up to 5 mm in length filling interstices in between quartz, plagioclase and biotite. Inclusions of quartz, plagioclase and biotite are very common. Unlike the two reactions (1) and (2) described earlier, muscovite did not react with quartz in the genesis of these fibrolites, otherwise the by-product potash feldspar should have been found in close association with fibrolite in the faserkiesels. This leads to the view that some metasomatic or metamorphic differentiation process may be responsible for the genesis of fibrolite and quartz in the faserkiesels and potash feldspar in the rock matrix at a late stage. Consideration has been given to the role of a fluid phase which may be involved in metasomatic activity. Dissolution of microcline at 500°C, 1-2 kb has been demonstrated by Morey and Hesselgesser (1951) and therefore, a fluid phase is likely to have some potassium content in dissolved form. In the light of the above data, Eugster (1970) suggested the following reaction for the nodules formation containing sillimanite and quartz with absence of K-feldspar.

$$2KAlSi_3O_8 + 2H^+ \rightleftharpoons Al_2SiO_5 + 5SiO_2 + 2K^+ + H_2O \quad \ldots\ldots\ (3)$$

K-feldspar sillimanite quartz

He further states that once the nucleation of sillimanite is initiated by the process of dealkalization, muscovite may also be likely to be involved in the formation of sillimanite by the following reaction:

$$2KAl_2AlSiO_3O_{10}(OH)_2 + 2H \rightleftharpoons 3Al_2siO_5 + 3SiO_2 + 2K^+ + 3H_2O \quad \ldots\ldots\ (4)$$

muscovite sillimanite quartz

The petrographic evidence for genesis of fibrolites in the present area appears to support the above two reactions (3) and (4). The presence of relicts of muscovite in some assemblages also lends support to the hypothesis of destruction of muscovite by ionic reaction in the presence of a hydrous phase (see also Carmichael, 1969, reaction 5). The potash content released by dealkalization reactions (3) and (4) will probably cause the formation of K-feldspar in the matrix at a later stage after the formation of fibrolite in the faserkiesels. This again may be due to metasomatic activity of the fluid phase, but the actual mechanism that forms K-feldspar in the matrix only, and not in the faserkiesels, is not clear. White (1956) has described a similar type of pelitic schists, south of Palmer, containing faserkiesels rich in fibrolite and quartz only and abundant K-feldspar in the matrix. He proposed a metasomatic origin for fibrolitic sillimanite in the schist.

(2) Some mats of fibrolites contain numerous inclusions of iron ores with closely associated quartz grains crowded by an enormous number of small needles of sillimanite. Occasionally skeletal muscovite with fine needles of sillimanite is also seen in direct contact with fibrolite. Like the previous example, again the potash bearing minerals K-feldspar and biotite are absent from the fibrolite tufts but are common in the surrounding matrix. The growth of fibrolite tufts or faserkiesels transects the planar schistosity indicating its origin later than the recrystallization of rock matrix. The textural and mineralogical assemblages of these rocks are similar to those described by Watson (1948), Tozer (1955), Francis (1956) and Shelley (1968). The close association of iron ores with fibrolite indicates that probably biotite has contributed to the genesis of fibrolite, according to the following reactions (Chinner, 1961).

$$\underset{\text{biotite}}{(OH)_8K_4Mg_4Fe_5Al_8Si_{10}O_{40}} \longrightarrow \underset{\text{sillimanite}}{4Al_2Sio_5} + \underset{\text{quartz}}{10SiO_2} + (H_8K_4Mg_4Fe_5O_8) \quad \ldots \quad (5)$$

As pointed out by many petrologists, the fate of expelled elements (K, Fe, Mg) after destruction of biotite can pose a serious problem in some mineral assemblages. In the present rock, K and Fe appear to be used up to form potash feldspar and iron ores, but the fate of

Mg is uncertain. As a part of the explanation it may be suggested that Mg may be re-established in the stable biotite of the matrix (cf. Francis, 1956). Tentatively, therefore, it can be concluded on the petrographical evidence that the fibrolite with its close association with iron ores and quartz has also been formed by breakdown of some biotite in the present area.

4.7. GRANITIC GNEISSES

Introduction

A small composite body of medium to coarse grained gneisses is well exposed near the Cooke Hill trig, in the northeaastern part of the area studied (Map 1). The body is lenticular, striking in north-south direction parallel to bedding trend of the adjacent quartzo-feldspathic schists. These gneisses occupy the core of a tight antiformal structure whose eastern limb has been cut off sharply by the Cooke Hill thrust fault. The length of the gneissic body is about 1 kilometre and it attains a maximum width of 250 m towards the southern end. Most of the outcrop is covered by tor-like boulders of varying sizes. Where the contact of the gneisses with the country rock can be seen, it appears to be conformable with the adjacent schists. A good section which is well exposed along the Mt. Pleasant-Sanderston road provides the best evidence for the concordant nature of gneisses.

The gneisses generally comprise two different varieties based on textural variation. One type is coarse grained and the other is medium grained, both are highly lineated and weakly foliated rocks. The foliation is less marked in coarse grained gneisses due to the small amount of biotite, but the medium grained gneisses have thin laminations rich in biotite. These laminations are widely spaced from less than 10 cm to more than 30 cm in thickness. The coarse grained gneiss makes prominent outcrops in which the medium grained gneiss appears as interbands or layers, ranging in thickness from about 10 cm to some metres. The layers may be traced up to hundreds of metres parallel to regional strike of the country rock. The contact between the coarse grained and medium grained gneisses is sharp but gradational.

Both gneisses are pink or deep flesh coloured in appearance and have a porphyroblastic texture. In hand specimen, the prominent porphyroblasts of quartz and feldspar can easily be seen with the naked eye. A strong linear structure is produced by oriented elongated grains of quartz and feldspar which are distinctly visible on the outcropping surfaces. Apart from the texture, two types of gneisses have similar mineral assemblages. These gneisses have been collectively called "granitic gneisses". The term "granitic" has been used here to express the strong granitic appearance shown by their texture and tor-like outcrops. In addition the absence of mineral layering contrasts sharply with the well laminated quartzo-feldspathic gneisses.

4.7.1. Petrography

The major components of the gneisses are quartz and plagioclase with subordinate amounts of potash feldspar and biotite. Primary muscovite is less common, but secondary muscovite is associated with altered plagioclase. Accessory minerals are apatite, sphene, zircon and iron ores. Epidote and chlorite are secondary minerals as alteration products of feldspar and biotite respectively.

Texture

In thin section, the coarse grained gneiss consists of a medium grained matrix (average grain size 0.5 mm) with abundant porphyroblasts of quartz and plagioclase (size range 1-5 mm) (Plate 32A). This abundance of porphyroblasts in the gneisses produces the coarse texture seen in hand specimen. The medium grained gneisses have an average grain size of 0.5 mm but have fewer porphyroblasts up to 2 mm in diameter.

Mineralogical composition

Plagioclase is the dominant mineral in both the medium and coarse grained gneisses. It ranges from about 50 to 60% of the total rock. The next abundant mineral is quartz which makes up 40-50%. The potash feldspar is generally less abundant in the coarse grained gneisses but it may be present up to 20% in some of the medium grained gneisses (Plate 32B). A brief account of individual minerals is given below:

Quartz is subrounded in shape, ranging in size from drop-like inclusions in plagioclase to xenoblasts up to 5 mm. It has marked undulose extinction. Deformation lamellae are frequently seen in porphyroblastic grains (cf. Christie et al., 1964; Carter et al., 1964; Ernst & Blatt, 1964). The strong deformation features are evident by parallel sets of lamellae (Plate 33A); sometimes two sets occur at right angles to each other. Small inclusions of biotite and plagioclase are seen in bigger grains (Plate 33B).

Plagioclase shows irregular outlines with curved boundaries against quartz. Both twinned and untwinned grains are found but the twinned are more abundant. Dusty appearance due to incipient alteration is a common feature. The highly sericitized plagioclase contains abundant white mica and frequently epidote grains. A weak normal zoning is often present. The composition of the plagioclase is An_{16-30}.

Potash feldspar is more irregular in its shape than plagioclase. It may be common in the matrix, but is rarely seen as porphyroblasts. Cross-hatch twinning is generally well developed. The embayment of potash feldspar against plagioclase or vice versa is common. Inclusions of quartz and plagioclase are seen in potash feldspar. It ranges from about 0 to 20% of the total rock.

Biotite (X = light brown, Y = Z = dark brown). It generally occurs as clusters of flakes with weak preferred orientation. Usually, biotite forms fine slender flakes 0.1-0.5 mm in length; occasionally it reaches 1 mm. It ranges in abundance from about 5 to 20%, but is more abundant in the biotite rich laminations.

Among the accessories, muscovite is predominant. It is commonly associated with biotite. Sphene is always anhedral in shape. Iron ores are irregularly distributed in the matrix or are found to be associated with biotites. Under reflected light, some have an altered margin of limonitic to hematitic composition. Small slender subhedral zircons are commonly found. Tourmaline and rutile are rare.

4.7.2. Origin of the granitic gneisses

The granitic gneisses are strikingly different in appearance, texture and colour from the adjacent quartzo-feldspathic schists. The quartzo-feldspathic schists are finer grained, greyish coloured rocks with more biotite and a more calcic plagioclase than the granitic gneisses.

The following possible hypotheses have been proposed by various authors for the origin of granitic gneisses:

1. magmatic intrusion into sediments and later deformation by tectonic movements
2. granitization (metasomatism) of the country rock
3. isochemical metamorphism leading to crystallization of an original sediment with different composition from the adjacent gneisses.

The following field and petrographic evidence argue against a magmatic origin:

1. no xenoliths of the country rock have been found in the gneisses.
2. the gradational contact between the gneisses and the country rock.
3. intercalated bands and layers of medium grained gneisses in the coarse grained gneisses.
4. abundance of non-twinned feldspar.
5. the normative composition of an analysed coarse grained gneiss (A285'144) plots on the synthetic granite system (Ab-Or-SiO_2) outside 500 bars pressure field, but closer to some of the metasedimentary rocks (see Fig. 28A and Table 3D).

Metasomatic granitization of a quartzo-feldspathic schist to give rise to the porphyroblastic granitic gneisses needs the following explanations:

1. there is a marked tendency towards impoverishment of biotite in the granitic gneisses as compared with the quartzo-feldspathic schists.
2. if the silica, alumina, soda and lime, necessary for the formation of quartz and plagioclase, were introduced from an outside source into an original quartzo-feldspathic rock to dilute the relative

> proportions of biotite, some reaction rims around the primary plagioclase would be expected to appear in some grains. There are no such reaction rims as albite forming the outer margin of plagioclase.

Due to the low content of biotite in the granitic gneisses, it was thought that probably some rock of sandstone composition may have been granitized by metasomatizing fluids released from nearby intrusive bodies (e.g. Härme, 1959; Gavelin, 1960; Mehnert, 1968, p. 288-302.), These authors have demonstrated that migration of certain elements (especiallly, K, Na, Ca, Al and Si) through intergranular diffusion can change quartzite into a granitic rock by "feldspathization". Such a process generally forms abundant porphyroblasts of potash feldspar with rim feldspathization around the original quartz and plagioclase. In the Cooke Hill area, porphyroblasts of quartz and plagioclase are abundant in the granitic gneisses but potash feldspar rarely appears as porphyroblasts as is commonly found in granitized rocks (see also Härme & Laitala, 1955; Härme, 1958; Simonen, 1948; Anderson, 1963). Thus it seems unlikely that a similar process was involved in the genesis of granitic gneisses. There are cases where sodium rich hydrothermal solutions can form porphyroblasts of plagioclase with microperthite and quartz during granitization (cf. Misch, 1949). In this case replacement perthite is a characteristic feature of metasomatic rock, whereas in the present granitic gneisses, potash feldspar is generally less common and no replacement of rim perthite in the plagioclase has been seen.

These granitic gneisses bear a strong resemblance to the Rathjen Gneiss (White, 1966b) which occurs about 7 km to the east. The Rathjen gneisses extends for about 12 km roughly N-S and parallel to the grain of the country. The outcrop is about 2-3 km wide but White believes it to be 'a gently folded sheet of the order of 800 metres in thickness'. The Cooke Hill granitic gneisses are only 250 m across and 1000 metres long. The mode of occurrence suggests that both bodies are due to metamorphism of a thick relatively homogeneous lens of sediment. In each instance the absence of the small-scale layering which is so characteristic of the 'normal' gneisses of the area leads to massive outcrops and to 'tor'-like boulders.

The Cooke Hill granite gneisses share with the Rathjen gneiss a richness in quartz (35-40%) but differ from the Rathjen gneiss in having less biotite and much less potash feldspar. On the other hand the analysis of one of the coarse granitic gneisses (144) is similar to that of one of the more leucocratic metasediments (139). This suggests that the granitic gneisses are derived from the same sediment source as the adjacent metasediments, but with the micaceous materials winnowed out during sedimentation leaving a quartz-plagioclase rock. The medium grained layers in the granitic gneiss represent stages during which some micaceous material accumulated and formed mica rich layers in the rock. These observations all point to a sedimentary origin for the 'granitic gneisses'.

4.8. PORPHYROBLASTIC POTASH FELDSPAR GNEISSES

Some pelitic and semi-pelitic gneisses contain bands or lenses of gneisses rich in potash feldspar. They range in thickness from a metre up to 6 metres and can be traced along the strike line to distances ranging from 6 metres to about a hundred metres (e.g. Location 182062). These gneisses form the integral part of the host rock except the occurrence of potash feldspar gives a characteristic porphyroblastic appearance. There is no well defined boundary line separating the potash feldspar gneisses from the country rocks and the groundmass texture is almost the same in both units. They can be easily identified in the field by the typical appearance of the greyish coloured potash feldspar porphyroblasts which project out from the outcropping surfaces as stumpy knots.

In thin section, all potash feldspar gneisses have similar mineral assemblages but potash feldspar may be found both as syntectonic or post-tectonic minerals (Plate 45A,B). The common minerals in the gneisses are quartz, potash feldspar, plagioclase, biotite and lesser muscovite. Apatite, sphene, zircon, tourmaline and iron ores are accessory minerals.

Specimen 285/72 (location 184059) shows poikiloblasts of potash feldspar containing numerous inclusions of quartz, plagioclase and biotite or some small granules of iron ores. Generally the poikiloblastis are 1-3 mm in diameter; some may reach 5 mm. Helicitic texture is well developed with the poikiloblasts having trails of oriented biotites as internal S-surfaces (S_i). Many porphyroblasts contain rows of inclusions constituting an internal schistosity (S_i). In many cases the S_{Ii} runs straight through the crystals and is continuous with the external schistosity (S_e); in other porphyroblasts, the S_i lie at small angles to as much as 60^o to the S_e (Plate 46A). Careful study of the fabrics of these rocks indicates that porphyroblasts grew by syntectonic crystallization during F_1 having an internal schistosity - S_i inherited from an earlier existing schistosity, but the latter deformation during F_2 rotated some of the porphyroblasts, leaving the S_i at an angle to S_e.

There are also evidences of post-tectonic crystallization of potash feldspar in some other gneisses. One sample 285/445 (Location 195992) was collected from a small band of potash feldspar gneiss adjacent to a small granitic intrusive. In thin section, megacrysts of potash feldspar are abundant with well developed cross-hatch twinning (Plate 45B). Petrographic studies indicate that some porphyroblasts of potash feldspar have been formed by post-tectonic crystallization displacing the pre-existing S_1 schistosity, whilst other syn-tectonic grains with helicitic texture (Plate 46B) are also seen. It thus appears that potash feldspars of two generations have developed in the gneisss, the earlier one as syn-tectonic crystallization of pre-existing mineral in the parent rock and the later one as post-tectonic crystallization from the metasomatic alkali fluids soaked in the gneiss from the adjacent granitic body.

Apart from the similar mineral assemblages in both the porphyroblastic potash feldspar gneisses and the host rock in which they appear as bands and lenses, no significant difference in relative proportions of quartz, plagioclase and biotite is found. The main characteristic difference is the abundance of potash feldspar in the porphyroblastic gneisses and its scarcity in the host semi-pelitic or pelitic gneisses. The occurrence of potash feldspar in the form of porphyroblasts is interesting inasmuch as they grew in two stages, one in the main deformation (D_1) when pre-existing smaller grains of potash feldspar recrystallized to form larger grains enclosing the dimensionally oriented grains of biotite, quartz and plagioclase. Interstitial potash feldspar in the matrix of these rocks is rare. In the second stage, the formation of porphyroblastic potash feldspar is considered to be a metasomatic process deriving its source materials from outside the host rock. However, the nature of the porphyroblasts indicates that tectonic control was an important factor during the development of porphyroblastic potash feldspar in the gneisses. During the latter stage of tectonic deformation, a drop in temperature generally occurs. Under such conditions, pore fluids are capable of potassium metasomatism leading to the formation of post-tectonic porphyroblasts of potash feldspar (e.g. Eskola, 1956; Marmo, 1956; Cannon, 1964).

The genesis of post-tectonic porphyroblasts of potash feldspar near the intrusives seems to be related to a potash metasomatism by a fluid phase released from the nearby granitic intrusives. Therefore, the residual fluids rich in potash can impregnate the adjacent country rock resulting in the formation of potash feldspar. This hypothesis can be supported from the field relationships as in the case of Sample No. 285/445 which is associated with a small granitic intrusive. A graphic intergrowth of the quartz-feldspar relationship has been observed probably indicative of a magmatic role in the genesis of potash feldspar (cf. Tuttle & Bowen, 1958, p. 84; Gates & Scheerer, 1963 etc.).

4.9. MICA SCHISTS

Two roughly parallel beds of mica rich schists, about 70 m in width, outcrop in the northern part of the area (Map 1). These beds can be traced out in the field along the bedding plane of the quartzo-feldspathic schists. The prominent feature of these schists is the occurrence of muscovite porphyroblasts, ranging in length from 1-3 mm. The porphyroblasts are randomly distributed in a fine grained matrix and mainly consist of quartz, plagioclase, muscovite and biotite. Generally, the muscovite porphyroblasts transect the schistosity as defined by oriented mica (Plate 47A), thus indicating a post-tectonic origin for the porphyroblastic muscovite.

Texturally and mineralogically the mica schist is similar to the adjacent quartzo-feldspathic schists except that muscovite is much more abundant. In the field, mica schist is distinguished from the quartzo-feldspathic schist by the presence of large silvery muscovite porphyroblasts. The contact between the mica schist and the adjacent country rock is diffuse but the internal planar schistosities between the two rocks are parallel to each other. No prominent compositional banding and layering is found in the mica schist.

In thin section, the texture of the mica schist is seen to be finer (0.2 - 0.3 mm) but gradually increases towards the southern part of the area, reaching 0.5 mm as an average grain size. Quartz and plagioclase have granoblastic shapes with a strong preferred orientation of biotite and muscovite flakes (0.1 to 0.3 mm). Plagioclases are generally untwinned but some twinned grains with albite law twinning may also be present. The composition measured by albite twinning is An_{35}. Potash feldspar is rare. Inclusions of quartz and plagioclase are found in the large crystals of muscovite (Plate 47B). Apatite, sphene, zircon and iron ores are the accessory minerals.

As far as the genesis of porphyroblastic muscovite is concerned, it appears to be formed by some metasomatic process. On textural grounds, post-tectonic crystallization is evident by its cross-cutting relationship to the schistosity. The absence of sillimanite (or fibrolite) is a notable feature among the porphyroblastic muscovite in the present high grade metamorphic rocks. Harte and Johnson (1969) have described the late stage muscovite porphyroblast showing the discordant relationship with the schistosity. They suggest a post-tectonic recrystallization for development of porphyroblastic muscovite with simultaneous growth of sillimanite. Contrary to the above fact, no such process was involved in the present area. Furthermore, the stability of muscovite in the presence of quartz is unlikely in the sillimanite grade rocks. Therefore, the development of late stage muscovite indicates an influx of potash metasomatism at the end of tectonic deformations (D_2), particularly in view of the fact that some potassium rich fluids are likely to be present as residual fluids from the Cooke Hill intrusives which were emplaced during the waning stage of the second deformation.

CHAPTER 5

PETROLOGY AND CHEMISTRY OF THE COOKE HILL MIGMATITES

5.1. INTRODUCTION

A number of terms have been used in the geological literature to describe rocks of migmatitic character. These terms are variously based on the occurrence, structure and the supposed geological history of such rocks. Mehnert (1968) states that a migmatite is a composite rock generally composed of two petrographically different parts: one the unaltered or only slightly modified metamorphic parent rock (paleosome) and a newly formed portion containing predominantly light coloured minerals, i.e. quartz and feldspar (leucosome). Sederholm (1907, p. 48-49) introduced the term "migmatite" merely to describe a process of "remelting" in the metasediments, from the Greek "tekein". However, he used the term to describe rocks which he believed were formed by the injection of magmatic liquid or "ichor" into a metamorphic host. Holmquist (1910, 1920, 1921) rejected this hypothesis and suggested that migmatites were formed in situ by ultra metamorphism. Since that time migmatites have been studied at various places throughout the world and it is accepted by many workers that a number of processes may be involved in the formation of migmatites. They can be formed either by metamorphic differentiation, metasomatism, partial melting and lit par lit injection of magmatic liquid or a combination of two or more of these processes at the same time (cf. Misch, 1968; Starmer, 1969).

Owing to this diversity of opinion, a more or less neutral definition has been proposed by Dietrich and Mehnert (1960). They describe migmatite as a "megascopically composite rock that once consisted of geochemically mobile and immobile (or less mobile) parts, i.e. consists of igneous or igneous looking and/or metamorphic host materials".

It is an established fact that migmatites are intimately associated with high grade metamorphic rocks. This observation had

led many petrologists to believe that there is some genetic relationship between the migmatization and the high temperature and pressure conditions of metamorphism. There are two schools of thought on the formation of migmatites in the high grade gneisses of the amphibolite facies. One believes that the migmatites have resulted by partial melting of country rocks and the other considers that they are formed by metamorphic processes in which the leucocratic veins or bands of igneous appearance were formed by migration of certain elements on a scale of a few inches to a few miles.
In recent years, the geological observations recorded in the field have been tested in the laboratories, and a great deal of experimental work has been done to prove the partial melting hypothesis (von Platen, 1965; Winkler, 1965; Mehnert, 1968 etc.). There seems little doubt that some migmatites have originated by selective melting of the country rocks, especially those described by Mehnert (1953a,b, 1962; Sharma, 1969 etc.), but this hypothesis is not applicable for all migmatites occurring in high grade metamorphic rocks (Barth, 1938; Read, 1927, 1931, 1951, 1957; Harme, 1959, 1966; Kizaki, 1964; King, 1965; Evans, 1966; White, 1966a; Misch, 1968; Kalsbeek, 1970; Leake, 1970 etc.).

Accepting that migmatites can be formed by different processes in different geological conditions, one of the main objects of the present investigation is to decipher the genesis of migmatites in the Cooke Hill area by means of the petrological, structural and chemical studies.

The term migmatite has been used here to describe the rocks of mixed character consisting of leucocratic veins in a metamorphic host. The degree of migmatization in the present area, and in the Mt. Lofty Ranges in general, varies considerably. The rocks have been termed "veined gneisses" (cf. White, 1956, 1966a; Mills, 1964; Fleming, 1971). To avoid the complexity involved in the genetic relationship of migmatitic veins with the country rock, the term "migmatite" has been adopted in preference to "veined gneisses".

The migmatites of the Mt. Lofty Ranges occur in an arcuate belt, 50 miles long, extending continuously from Angaston to the south of Monarto. They are confined to the high grade rocks of the

Amphibolite Facies. A large number of small intrusive bodies of granitic composition are closely associated with the migmatites. The Palmer Granite which is the largest intrusive body in the region is enveloped by migmatites. The Cooke Hill area forms a small part of the migmatitic belt and is predominantly composed of biotite gneisses rich in plagioclase. In places, small intrusive bodies of tonalitic to granodioritic composition are intruded in the migmatitic rocks. The whole area was subjected to a high grade regional metamorphism reaching the sillimanite zone. Tectonic deformation appears to have played an important role in the development of migmatites. Two generations of folds are widely distributed in the migmatites and other metasedimentary rocks.

5.1.1. Megascopic features of the migmatites

The migmatites in the area studied are marked by the presence of leucocratic veins in quartzo-feldspathic gneisses. The layering in the gneisses is defined by alternate light and dark bands, rich in quartzo-feldspathic and biotite components respectively. Generally, veins occur as lenses, sheets, pods and "pinch and swell" like boudins conformable with the foliation of the gneisses. The degree of vein formation systematically increases in the area from north to south, following the increasing grade of metamorphism from the sillimanite-muscovite zone to the sillimanite-potash feldspar zone. There also appears a gradual increase in grain size of the host rock from north to south. In general, the degree of migmatization is also related to intensity of deformation: there is a tendency for the migmatitic veins to be much more common in the stronglyfolded gneisses. The major structures are relatively open in the north but get tighter towards the south in the closure of the main synform.

Although the leucocratic veins occur in variable shapes and sizes, the boundary between the vein and the gneissic host is sharp and concordant (Plate 27A). Rarely, some of the veins show discordant features on a small scale, where the schistosity of the adjacent host rock merges into veins along the strike. Also in a few cases, basic clots of biotite rich rock appear as xenoliths in the veins.

The length of the leucocratic veins rangcs from a few cm to some tens of metres; the thickness may vary from less than 2 cm to as much as 30 cm; it rarely exceeds more than a metre. In some places, a dark rim of biotite selvedges may be found bordering the concordant veins. The thickness of the basic selvedge is generally a few centimetres, or even less in some cases. However, there are some migmatites, where the biotite selvedges do not occur at all. This characteristic feature of migmatite veins rimmed by biotite rich selvedges has also been recorded in other migmatitic rocks of the Mt. Lofty Ranges by White (1956, 1966a), Mills (1964) and Offler (1966).

The leucocratic veins of migmatite are medium to coarse grained rocks (average grain size 1-2 mm). In general, the composition of veins ranges from tonalite to granodiorite in the rocks of granitic appearance. In some cases, thin pegmatitic veins may be found interleaved with the gneisses, but such veins are rather less common than veins of granitic aspect. Late pegmatitic veins of intrusive character usually have a discordant nature and cut across both the leucocratic veins of migmatite and the host rock.

5.1.2. Petrography

The description of the petrography of the migmatites is divided into two parts dealing with two distinct petrological units, namely the leucocratic veins and the gneissic host. The mineralogical assemblages in these two lithological units are similar but the relative proportions of the individual minerals vary considerably. Texturally, the leucocratic veins are two to four times coarser than the host rock. The internal fabric of the two units show complete parallelism as defined by preferred orientation of biotites and to a lesser extent by the dimensional orientation of quartz and/or plagioclase. The degree of preferred orientation of minerals is much greater in the host gneisses than in the veins. The degree of preferred orientation is related to the amount of biotite present in the rock, and since the veins have only a small amount of biotite, the preferred orientation of the minerals is not as apparent. There is some evidence of post-crystalline deformation. Undulose extinction and deformation lamellae are common in quartz grains, both in the veins and in the gneissic host (Plate 28A). Such features are considered to be the evidence of ductile flow of quartz in deformed rocks (Carter et al., 1964; Christie, et al., 1964; Christie & Green, 1964; Christie & Raleigh, 1959 etc.). Wavy extinction, bending and glide twinning and kink structures are often seen in plagioclase of the veins (Plate 28B). There is also some evidence of induced deformational twinning of pericline law superimposed on the albite law twinning of plagioclase as described by Borg and Handin (1966) and Carter and Raleigh (1969) (Plate 29A).

The most common mineral assemblages of the leucocratic veins and the host rocks are as follows:

(1) Quartz-plagioclase-biotite ± muscovite
(2) Quartz-plagioclase-potash feldspar-biotite ± muscovite
(3) Quartz-plagioclase-potash feldspar-biotite ± muscovite ± fibrolite

Apatite, zircon, tourmaline, rutile, epidote, chlorite and iron ores are notable among the accessory minerals.

LEUCOCRATIC VEINS OF THE MIGMATITES

The leucocratic veins of the migmatites have the following variations in the proportions of major minerals: quartz, 25-40%; plagioclase, 45-70%; potash feldspar, 0-20%; biotite, 2-10%; muscovite, 0-10%.

Plagioclase is the most abundant, occurring in various shapes and sizes. It has a subidiomorphic to zenoblastic shape in general, but a few rare crystals show a tabular habit. The grain size generally ranges from less than 1 mm to over 3 mm, some reach 4-5 mm in length. It has smooth curving interlocking boundaries with

TABLE 2

Modal analyses of leucocratic veins and host rock of migmatites

Sample No.	Leucocratic Veins			
	608	479	583	372
Quartz	30.1	31.5	30.1	32.1
Plagioclase	54.0	55.5	51.5	61.4
K-feldspar	5.8	3.5	8.1	0.7
Biotite	9.4	7.2	6.5	4.2
Muscovite	0.4	2.1	3.4	1.3
*Others	0.3	0.2	0.4	0.3
Total	100.0	100.0	100.0	100.0
Total no. of counts	5193	6173	5110	4219

Sample No.	Leucocratic Veins		Host Rock	
	398	474	583	372
Quartz	28.0	26.1	37.2	46.0
Plagioclase	67.9	62.4	31.2	19.9
K-feldspar	0.4	0.0	0.4	0.6
Biotite	1.9	4.3	23.4	32.5
Muscovite	0.8	7.1	7.0	0.1
*Others	1.0	0.1	0.8	0.9
Total	100.0	100.0	100.0	100.0
Total no. of counts	7451	10563	2394	1583

* includes iron ores and accessory minerals

other feldspar and quartz grains, Both twinned and untwinned grains are equally abundant , mostly on albite law, rarely on pericline law. Zoned grains are infrequently present, generally as normally zoned but rarely as weakly oscillatory zoned with rather diffuse boundaries. The compositional difference in the zoned grains is not more than 5% An from core to marginal rim. Post-crystallisation deformational features such as bending and fracturing of twin lamellae or appearance of kink bands are often seen (Plate 29B). A notable feature of most plagioclases is that they are generally clouded with very fine unidentified "dust" and some contain numerous inclusions of sericite and flaky muscovite. Myrmekitic intergrowths of vermicular quartz into plagioclase are occasionally seen. The common range in composition of plagioclase is An_{20-28} but some may be sodic down to An_{15} and others calcic up to An_{34}.

Quartz occurs in variable shapes and sizes. It may be found as droplike inclusions in plagioclase to irregular aggregates between plagioclase and/or potash feldspar crystals. Megacrysts of quartz (3-5 mm) containing small inclusions of biotite and plagioclase are also seen. Undulose extinction is a common feature among all sizes of quartz grains. Replacement phenomena is exhibited by quartz, partially resorbing the plagioclase crystals forming irregular and curved boundaries (Plate 30A). Quartz also replaces potash feldspar in a similar way and less commonly biotite. The quartz is the next most abundant mineral after plagioclase.

Potash feldspar is usually less common. It is readily recognised by its cross-hatched twinning. Some veins are completely devoid of K-feldspar, and in others as much as 20% of the total rock may be K-feldspar. It has an irregular shape and usually occurs as smaller grains filling the interstices between quartz and plagioclase. Occasionally it exhibits a strong tendency to form bigger xenoblastic grains with inclusions of quartz, plagioclase and biotite. It commonly alters to form secondary muscovite (Plate 30B).

Biotite is pleochroic from light brown to dark brown. It shows a weak preferred orientation in the leucocratic veins. The amount ranges from 2-10%.

Muscovite is generally found in amounts from less than 1% up to as much as 10% of the total rock. There are two generations of muscovite. Primary muscovite is less common and occurs as small flakes associated with biotite. The secondary muscovite either develops in the altered cores of plagioclase and microcline, along and across the cleavage planes, or as stumpy porphyroblasts of skeletal shape up to 6 mm in length. These muscovites start growing up along the edges of plagioclase and then finally penetrate irregularly in all directions in the surrounding plagioclase crystals (Plate 31A).

Fibrolitic sillimanite is found as small needles in quartz and muscovite. Iron ores occur as skeletal to disseminated aggregates, distributed unevenly in the quartz-feldspar matrix. Zircon is commonly seen as euhedral to subhedral shaped slender crystals. Apatite is a ubiquitous accessory mineral as prismatic crystals ranging up to 0.5 mm in length. Sphene and tourmaline are rare. As secondary minerals, epidote, white mica and chlorite are also found. Epidote is a common secondary product of plagioclase. White mica is found in the more highly sericitized plagioclase. Chlorite forms along the altered edges of biotite.

HOST ROCK OF THE MIGMATITES

These are well crystallised medium grained biotite gneisses which form an integral part of migmatites as a highly schistose metamorphic host. Their darker colour, finer texture and different mineral proportions make them readily perceivable as a separate rock unit from the light coloured coarse grained leucocratic veins of the migmatites.

The major mineral proportions of host gneisses are as follows: quartz, 25-50%; plagioclase, 15-45%; potash-feldspar, 0-20%; biotite, 10-40%; muscovite, 0-10%. The relative proportions of minerals in the gneisses vary considerably even on hand specimen scale as a result of compositional banding or layering. These rocks have average grain size of about 0.5 mm with excellent mica schistosity (Plate 31B).

Quartz occurs as granoblastic aggregates in variable shapes and sizes. The boundaries between quartz-quartz and quartz-plagioclase grains are commonly curved. Undulose extinction is a common feature. Small inclusions of other minerals are commonly seen in bigger grains of quartz.

Plagioclase is generally clear and less altered than the plagioclase in the leucocratic veins. Untwinned plagioclase grains are more common than twinned grains. Simple albite twinning is predominant but Carlsbad, and Carlsbad/albite twins may be present in some. The bending and dislocation of twin lamellae are frequently seen in larger grains. Usually, plagioclase alters in the core, exhibiting a dusty appearance with the formation of white mica. The compositional range of plagioclase is An_{22-34}.

Potash feldspar is extremely variable in its amount but is usually less common than in the adjacent vein. Mostly it is untwinned and more irregular in shape than plagioclase. Perthitic intergrowth is seen in some grains.

Biotite is the next abundant mineral after quartz and plagioclase. It is found in pleochroic colours of light brown to chocolate brown, occasionally as reddish brown. Pleochroic halos are uncommon. It forms a strong penetrative schistosity most commonly parallel to S_1, but in some rocks two schistosities (S_1 and S_2) are seen, at different angles.

Muscovite as a primary mineral is less common, and is always found as small flakes oriented parallel to the schistosity. Secondary muscovite grows in and around altered plagioclase.

Sillimanite occurs as tiny needles in the quartz grains. Accessory minerals include prismatic apatite, subhedral zircon, disseminated irregularly shaped iron ores, subrhombic to anhedral sphene, columnar aggregates of tourmaline and acicular rutile. Epidote and chlorite are secondary minerals. Small granules of colourless epidote are confined to the altered core of plagioclase, while greenish, weakly pleochroic chlorite is associated with biotite.

PEGMATITE VEINS

Thin veins of pegmatitic aspect are seen infrequently in the migmatite zone. Unlike the intrusive pegmatites which show cross-cutting structures and which fill fractures in the narrow joints of the country rock, these veins show concordance with planar structures of the migmatites. They are generally devoid of mica, a characteristic feature which distinguishes them from the intrusive pegmatites. Therefore, these pegmatitic veins are considered to have originated during regional metamorphism.

Most commonly, they occur as conformable veins or eye shaped boudins in the migmatites. Generally, they are fine to medium grained (0.2-1 mm) and mainly composed of quartz, potash-feldspar and oligoclase. Quartz is xenoblastic with subrounded shape, and undulatory extinction is a common feature in some grains. Usually, quartz does not show any straining effect. Potash feldspar lacks twinning in general, and bigger grains contain inclusions of quartz and plagioclase. Plagioclase shows simple albite twinning, but untwinned plagioclase is also present. Tourmaline, apatite and subhedral shaped zircons are the most common among the accessories.

5.1.3. Age of the Migmatization

Structural studies of migmatites show a genetic connection between migmatization and tectonism. The leucocratic veins of migmatites are commonly concordant with the metamorphic layering and planar schistosity of the enclosing host gneisses. During F_1 deformation, a mobile phase of granitic composition was either accumulated in situ parallel to planar structures of the host rocks, or migrated into zones where the pressure was not as great, in some cases parallel to axial surfaces of the folds. These notable features have been recorded in the field, when a single vein or a group of veins has developed in the gneisses along the axial planes of F_1 folds. During the second deformation, the older structures with concordant veins were refolded with complete harmony in the style of F_2 folding. Often older deformed veins are crossed by another group of leucocratic veins along the hinges of F_2 folds. On the basis of these studies it can be concluded that migmatization commenced during the first phase of deformation and continued into the second phase of deformation. North west of the present area, Offler (1966, p. 24-25) recorded similar observations in the Pewsey Vale area, concluding that the migmatitic veins were formed during two phases of deformation.

5.2. PETROCHEMISTRY OF THE COOKE HILL MIGMATITES

Introduction

One of the main objectives of the present investigation was to test the validity of two different hypotheses proposed for the genesis of migmatites in the high grade rocks of the Kanmantoo Group metasediments. White (1956, 1966a) and Offler (1966) favoured metamorphic differentiation for development of migmatites in the Palmer and the Pewsey Vale areas, whilst Mills (1964) favoured selective melting of the country rock for the genesis of migmatites in the Cambrai area, just north of the present area. Mills stated that partial melting may be responsible for formation of migmatites due to increase in temperature by syn-tectonic intrusion of small bodies of granodiorite and the abundance of water to promote the selective melting of the low temperature quartz-feldspar fraction of the quartzo-feldspathic rocks.

By means of a petrochemical study it was hoped to reveal whether material has been introduced metasomatically by aqueous fluids released from the igneous intrusives or whether partial melting of the country rock or metamorphic differentiation has contributed in the development of migmatites in this area. The major and trace element chemistry of the migmatic veins,* a host rock and a few metasedimentary rocks are discussed in this chapter.

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* see Appendix 3 for selection of samples

5.2.1. Major Element Chemistry

The most striking feature of the migmatitic veins is the high Na and low K contents of these rocks. The average Na/K ratio of the veins is 6.88, as compared to the average Na/K ratio of 1.64 for the host rock, 3.24 for the Cooke Hill tonalites, 1.20 for the Cooke Hill granodiorite and 4.06 for the Massive granodiorite. It is interesting to note that the major oxides SiO_2 and Al_2O_3, are almost constant in all the rocks examined (Table 3A, B), but there is some variation in total Fe_2O_3, K_2O, Na_2O, CaO and MgO of the veins due to different amounts of quartz, feldspar and mica present in the analysed samples.

Comparison of major element co-variation in migmatites and other associated rocks

One of the major steps in the understanding of the process of migmatization was to determine the nature of the original rock before migmatization. It is very difficult to determine the pre-migmatite composition directly, so it is hoped that comparison and correlation of chemical analyses of different coexisting rocks may indicate some relationship of the migmatites to either metasediments or igneous intrusives.

Values of CaO, Na_2O and K_2O, plotted on the triangular diagram in Figure 18, show the high soda content of the migmatite veins, a feature which distinguishes them from the Cooke Hill intrusives. The Cooke Hill intrusives generally fall near the soda-rich limits of the fields delineated by Nockolds (1954) for the common igneous rocks. The metasediments are more variable, some plot close to the migmatite veins, and some close to theoretical granodiorite.

The ionic percentages of Fe^3 and Fe^2 are plotted in Figure 19 and show that although the Fe^2/Fe^3 ratios are similar for all the rock types, the veins have lower absolute values than all of the igneous rocks, host rock of migmatite (372AG) and the metasediments. The MgO and FeO (total iron) (Fig. 20) also show a

TABLE 3A

Chemical analyses and Barth mesonorms of migmatites

Sample No.	372	377	398	515	514	479	483
SiO_2	73.06	74.44	71.63	74.66	70.99	74.07	74.43
TiO_2	0.15	0.11	0.21	0.15	0.31	0.11	0.20
Al_2O_3	15.56	15.50	15.75	13.99	16.56	14.68	14.31
Fe_2O_3	0.31	0.15	0.96	0.29	0.68	0.24	0.40
FeO	0.76	0.37	1.00	0.74	1.36	0.58	0.70
MnO	0.12	0.07	0.07	0.07	0.07	0.06	0.07
MgO	0.32	0.23	0.41	0.37	0.36	0.28	0.38
CaO	1.56	0.87	1.75	1.87	2.27	1.57	1.59
Na_2O	7.33	7.21	6.66	5.34	5.78	6.75	5.97
K_2O	0.77	0.73	0.75	1.30	1.20	0.65	1.44
P_2O_5	0.19	0.14	0.12	0.11	0.13	0.21	0.08
L.o. Ignit.	0.50	0.69	0.31	0.54	0.38	0.52	0.37
Total	100.63	100.51	99.62	99.43	100.09	99.72	99.94
Q	22.38	25.50	24.56	32.64	26.04	27.22	28.64
Or	2.31	0.00	0.11	4.49	0.07	1.32	6.54
Ab	64.60	63.95	59.53	48.45	51.74	60.38	53.51
An	6.48	3.49	8.04	8.78	10.56	6.49	7.43
Sill	0.00	0.43	0.00	0.00	0.00	0.00	0.00
Bi	2.39	1.40	2.43	2.52	3.06	1.92	2.24
Ms	0.92	4.82	3.89	2.38	7.11	1.84	0.77
Mt	0.32	0.15	1.00	0.31	0.71	0.25	0.42
Il	0.21	0.15	0.29	0.21	0.43	0.15	0.28
Ap	0.39	0.29	0.25	0.23	0.27	0.44	0.17

Migmatite veins (372, 377, 398, 515, 514, 479, 483)

TABLE 3B

Chemical analyses and Barth mesonorms of migmatites

Sample No.	497	608	705	474	Mean*	372AG
SiO_2	73.98	72.74	74.34	72.06	73.30	74.86
TiO_2	0.15	0.17	0.15	0.18	0.17	0.49
Al_2O_3	14.18	15.58	14.63	16.15	15.17	11.63
Fe_2O_3	0.40	0.23	0.24	0.54	0.40	0.84
FeO	0.73	0.94	0.43	0.84	0.76	q 3.74
MnO	0.06	0.07	0.07	0.06	0.07	0.10
MgO	0.42	0.48	0.27	0.25	0.34	1.15
CaO	2.06	1.55	1.18	1.33	1.60	0.92
Na_2O	5.67	7.14	5.55	6.25	6.33	3.82
K_2O	0.66	1.06	3.12	1.13	1.16	2.06
P_2O_5	0.06	0.16	0.17	0.16	0.13	0.08
L.o. Ignit.	0.78	0.49	0.50	0.78	0.53	0.46
Total	99.15	100.61	100.65	99.73	99.96	100.15
Q	32.06	22.06	25.66	26.52		40.54
Or	1.04	3.53	16.57	0.00		1.69
Ab	51.60	62.95	49.41	56.17		35.08
An	10.13	6.60	4.97	5.69		4.63
Sill	0.00	0.00	0.00	0.88		0.00
Bi	2.56	3.27	1.51	1.84		10.72
Ms	1.84	0.81	1.07	7.74		5.69
Mt	0.42	0.24	0.25	0.57		0.90
Il	0.21	0.23	0.21	0.25		0.70
Ap	0.13	0.33	0.35	0.33		0.06

Migmatite veins (497, 608, 705, 474)

Migmatite host rock (372AG)

* Mean of migmatite veins

similar variation between migmatite veins and the igneous rocks, but there is wide scattering of the values in the metasediments. The ionic percentages of Ti and Fe (total), plotted in Figure 21, show a similar trend between all types of rocks, but the migmatite veins have a lower absolute value of Fe/Ti ratios.

Thus the migmatite veins differ from the metasediments in that they have higher absolute values of Na_2O and lower amounts of K_2O and consequently high Na/K ratios. In addition, the veins have a higher CaO content. This suggests that if the veins are derived from the metasediments, some process has operated to increase the Na_2O and CaO content of the veins. Inasmuch as the Na/Al ratio has remained about constant, this implies that the "veins" have been enriched in the component of plagioclase feldspars and depleted in biotite content.

The lower absolute values of Fe(total), Mg and Ti in the migmatite veins are merely related to lower amounts of biotite and iron oxides in the vein than in the host rocks.

5.2.3. Trace Element Chemistry

Data for trace elements Rb, Sr and Ba have been collected to test the hypotheses that the migmatite veins were produced by injection or permeation of magmatic fluid from the Cooke Hill intrusives. The geochemical behaviour of Rb, Sr and Ba in igneous rocks has been well established in recent times (Taylor, 1965; Heier & Taylor, 1959; Taylor & Heier, 1960). If the veins have been evolved from magmatic source, these critical trace elements will be of great help in elucidating their genesis.

The trace element determinations were made both for the total rocks and their respective biotites in some 23 samples, using X-ray fluorescence methods. The precision (reproducibility) of analyses is reasonably good (see Appendix 3).

The concentration of trace elements in leucocratic veins and host rock of migmatite, the Cooke Hill intrusives and metasedimentary rocks is presented in Table 4. There is some

TABLE 3C

Chemical analyses and Barth mesonorms of the Cooke Hill intrusives

Sample No.	75	140	199	233	295	511	137	232	208
SiO_2	69.62	66.65	68.71	68.71	69.67	68.83	69.82	71.31	68.59
TiO_2	0.34	0.43	0.25	0.37	0.36	0.37	0.26	0.26	0.40
Al_2O_3	16.45	17.15	17.23	16.19	15.81	16.29	16.08	16.01	15.49
Fe_2O_3	0.52	0.59	0.44	0.35	0.38	0.54	0.44	0.52	0.98
FeO	1.73	2.02	1.72	1.89	1.53	1.93	1.34	1.20	2.11
MnO	0.09	0.09	0.04	0.14	0.10	0.10	0.06	0.08	0.08
MgO	1.12	1.26	0.96	0.96	1.23	0.97	0.78	0.68	1.06
CaO	3.01	3.02	2.94	3.05	2.75	3.12	1.97	1.81	3.66
Na_2O	5.52	5.93	5.47	5.52	5.32	4.69	5.22	4.11	4.93
K_2O	1.20	1.57	1.71	1.40	1.73	2.10	3.49	3.44	1.08
P_2O_5	0.18	0.13	0.10	0.18	0.08	0.20	0.11	0.15	0.16
L.o. Ignit.	0.42	0.63	0.77	0.75	0.57	0.40	0.51	0.47	0.76
Total	100.20	99.47	100.34	99.51	99.53	99.54	100.08	100.04	99.30
Q	24.68	18.38	22.24	23.25	24.49	25.43	20.21	27.09	26.68
Or	0.62	3.42	3.13	3.31	5.11	5.62	16.97	12.26	2.05
Ab	49.30	53.14	48.87	49.79	47.93	42.41	46.63	37.15	44.95
An	14.11	14.45	14.31	14.55	13.37	14.59	9.58	9.61	17.21
Ms	3.31	1.49	4.22	1.10	1.16	3.82	0.72	7.79	0.00
Bi	6.73	7.64	6.27	6.75	6.90	6.62	4.85	4.21	7.08
Mt.	0.54	0.62	0.46	0.37	0.40	0.57	0.46	0.55	1.04
Il	0.47	0.60	0.35	0.52	0.50	0.52	0.36	0.36	0.41
Ap	0.37	0.27	0.17	0.38	0.15	0.42	0.23	0.32	0.34
Sp	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.23

Cooke Hill tonalites	(75, 140, 199, 233, 295, 511)
Cooke Hill granodiorites	(137, 232)
Massive granodiorite	(208)

TABLE 3D

Chemical analyses and Barth mesonorms of metasedimentary rocks

Sample No.	138	139	205	201*	184	149	144
SiO_2	73.57	78.47	65.12	nd	76.71	75.82	77.95
TiO_2	0.43	0.37	0.79	nd	0.29	0.40	0.25
Al_2O_3	13.34	10.88	15.34	nd	12.26	12.25	12.51
Fe_2O_3	0.85	0.31	1.69	nd	0.90	0.55	0.42
FeO	1.92	1.83	5.18	nd	1.05	1.64	1.01
MnO	0.07	0.08	0.13	nd	0.06	0.12	0.07
MgO	1.20	1.15	3.05	0.59	0.17	1.11	0.56
CaO	0.62	0.93	0.47	nd	0.95	1.10	0.77
Na_2O	4.21	4.41	3.40	3.79	3.68	4.40	5.02
K_2O	4.02	1.18	4.19	1.65	4.25	2.49	0.63
P_2O_5	0.17	0.09	0.18	nd	0.05	0.19	0.02
L.o. Ignit.	0.34	0.31	1.02	nd	0.32	0.52	0.35
Total	100.74	100.01	100.56	-	100.69	100.59	99.56
Q	30.37	43.79	26.96	-	34.72	36.13	41.18
Or	15.97	0.18	0.00	-	24.34	8.94	0.00
Ab	37.84	40.23	30.90	-	33.38	39.79	45.78
An	2.18	4.17	1.28	-	4.43	4.56	3.75
Sill	0.00	0.00	0.11	-	0.00	0.00	2.73
Bi	6.95	7.32	19.40	-	1.64	6.52	3.49
Ms	4.85	3.26	18.10	-	0.00	2.52	2.73
Mt	0.89	0.33	1.79	-	0.95	0.58	0.45
Il	0.60	0.52	1.11	-	0.35	0.56	0.35
Ap	0.36	0.19	0.38	-	0.11	0.40	0.04
Sp	0.00	0.00	0.00	-	0.08	0.00	0.00

* nd - not determined

TABLE 4

Major and trace element data for migmatites and other associated intrusives and metasedimentary rocks

Specimen No.	Na%	K%	Ca%	Rb ppm	Rb ppm	Ba ppm	K/Rb	Sr/ Ca.10^3	Ba/Sr	Ba/R	Ba/ K.10^3	Na/K
372	5.44	0.64	1.11	17	251	340	376	23	1.35	20	53	8.50
398	4.94	0.62	1.25	13	445	336	477	36	0.75	25	54	7.97
474	4.64	0.94	0.95	28	336	230	336	35	0.68	8	24	4.93
377	5.35	0.61	0.62	13	229	138	469	37	0.60	11	23	8.77
497	4.21	0.55	1.47	20	381	224	275	26	0.59	11	41	7.65
514	4.29	1.00	1.62	30	361	373	333	22	1.04	12	37	4.29
515	3.96	1.08	1.34	24	355	887	450	26	2.50	37	82	3.66
479	4.01	0.54	1.12	10	295	108	540	26	0.37	11	20	9.27
Mean	4.85	0.74	1.18	19	331	329	407	29	0.98	17	42	6.88
372AG	2.83	1.71	0.66	67	118	572	255	18	4.84	8	33	1.64
295	3.95	1.44	1.96	46	551	785	313	28	1.42	17	54	2.74
233	4.09	1.16	2.18	49	653	1759	237	30	2.69	36	152	3.52
199	4.06	1.42	2.10	52	542	588	275	26	1.08	11	41	2.85
75	4.09	1.00	2.15	35	664	1198	286	31	1.80	34	119	4.09
140	4.40	1.00	2.12	35	699	675	363	33	0.96	19	53	3.56
137	3.87	2.90	1.41	93	665	1906	312	47	2.87	20	66	1.33
232	3.05	2.85	1.29	100	647	1667	285	50	2.58	17	58	1.07
208	3.66	0.90	2.61	41	342	565	219	13	1.65	14	63	4.06
Mean	3.89	1.61	1.97	56	595	1142	286	32	1.88	18	76	2.89
139	3.27	0.98	0.66	36	139	238	272	21	1.71	7	24	3.33
138	3.12	3.34	0.44	113	108	894	295	24	8.28	8	27	0.93
205	2.52	3.48	0.30	168	45	524	207	14	11.64	3	15	0.72
201	2.81	1.37	n.d.	32	203	270	428	-	1.33	8	20	2.05
184	2.73	3.53	0.68	134	69	535	263	10	7.75	4	15	0.77
149	3.26	2.07	0.79	76	312	1202	272	39	3.85	16	58	1.57
Mean	2.95	2.46	0.57	93	146	610	289	21	5.76	8	26	1.56

Migmatite veins	(372, 398, 474, 377, 497, 514, 515, 479)
Migmatite host rock	(372AG)
Cooke Hill tonalites	(295, 233, 199, 75, 140)
Cooke Hill granodiorites	(137, 232)
Massive granodiorite	(208)
Metasedimentary rocks	(139, 138, 205, 210, 184, 149)

variation within the same rock type, but the average values for each rock type show significant differences. These results will be discussed in detail in the following paragraphs and Table 4 will provide the data referred to in the text.

Strontium

Considerable differences in distribution of Sr in the migmatitic veins, granitic intrusives and metasedimentary rocks are noted. The ranges for this element are 251 to 445 ppm (average 331 ppm) in migmatitic veins, whilst a host rock (A285/372AG) has 118 ppm and the metasedimentary rocks have 45 to 312 ppm (average 146 ppm). The Sr content ranges from Cooke Hill intrusives (595 ppm) to migmatitic veins (331 ppm) to host rock (118 ppm) and metasediments (146 ppm). Sr/Ca ratios are generally variable but the mean values are fairly distinct with respect to different rock types (Fig. 22). The average Sr/Ca.10^3 ratios range from 32 (Cooke Hill intrusives) to 29 (migmatitic veins) to 18 (host rock) and 21 (metasediments).

Rubidium

A notable feature of the Rb values is the low amount of Rb in the migmatite veins. The mean value of 19 ppm of the veins contrasts markedly with the mean value of 93 ppm in the metasediments, 43 ppm in the Cooke Hill tonalites, 93 ppm in the Cooke Hill granodiorites and 41 ppm in the Massive granodiorite. K/Rb ratios show that there is a general correlation between K and Rb values in the rocks. The migmatite veins are lower in both potash and rubidium than other rocks mainly because of their lower content of biotite and potash feldspar. The relatively higher K/Rb ratios for the veins (mean value 407) are quite distinct from those of the intrusives (mean 295 for Cooke Hill tonalites, 299 for Cooke Hill granodiorites and 219 for Massive granodiorite) and the metasediments (mean 289). This is illustrated in Figure 23.

Barium

The absolute value of Ba is lowest in the migmatitic veins and highest in the Cooke Hill intrusives. The concentration of Ba is extremely variable in the metasediments with some values lower than in the migmatites and the gneissic host of migmatite (Sample No. 372AG). In general, the Ba values of the host rock of migmatite and mean value of metasediments are about twice that of migmatitic veins. Ba is nearly identical in size to K and substitutes for K alone among the common cations. According to data of Nockolds and Allen (1953), Ba enters K-feldspar more readily than it does into biotite. Thus the migmatite veins which contain a lesser amount of K-feldspar and biotite have a lower Ba content than the Cooke Hill intrusives which are richer in biotite and/or K-feldspar.

Ba in the Cooke Hill tonalites ranges from 588 to 1759 ppm (mean 1001), in the Cooke Hill granodiorite from 1677 to 1906 ppm (mean 1786), in the Massive granodiorite 565 ppm, in migmatitic veins from 108 to 887 ppm (mean 329), in migmatite host rock 572 ppm and in the metasediments from 238 to 1202 ppm (mean 610). Ba/Rb, Ba/K and Ba/Sr ratios of the veins are markedly different from the ratios in the intrusive igneous rocks. Generally, these ratios are lower in extremely fractionated magmatic rocks than in less fractionated rocks (Taylor & Heier, 1960; Taylor, 1965 etc.). The Ba/Rb ratio, which is a critical index of magmatic differentiation, should be significantly lower in the veins than in the igneous rocks if they have been derived from a residual liquid of the Cooke Hill magma. The veins have mean Ba/Rb ratios of 17, the Cooke Hill tonalites 23, the Cooke Hill granodiorite 18, and the Massive granodiorite 14, but the absolute values of Ba and Rb are significantly lower in the veins than in the intrusive rocks. These are illustrated in Figure 24. The Ba/K ratios are generally lower in the veins than in all igneous intrusives (Figure 25), and the absolute values are also lower. The Ba/Sr ratio is lower in the veins than in the intrusive rocks and also lower than in the metasediments.

5.2.4. Chemistry of the Biotites

Mineralogically, the major constituent minerals of leucocratic veins of migmatite and adjacent host rocks are similar, but the relative proportions of quartz, feldspar and biotite are significantly different (Table 2; Figure 16). The main potassium bearing mineral is biotite, since the amount of K-feldspar is low both in most of the veins and in the host rock of the migmatites (see section 5.1.2. and Fig. 16). In view of the sensitivity of biotite composition to certain conditions of formation (temperature, water pressure, oxygen fugacity etc.), it was thought that major and trace element chemistry of biotites from the migmatitic veins and host rock might throw light on the genesis of migmatites.

For comparison, biotites from selected samples of migmatites and from associated country rock and igneous intrusives were analysed. The analyses of biotites with their structural formula is given in Table 5A, B, C and trace element data for Rb, Sr and Ba and their respective ratios with other elements is presented in Table 7.

The Fe/Fe+Mg ratio of biotite is indicative of its petrogenic history. Recent experimental work on biotite by Eugster and Wones (1962) and Wones and Eugster (1965) demonstrates that the relative proportion of Fe and Mg in biotite is dependent upon P_{H_2O}-f_{O_2}-T conditions. Therefore, the Fe/Fe+Mg ratio of biotite in rocks of the present area may give a picture of the condition of formation. Chemically the biotites from migmatitic veins do not differ much from biotites of the Cooke Hill intrusives and metasediments. Only two biotites from the metasedimentary rocks (194 and 144) show significantly different Fe and Mg contents. Both of these specimens are atypical. Sample 144 is described as granite gneiss in section 4.7. and sample 184 was collected within a few metres of the diopside gneisses (see chapter 6). The average values of $FeO + Fe_2O_3 \times 100/FeO + Fe_2O_3 + MgO$ ratio of biotites (Table 6) from the migmatitic veins is 63%, from the host rock of migmatite 69%, igneous intrusives 66% and metasedimentary rocks (omitting Sample Nos. 184 and 144) 65%. All of the values are very close, indicating only slight variation in P-T conditions at the time of crystallization of the biotites.

The triangular plot of biotite compositions, on a diagram representing octahedral groups of (Al + Fe^{+3} + Ti), Fe^{+2} + Mn) and Mg values, in Figure 26, indicates that all the biotites have crystallized in a narrow range of P_{H_2O}-P_{O_2}-T conditions, resulting in homogeneity of composition. All biotites fall in a fairly restricted field within the field of biotites from schists and gneisses as delineated by Foster (1960). This suggests that all the biotites in the different rocks recrystallized in very similar P - T conditions of metamorphism. The amounts of trace elements Sr, Rb and Ba in the biotites of the migmatite veins, metasediments and igneous rocks are listed on Table 7. Sr is present in such small amount that it will not be further considered. The concentrations of Ba and Rb are variable amongst biotites of one rock type, but some weak trends may be seen. The biotites of the metasediments appear to be poorer in Ba and richer in Rb than the biotites of the migmatite veins or the igneous rocks. The biotites of the migmatite veins have similar Ba and Rb to the biotites of the igneous rocks and the Ba/Rb ratio of the migmatite biotites covers the range of the Ba/Rb ratio of the igneous biotites. However the absolute values of both Ba and Rb tend to be lower in the biotites of the migmatite veins. No clear picture emerges from the data beyond the fact that the biotites of the metasediments are relatively low in Ba compared with both the veins and intrusive rocks. However, when the compositions of the rocks are considered all except the veins have similar Ba and Rb content. This suggests that the trace element geochemistry of the biotite only reflects the ability of the biotites to take up Ba and Rb into the crystal lattice.

5.3. ORIGIN OF THE MIGMATITES

The history of migmatization appears to be closely linked with tectonism accompanied by high grade regional metamorphism but there are other factors which may have contributed to the evolution of migmatites. From structural considerations the migmatites appear to have developed Syn-kinematic to Late-kinematic during two successive

TABLE 5A

Chemical analyses of biotites

Sample No.	372	377	398	497	483	514	515	474	372AG
SiO_2	35.19	37.13	35.54	36.32	35.40	35.44	37.01	36.74	34.97
TiO_2	2.40	2.15	3.43	2.64	2.91	2.74	2.57	2.64	2.43
Al_2O_3	16.92	19.46	16.25	17.32	18.51	17.30	17.31	17.55	17.44
Fe_2O_3	2.74	4.89	4.69	2.94	3.71	3.86	3.17	5.11	3.24
FeO	18.16	12.65	16.35	14.65	15.05	14.69	14.90	13.93	18.30
MnO	0.24	0.23	0.22	0.13	0.18	0.14	0.18	0.11	0.22
MgO	10. 1	11.85	10.99	11.18	11.60	11.07	10.58	11.53	9.55
CaO	0.03	0.15	0.13	0.05	0.05	0.02	0.09	0.05	0.05
Na_2O	0.20	0.33	0.18	0.18	0.25	0.15	0.21	0.24	0.15
K_2O	8.70	7.52	8.79	8.81	9.48	9.01	8.81	9.58	9.55
L.o. Ignit.	4.97	3.48	3.36	5.11	2.47	4.85	5.66	3.12	3.18
Total	100.06	99.84	99.93	99.33	99.61	99.27	100.49	100.60	99.09
Structural formula on anhydrous basis of O = 22									
Si	5.38	5.40	5.34	5.49	5.24	5.38	5.56	5.40	5.34
Al	2.62	2.60	2.66	2.51	2.76	2.62	2.44	2.60	2.66
Z	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00
Al	0.43	0.74	0.21	0.58	0.48	0.48	0.62	0.44	0.48
Fe^{+3}	0.31	0.53	0.53	0.33	0.41	0.44	0.36	0.56	0.37
Ti	0.28	0.23	0.39	0.30	0.32	0.31	0.29	0.29	0.28
Mg	2.39	2.57	2.46	2.52	2.56	2.51	2.37	2.53	2.17
Mn	0.03	0.03	0.03	0.02	0.02	0.02	0.02	0.01	0.03
Fe^{+2}	2.32	1.54	2.05	1.85	1.86	1.87	0.87	1.71	1.34
Y	5.76	5.64	5.67	5.60	5.65	5.63	5.72	5.54	5.67
Ca	0.00	0.02	0.02	0.01	0.01	0.00	0.01	0.01	0.01
Na	0.06	0.09	0.05	0.05	0.07	0.04	0.06	0.07	0.04
K	1.70	1.40	1.68	1.70	1.79	1.75	1.69	1.80	1.86
X*	1.78	1.52	1.79	1.78	1.88	1.81	1.78	1.88	1.91

* includes Ba and Rb

Migmatite veins	(372, 377, 398, 497, 483, 514, 515, 474)
Migmatite host rock	(372AG)

TABLE 5B

Chemical analyses of biotites

Sample No.		75	140	199	233	295	511	232	137	208
SiO_2		34.00	35.29	35.10	34.34	35.77	35.86	34.54	36.83	24.90
TiO_2		3.06	3.18	2.80	3.09	2.90	3.01	3.02	3.55	2.81
Al_2O_3		17.06	16.60	17.23	16.34	17.20	15.39	16.59	14.99	16.36
Fe_2O_3		4.60	5.18	3.59	2.51	4.68	3.08	3.73	5.25	2.61
FeO		17.51	14.65	16.73	18.55	15.27	16.84	16.53	15.94	18.02
MnO		0.32	0.23	0.32	0.18	0.28	0.22	0.20	0.36	0.23
MgO		9.94	11.04	9.62	10.27	11.89	12.00	10.36	10.69	10.83
CaO		0.13	0.11	0.11	0.01	0.07	0.21	0.01	0.09	0.42
Na_2O		0.17	0.15	0.16	0.12	0.19	0.12	0.10	0.14	0.19
K_2O		8.91	9.64	9.15	8.88	9.62	9.04	9.06	9.43	8.18
L.O. Ignit.		4.27	4.15	5.38	5.95	2.71	3.87	5.93	3.05	5.45
Total		99.97	100.22	100.19	100.24	100.58	99.61	100.07	100.32	100.10
Structural formula on anhydrous basis of 0 = 22										
Si		5.20	5.32	5.37	5.32	5.29	5.42	5.33	5.50	5.35
Al		2.80	2.68	2.63	2.68	2.71	2.58	2.67	2.50	2.65
	Z	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00	8.00
Al		0.27	0.26	0.48	0.31	0.28	0.17	0.3e5	0.14	0.31
Fe^{+3}		0.53	0.59	0.41	0.29	0.52	0.35	0.43	0.59	0.30
Ti		0.35	0.36	0.32	0.36	0.32	0.34	0.35	0.40	0.32
Mg		2.26	2.48	2.19	2.37	2.62	2.71	2.38	2.38	2.50
Fe^{+2}		2.24	1.84	2.14	2.40	1.89	2.13	2.13	1.99	2.31
Mn		0.04	0.03	0.04	0.02	0.03	0.03	0.03	0.05	0.03
	Y	5.69	5.57	5.58	5.75	5.66	4.73	5.67	5.55	5.77
Ca		0.02	0.02	0.02	0.00	0.02	0.03	0.00	0.01	0.07
Na		0.05	0.04	0.05	0.04	0.05	0.03	0.03	0.04	0.06
K		1.74	1.85	1.79	1.76	1.81	1.75	1.78	1.80	1.60
	X*	1.85	1.94	1.87	1.82	1.89	1.84	1.83	1.86	1.75

* X includes Ba and Rb

Cooke Hill tonalites	(75, 140, 199, 233, 295, 511)
Cooke Hill granodiorites	(232, 137)
Massive granodiorite	(208)

TABLE 5C

Chemical analyses of biotites

Sample No.		138	139	205	201	184	149	144	3
SiO_2		36.18	35.69	33.91	35.64	36.60	34.76	35.62	36.70
TiO_2		3.53	3.14	2.23	2.74	2.66	3.32	3.73	4.41
Al_2O_3		16.87	18.02	18.64	18.48	13.73	13.89	14.62	14.50
Fe_2O_3		4.22	2.62	2.40	4.85	4.98	3.54	5.04	3.46
FeO		15.31	16.78	17.01	15.87	20.18	15.48	13.33	17.99
Mno		0.22	0.19	0.13	0.20	0.51	0.14	0.18	0.19
MgO		10.98	9.92	10.25	9.50	7.80	12.80	13.76	10.57
CaO		0.02	0.07	0.02	0.09	0.19	0.07	0.36	0.16
Na_2O		0.13	0.16	0.23	0.13	0.19	0.12	0.17	0.36
K_2O		9.80	9.77	9.41	9.28	9.02	9.02	8.98	8.60
L.o. Ignit.		2.98	3.43	4.90	3.62	3.99	5.98	3.43	3.13
Total		100.06	99.79	99.13	100.40	99.85	99.12	99.22	100.02
Structural formula on anhydrous basis of O = 22									
Si		5.38	5.36	5.23	5.32	5.66	5.41	5.35	5.51
Al		2.62	2.64	2.77	2.68	2.34	2.55	2.59	2.49
	Z	8.00	8.00	8.00	8.00	8.00	7.96	7.94	8.00
Al		0.33	0.55	0.61	0.57	0.16	0.00	0.00	0.08
Fe^{+3}		0.47	0.30	0.28	0.55	0.58	0.41	0.57	0.39
Ti		0.39	0.35	0.26	0.31	0.31	0.39	0.42	0.49
Mg		2.43	2.22	2.35	2.12	1.80	2.97	3.08	2.37
Mn		0.03	0.02	0.02	0.02	0.07	0.02	0.02	0.02
Fe^{+2}		1.90	2.11	2.19	1.98	2.61	2.02	1.67	2.26
	Y	5.55	5.55	5.71	5.55	5.53	5.81	5.76	5.61
Ca		0.00	0.01	0.00	0.01	0.03	0.01	0.06	0.02
Na		0.04	0.05	0.07	0.04	0.06	0.04	0.05	0.10
K		1.86	1.87	1.85	1.77	1.78	1.79	1.72	1.65
	X*	1.91	1.95	1.93	1.83	1.88	1.86	1.83	1.79

* X includes Ba and Rb
Metasedimentary rocks (138, 139, 205, 201, 184, 149, 144, 3)

TABLE 6

BIOTITE COMPOSITIONS

Weight Percentages of Fe_2O_3, FeO, MgO and the factor "F"

where $$F + \frac{Fe_2O_3 + FeO}{Fe_2O_3 + FeO + MgO} \times 100$$

Sample No.	Fe_2O_3%	FeO%	MgO%	"F"
		Migmatite Veins		
372	2.74	18.16	10.51	66
377	4.89	12.65	11.85	60
398	4.69	16.35	10.99	66
497	..94	14.65	11.18	61
483	3.71	15.05	11.60	62
514	3.86	14.69	11.07	63
515	3.17	14.90	10.58	63
474	5.11	13.93	11.53	62
Mean	-	-	-	63
		Migmatite Host Rock		
372AG	3.24	18.31	9.55	69
		Cooke Hill Intrusives *		
75	4.60	17.51	9.94	68
199	3.59	16.73	9.62	68
233	2.51	18.55	10.27	67
140	5.18	14.65	11.04	66
295	4.68	15.27	11.89	64
511	3.08	16.84	12.00	62
137	5.25	15.94	10.69	63
232	3.73	16.53	10.36	66
208	2.61	18.02	10.93	65
Mean	-	-	-	65
		Metasedimentary Rocks		
149	3.54	15.48	12.80	60
138	4.22	15.31	10.98	64
139	2.62	16.78	9.92	66
205	2.40	17.01	10.25	65
201	4.85	15.87	9.50	69
184	4.98	20.18	7.80	76
144	5.04	13.33	13.76	57
3	3.46	17.99	10.57	67
Mean	-	-	-	66

* Cooke Hill tonalites (75, 199, 233, 140, 295, 511)
Cooke Hill granodiorites (137, 232) Massive granodiorite (208)

phases of deformation. The late kinematic migmatites of second deformation are likely to be contemporaneous with the Cooke Hill granitic intrusives phase which were emplaced before the completion of F_2 folding. Seemingly, this last phase of migmatization was operating during the intrusion of numerous igneous bodies and hence the residual liquids from the crystallising magmas could possibly have been injected to form late kinematic migmatites. All these facts indicate that migmatites have a complex geological history .

Hypotheses generally proposed in the literature for the origin of migmatites are given below. These are examined to determine whether they are applicable to the migmatites in the present area. For comparative study, the mineralogical composition of the leucocratic veins of migmatites and the host rock is expressed in terms of quartz (Q), plagioclase (Ab + An) and K-feldspar (Or) calculated from the analyses (Fig. 27).

Hypotheses that have been invoked to explain the formation of migmatites include:

1. "Lit par lit" injection of magma to give rise to the leucocratic veins in the migmatized gneisses.
2. Selective melting of country rock "anatexis" and subsequent consolidation of melted material either in situ, or after local migration of the resulting melts into zones of real or potential low energy developed in the rocks by tectonism.
3. Metamorphic differentiation involving the differential movement of materials in a vapour phase, and resulting in concentration of quartz-feldspar in the veins and enrichment of biotite in host gneisses of migmatites.
4. Metasomatism. Formation of leucocratic veins in migmatites by metasomatic process involving large scale migration of alkali-rich fluids.

5.3.1. Lit par lit injection of magma

The idea of injection of granitic magma for the formation of migmatites is excluded by the following facts:

(a) If the leucocratic veins of migmatites have been formed by the injection of residual liquids from the Cooke Hill intrusives, then they must be enriched in K and Rb. The chemistry of veins from present migmatites shows strikingly low values of K and Rb as compared to parent intrusive "granites"

It is generally observed that the K/Rb ratio systematically decreases during a continuous fractional crystallization of magma, hence this ratio would be expected to be lower in the late stage residual granite than in the parent Cooke Hill tonalitic and granodioritic magmas. Contrary to this, the average K/Rb ratio of migmatite veins is 407 against the average ratios of tonalites as 295 and granodiorite as 299 (see Table 4 and Fig. 23).

(b) If it is considered that the injected material of veins was derived as "emanation" from some underlying concealed granitic intrusives other than the Cooke Hill intrusives, then the hypothetical magmas should be highly fractionated rocks and a K/Rb ratio of less than 100 would be expected for such granitic rocks (Taylor, 1965). However, the K/Rb ratios of analysed samples of migmatite veins range from 275 to 540.

(c) The trace element data of biotites, both from migmatite veins and the Cooke Hill intrusives provide further evidence against the injection of granitic liquids in the development of leucocratic veins of migmatites. It has been shown by Lange, et al. (1966) that the average K/Rb ratio of magmatic biotites decreases from high temperature granodiorite to intermediate temperature quartz-monzonite to low temperature granite in a systematic order. The observed trend for average K/Rb ratios of biotites in the Cooke Hill tonalite is 200 and in the granodiorite is 125, which generally follow their predictions but a rather reverse trend has been found in the case of biotites from the migmatite veins (Table 7). The K/Rb ratios of biotites in strongly fractionated granites have a range of 35-100 (Kolbe, 1965; Lange, et al. 1966; Hurley, 1958 and Fairbairn, et al. 1960), whereas biotites from the migmatite veins have K/Rb ratios in the range of 135-300, much higher than is expected for a fractionated magmatic granite.

In the light of above facts it is concluded that the leucocratic veins of the migmatites were not formed by bodily injection of granitic liquids, The field evidence also shows no obvious connection between the granitic intrusives and migmatitic veins. The general conformity of veins with the schistosity of host gneisses does not support the forceful injection of magma.

TABLE 7

Observed concentration ranges in biotites from migmatite veins, Cooke Hill tonalites, Cooke Hill granodiorites, massive granodiorite and metasedimentary rocks

Sample No.	K%	Sr ppm	Rb ppm	Ba ppm	K/Rb	Ba/Rb	Ba/K.10^3
372	7.22	4	271	2001	266	7.38	27.71
514	7.48	7	413	1971	181	4.77	26.35
515	7.31	14	540	1934	135	3.58	26.45
474	7.95	4	587	1114	135	1.89	14.01
497	7.31	3	534	1797	137	3.36	24.58
398	7.30	9	242	3056	301	12.62	41.86
Mean	7.42	6	431	1978	192	5.60	26.82
140	8.00	6	364	3911	220	10.74	48.88
233	7.37	8	367	3197	201	8.71	43.37
75	7.40	12	331	3120	223	9.42	42.16
199	7.59	9	488	2408	155	4.93	31.72
Mean	7.59	9	387	3159	200	8.45	41.53
137	7.83	10	533	1348	147	2.52	17.21
232	7.52	5	719	1818	104	2.52	24.17
Mean	7.67	7	626	1583	125	2.52	20.69
208	6.79	9	375	2200	181	5.86	32.40
139	8.11	6	430	2270	189	5.27	27.99
205	7.81	4	614	630	127	1.02	8.06
149	7.49	5	1144	1186	65	1.03	15.83
184	7.49	5	1539	782	49	0.50	10.44
201	7.70	4	348	585	221	1.68	7.59
138	8.13	4	659	601	123	0.91	7.39
Mean	7.78	4	789	1009	129	1.73	12.88

Biotites from migmatite veins	(372, 514,515,474,497,398)
Biotites from the Cooke Hill tonalites	(140, 233,75,199)
Biotites from the Cooke Hill granodiorites	(137, 232)
Biotites from the massive granodiorite	(208)
Biotites from the metasedimentary rocks	(139, 205,149,184,201,138)

5.3.2. Anatexis

Considering the high grade metamorphic condition prevailing during regional migmatization it could be postulated that the leucocratic veins in migmatites originated by differential anatexis (partial melting) of gneisses in situ. Therefore, an estimate of P-T conditions of metamorphism is an important factor for consideration of the "anatexis" hypothesis for the origin of migmatites in this area. In the studied area, metamorphism has reached the sillimanite-orthoclase zone, though some muscovite escaped complete destruction or still occurs as relicts. The complete absence of andalusite (a mineral which is present in less metamorphosed rocks of adjacent areas) and the stability of fibrolitic sillimanite in the present area, suggests that P_{total} = (P_{fluid}) has exceeded the experimentally determined curves (1) and (2) in Figure 15B. The stability of muscovite is shown in Figure 15A under the increasing temperature and pressure conditions. In the light of experimental data, it seems likely that a fluid pressure of over 3kb and temperature up to 650^{c} C was reached in the present area. A similar estimate of P-T conditions in the migmatite zone of Mt. Lofty Ranges is presented by Fleming (1971, p. 33-35), viz. P = 3.6kb and T = 650°C approximately. These temperature-pressure conditions are favourable for partial melting of the low temperature quartz-feldspar constituents of the biotite gneisses if $P_{H_2O} = P_{total}$. Winkler (1965, p. 179) states that if the temperature and pressure of metamorphism reaches P_{H_2O} = 2,000 bars at 700°C and P_{H_2O} = 4,000 bars at 670°C, gneisses are likely to be partially melted, depending upon the composition of rock (especially the Ab/An ratio of plagioclase).

The composition of migmatites has been studied in the light of experimental work on anatexis. The migmatite veins of the present area have more than 80% normative Ab + Or + Q and hence can be applied to the Ab - Or - SiO_2 diagram of Tuttle and Bowen (1958) and Luth, et al. (1964). The plots of normative Q, Or and Ab on the synthetic granite system of Tuttle and Bowen and Luth et al. requires the absence of normative An in the rocks. This, however, is clearly not the case for the analysed rocks of migmatite veins and granitic intrusives, thus their plots on the Q - Ab - Or ternary diagram (Figure 28) give only a rough comparison, as the addition of An produces a marked change in the position of the critical silica-alkali feldspar boundary.

Two general conditions for anatexis have been established by the experimental work of Tuttle and Bowen (1958), Luth, et al. (1964), Winkler (1961, 1965, p. 178-203), Von Platen (1965), Knabe (1966) etc., To promote anatexis in metasediment, (1) the lowest temperature cotectic composition must have a low An content; the higher the An content in the rock composition the higher the temperature that would be required to begin melting; (2) the melt phase will always contain Or in significant amount even if derived from K-feldspar lacking sediments, the potassium needed for the formation of Or (alkali feldspar) in the melt being derived from the breakdown of biotite or muscovite in the original sediments during anatexis.

These two rules are the basic requirements for differential anatexis (partial melting) of country rock. Apart from favourable P-T conditions of metamorphism for anatexis, none of the above rules is obeyed by the Cooke Hill migmatites. It will be shown in the following paragraphs that there are several arguments against this hypothesis:

(a) The normative Or in most of the analysed samples of migmatite veins is extremely low (Tables 3A, B; Figs. 28A, B). This contrasts strongly with the experimental data of Tuttle and Bowen (1958) and Luth et al. (1964); the composition of melts formed at the beginning of partial melting are shown by then as Q : Ab : Or = 35 :

Q : Ab : Or = 35 : 40 : 25 for 2 kb and 27 : 50 : 23 for 5 kb

(b) The mineralogical composition of the migmatite veins is close to tonalite and granodiorite with respect to quartz, plagioclase and K-feldspar ratios. To form a melt of tonalitic/granodioritic composition by anatexis, a temperature range of 690-740°C at 3-4 kb hydrostatic pressure will be needed (e.g. Piwinskii and Wyllie, 1968; 1970).

(c) If the normative compositions of migmatite veins are plotted on the diagram representing the Ab-An-Or-SiO_2 system of Kleeman (1965), they fall away from the differentiation trend of magmatic rocks (note the trend of the Cooke Hill tonalite and granodiorite on the diagrams) as shown in Figs. 29 A&B. The rather close grouping of points representing the composition of migmatite veins near the Ab apex of the diagrams indicate a different petrogenic history than the magmatic/anatectic rocks.

(d) The trace element data of migmatite veins also is not consistent with the composition of a partial melt. It is generally believed that anatectic melts would be enriched in Rb relative to Ba + Sr (White, 1966; Hall, 1967 and Condie, 1969). Experimental studies on anatexis show that the early formed liquids are mainly composed of low temperature fractions such as quartz, sodic plagioclase and K-feldspar, similar to that of late stage fractionated granites. Therefore, Rb which is abundant in low temperature acid granites, would also be expected to be concentrated in early formed low temperature partial melts. Ba and Sr which are abundant in high temperature calcic granites cannot be expected to concentrate in relatively lower temperature partial melts. The Cooke Hill migmatite veins show the reverse trend that is Sr and Ba are much higher than Rb and cannot support the partial melting of gneisses as the cause of migmatization (Table 4).

(e) The possibility has also been considered that an anatectic magma might have been generated outside the migmatite zone in a deeper part of the geosyncline where higher temperature and pressures may have been available to promote melting of plagioclase rich sediments comparable to mineralogical composition of veins. The resulting anatectic liquid may then have migrated some distance during tectonic movement. Applying this idea to the migmatite veins of the present area, one concludes that a magma of such a composition that it plots near the albite corner of the synthetic Ab-An-Or-SiO_2 system, must have been derived from sedimentary rock of a very unusual composition.

None of the above facts support the proposition that partial melting of gneisses was a major cause for the development of leucocratic veins in the Cooke Hill migmatites.

5.3.3. Metamorphic differentiation

During past few decades two major hypotheses have been proposed for the development of quartz-feldspar veins in migmatized gneisses. One of these suggests that vein-forming matter was derived from a remote source and injected in a molten state, while the other hypothesis suggests that vein forming material was derived from the surrounding host rock by some metamorphic process. The first hypothesis has been discussed in the preceding sections and the results of geochemical and petrological data show that veins in the migmatites have not arisen from the injection of magmatic liquids from outside sources nor by differential anatexis of country rock in situ or at deeper zones and the subsequent migration of anatectic liquids to an upper level. The second hypothesis is left as the only alternative cause of migmatization and indeed many structural and petrological features of migmatites can be explained by this hpothesis.

The field and structural studies provide convincing evidence that the migmatization was essentially a metamorphic process. A noteworthy feature of the migmatites is that the mineral species found in the veins are precisely those that make up the enclosing gneissic part of migmatites. This provides a strong indication that some or all of the vein forming matter was locally derived by metamorphic differentiation process.

The actual mechanism of metamorphic differentiation has never been precisely described. Some authorities believe that metamorphic differentiation is a process by which the recrystallization and readjustment of materials take place in the form of transfer of matter on a local scale of the order of a few metres without addition of foreign substances from the remote source. This produces a local segregation of quartz-feldspar in the leucocratic veins and subsequent migration of dark minerals towards the adjacent host gneisses. Other authorities invoke the movement of ions over large distances.

In this thesis, the term "metamorphic differentiation" is used to include considerable movement of materials over short distances, and the term "metasomatism" to include introduction of materials into the local area from sources outside, involving distances of some kilometres. Generally, metasomatism involves substantial addition or removal of materials during this process as defined by Heitanen (1954), Eskola (1939 and 1950), Ramberg (1952, p. 174) and Turner & Verhoogen (1960, p. 562.).

There is evidence that metamorphic differentiation was active on small scale to produce veined gneisses (migmatites) by local segregation of quartz-feldspar in the veins and of biotite at the margins of the adjacent gneisses. Petrographically, the composition of the veins and the host gneisses are the same except that the relative proportion of quartz and feldspar (mainly plagioclase) is much higher in the veins than in the gneisses. A dark rim of biotite surrounds some veins. The scarcity of biotite in the veins compared to neighbouring host gneisses, and the occasional enrichment of biotite in a narrow zone bordering between two rock units of migmatites, may be explained by metamorphic differentiation.

The presence of these biotite selvedges has been interpreted in different ways by some petrologists. Some believe that the scarcity of biotite in the vein and its concentration in the selvedges is due to expulsion of biotite during the growth of vein (Mehnert, 1951, 1968; Ramberg, 1956, 1960; Kretz, 1961 etc.) Misch (1968) disagrees with the concept of simple migration of biotite towards bordering selvedges and the consequent enrichment of leftover quartz-feldspar in the vein. He remarked that if the biotite of selvedges is added to the leucocratic vein, then the resultant composition is still more leucocratic than the adjacent host biotite gneiss. He therefore suggested that metamorphic differentiation is a process in which simultaneous migration of biotite from the growing vein and quartz-feldspar from the neighbouring gneiss takes place. Hence some quartz and feldspar have migrated into the site of vein from the adjacent host rock thus increasing the volume percentage of light coloured minerals.

Hughes (1970) interprets the presence of biotite selvedges as an indication of incipient anatexis. He suggests however that the amount of anatectic liquid is quite small (1 or 2%). This anatectic liquid develops in layers with the appropriate composition and other layers contain aqueous pore fluid. The biotite selvedges mark the boundary between these two layers and are the locus of reaction between the highly mobile aqueous fluid and the relatively immobile anatectic liquid. He moreover states that where anatectic liquids are predominant biotite selvedges do not occur. This hypothesis then suggests that the Cooke Hill migmatites have just reached the temperature (about 650°C at 4 kb P_{H_2O}) at which anatexis may occur in layers containing K-feldspar but that the main activity was due to the aqueous fluid phase, coexisting in the bulk of the rock.

These arguments suggest that migmatization of gneisses was essentially a metamorphic process.

In addition mesoscopic structural elements of veins are conformable with the enclosing host gneisses, that is the mica schistosity and mineral lineation show complete parallelism in both the vein and the adjacent gneiss. This structural interrelationship suggests that quartzofeldspathic material was segregated into the vein by metamorphic differentiation.

A notable feature of the migmatite veins is a considerable variation in shape and size. Some are thin impersistent veins of uniform thickness and others are irregularly large and small boudins. This irregularity of veins suggests that shape was essentially dependent on the vein forming process.

The trace element data of migmatite veins with respect to host gneiss are easily explained if they have originated by metamorphic differentiation process. A comparative study of distribution of trace elements in the vein (sample 372) and its host rock (sample 372 AG) shows that Rb and Ba are abundant in the gneiss, while these elements are relatively less abundant in the vein but that the vein has a higher value for Sr.

Examination of the geochemical data shows that the relative abundance of Rb and Ba in the host gneiss (372 AG) is directly related to the higher modal percentage of biotite and to a lesser extent the greater amount of K-feldspar, whereas the abundance of Sr in the veins (372) is attributable to the high percentage of plagioclase in the vein and the comparative lack of K-bearing minerals (Table 2).

The observed variations of trace elements in other analysed veins of migmatites are to some extent due to variable proportions of plagioclase, biotite, muscovite and less common K-feldspar composing these veins. Similarly, Ca/Sr, K/Rb, Ba/Rb and K/Ba ratios also vary with the changing proportions of major minerals present in the veins.

The above discussions lead to the conclusion that metamorphic differentiation was the major cause of migmatization of gneisses. The strong mineralogical interdependence between vein and gneissic host rock of migmatite suggests that nearly all of the vein forming material was derived from the host rock gneiss. The cause of vein formation and nature of migration of minerals from inside and outside the veins may be related to pressure gradients which generally develop in recrystallisation of sediments.

5.3.4. Metasomatism

The high soda content of all analysed veins is a striking feature of the Cooke Hill migmatites. This is indicated by the higher proportion of plagioclase in the vein than in the enclosing host gneiss. Although there is a strong evidence that metamorphic differentiation played a major role in the formation of the veined gneisses as migmatites, some of the field and petrographic observations can only be explained if some

Na metasomatism is also invoked as a simultaneous process.

The formation of the migmatite veins is considered to have occurred in the presence of a hydrous fluid phase. In this respect, it is pertinent to consider whether some of the sodium may have come from distant sources rather than from a merely local rearrangement of material. Generally the plagioclase of the veins is corroded and in part replaced by sericite or muscovite. Moreover, the vein plagioclase is more sodic than the plagioclase of the host rock. This difference is not great, of the order of 1 to 4% An but plagioclase of the veins is always consistently more sodic than that in the host. This is reflected in the Na/Ca ratio of 372 (vein) and 372 AG (host gneiss) which are 4.9 and 4.4 respectively. This may indicate that some Na has been introduced into the veins from distant sources, but it may also indicate that Na was more mobile than Ca under the condition of metamorphic differentiation .

It is believed that both addition and removal of some material has taken place during metasomatic alteration of plagioclase by the hydrothermal fluids migrating through the pore space of recrystallizing rock. It is difficult to make a quantitative estimate of material introduced or removed during metasomatism without knowing the volume changes and the original composition of pre-migmatized gneisses (Misch, 1968). The circulating fluids acting as a metasomatic agent were not of igneous origin (see discussion in the preceding section), therefore such fluids must be of hydrothermal nature and may be derived from the pore solution of original sediments which are capable of dissolving certain elements, especially alkalies during high temperature and pressure conditions of metamorphism. It is generally considered that below a depth of 5 km. at 300°C, albite is more soluble than quartz (Bowes, 1967). Also Orville's data (1963) indicates that in solid-vapour system for feldspar, and increase in pressure increases the amount of sodium (albite component) in solution. Therefore, fluids migrating from greater depths would be expected to be sodic rather than potassic and thus were involved in soda metasomatism in the migmatite veins of the present area.

5.4. CONCLUSIONS

In the light of structural, petrological and chemical data, it can be concluded that bodily injection of granite magma or magmatic "juices" was not involved in the formation of leucocratic veins of migmatites. Similarly, the veins in the migmatized gneisses have not originated by partial melting of the country rock as the maximum temperature attained during the regional metamorphism of the Cooke Hill area was about 650°C, whereas the composition of veins suggest that a temperature range of 690-740°C would be required to form an anatectic magma for such veins. Also the trace element chemistry of migmatite veins is not consistent with the partial melts.

All the observed facts related to chemical and mineralogical compositions of veins and the structural parallelisms between the conformable veins with the surrounding host gneisses can be simply explained by the metamorphic differentiation. During tectonic deformation and metamorphic recrystallisation, the rock material was redistributed in the form of leucocratic veins giving rise to the migmatites. It is inferred from the field relations and petrographic studies that the process of migmatization was one of the metamorphic differentiation in which the felsic material became more mobile in recrystallizing sediments on a scale of a few cm to a few metres and tended to be concentrated in structurally controlled areas to form concordant veins in the gneisses. The sodic tendency and the extensive clouding and sericitization of the vein plagioclase is related to some Na-metasomatism by the hydrothermal fluids, which was almost simultaneous with the metamorphic differentiation during the growth of veins in migmatites.

5.5. COMPARISON BETWEEN THE COOKE HILL MIGMATITES AND THE PALMER MIGMATITES

There are considerable differences between the Cooke Hill migmatites and the Palmer Migmatites, despite the apparent similarities in the field. In both areas the veins are developed along the foliation of the enclosing gneisses and in both areas the grade of metamorphism was sillimanite-orthoclase sub-facies. Both the data in this thesis and that of White (1966a) point to the origin of the veins by metamorphic differentiation and give no support to the suggestion that the granites had made any appreciable contribution to the migmatite veins.

The Cooke Hill veins are rich in sodium and are of tonalite composition, whereas the Palmer veins are close to the composition of potash aplites. The data for Rb, Ba and Sr reflect the gross mineralogy. It is a well established fact that Rb enteres more readily into the K position of micas than into K-feldspar and only to a limited extent into plagioclase (Heier & Taylor, 1959: Taylor & Heier, 1960, Iida, 1961). Ba on the other hand enters more readily into K-feldspar than into mica, with less going into plagioclase (Taylor, 1965, Heier, 1962). At magmatic temperatures (circa, 600-700°C) Sr goes into K-feldspar and plagioclase in equal quantities with little or none going into the mica (Sen et al., 1959, Brooks, 1968), Berlington & Henderson, 1969). This trend is for Ba to increase with increase of the amount of K-feldspar, Rb to increase with K-feldspar and micas and Sr to follow the amount of total feldspar.

The Cooke Hill veins which have low proportions of K-feldspar and biotite have only 19 ppm of Rb in contrast to 531 ppm in the Palmer veins. Ba follows the K-feldspar vein closely with 329 ppm in the Cooke Hill veins as against 2707 ppm in the Palmer veins. Sr, which enters into the two feldspars with equal ease shows similar values for Cooke Hill (331 ppm) and Palmer (284 ppm). This variation of the abundance of Rb, Ba, Sr is consistent with metamorphic differentiation under equilibrium conditions.

The difference between the migmatite veins of the Cooke Hill and the Palmer area appears to be related to the composition of metasediments from which they have been derived by metamorphic differentiation accompanied by some metasomatism. The metasediments of the Cooke Hill area have a high Na/K ratio (average 1.56) which produced an aqueous fluid phase rich in soda and calcium whilst the metasediments in the Palmer area have low Na/K ratio (average 0.36), thus producing a fluid in which potash predominates (Orville, 1963). Evidently, Rb and Ba followed K in the fluids of the

Palmer area which with increase of potash feldspar also increased these trace elements in the veins, while Sr and some Ba followed Ca and Na in the fluids of the Cooke Hill area, causing an enrichment of plagioclase along with these elements in the veins of migmatites.

CHAPTER 6

PETROLOGY OF THE CALC-GNEISSES AND CALC-SCHISTS

6.1. INTRODUCTION

The western portion of the area mapped is occupied by a belt of mildly calcareous rocks which are shown on the Adelaide 1:250,000 map sheet as "diopside gneiss locally grading into granitic gneiss". These rocks were grouped with the Rathjen Gneiss and other "granitic" gneisses as being rocks of igneous aspect. Their position in the stratigraphic column is doubtful. White (1956) considered them to be metasediments but was uncertain of their stratigraphic position.

The calc-gneisses are resistant to weathering and stand out as a long continuous ridge with a flat top, surrounded on either side by deeply eroded valleys occupied by the calc-schists. The contacts between the gneisses and schists are structurally conformable and extend in a north-south direction parallel to strike of the quartzo-feldspathic gneisses and migmatites which occupy most of the area (Map 1). They disappear into the country rock and alluvium in the north but apparently form a tight anticlinal structure probably a first generation fold (Sect. 2.4.3.). These rocks are essentially quartzo-feldspathic but contain amphibole and clinopyroxene as essential constituents, and therefore have been called the calc-schists (actinolite-tremolite schists) and the calc-gneisses (diopside gneisses). These calc-gneisses and calc-schists are quite different both in physical appearance and mineralogical composition from the calc-silicate rocks associated with the Milendella Marbles. Calcite, scapolite and/or K-feldspar are the important constituents of the Milendella calc-silicate rocks, but these minerals are completely absent from the "calc-schists" and "calc-gneisses". Although the presence of amphibole and clinopyroxene is indicative of the calcareous nature of the calc-schists and the calc-gneisses, these rocks are far less calcareous than the Milendella calc-silicate rocks (see Sect. 3.3.). The present mapping and structural studies indicate that the Milendella Marbles and calc-silicate rocks belong to the Strangway Hill Formation which is the oldest formation, while the calc-schists and calc-gneisses of the western part of the area studied belong to the upper part of the Inman Hill Formation which overlies the Strangway Hill Formation.

The calc-gneisses occur as two continuous uninterrupted beds, each about 7 km in length (Map 1). The thickness of the beds is variable, locally thickening and thinning in places, probably due to tectonic deformation. The maximum thickness is about 800 m. The calc-schists are wrapped around the gneissic bodies with gradational contacts. In the field the schists are characterised by a finer texture and more highly foliated and lineated than the gneisses (Plate 40A). The gneisses are coarser than the schists and are moderately lineated rocks. Tor-like boulders are commonly found in the exposed outcrops of the gneisses (Plate 40B). The schists split along the foliation planes but the gneisses, which have a texture comparable to the granitic gneiss, break unevenly. Hence the gneisses behaved as resistant rocks during the plastic deformation. During the tectonic deformation, the mafic minerals segregated into streaks and lenticles which together with dimensionally oriented quartz and/or plagioclase, resulted in the development of pronounced lineation. Banding and compositional layering (relict bedding) is quite rare in the gneisses except as thin streaks and lenticles rich in diopside, ranging in thickness from a fraction of 1 to 10 cm.

The calc-schists are marked by alternate dark and light bands formed due to concentration of amphibole and quartz-plagioclase in respective bands. Generally, the schists are soft friable rocks with a saccharoidal texture. Due to poor coherency of grains in some schists, they can be powdered between the fingers. Mechanical weathering has deeply eroded these rocks, and therefore the outcrops are either concealed under the soil cover or discontinuously exposed in the field. Preferred orientation of amphibole prisms produces a strong lineation and foliation.

The calc-schists are essentially of two types, actinolite schists and tremolite schists, the actinolite schists being much more abundant. In hand specimen, both types are fine to medium grained (average grain size 0.5 mm) light coloured rocks. The tremolite schists form compositional bands or layer in the actinolite schists, ranging in thickness from about 10 cm to 8 m. The regularity and uniformity of bands or layers reflects the original sedimentary character suggesting the different compositions of sediments. In the field, the tremolite schists are easily identified by their white colour because the colourless amphibole imparts no tone to these rocks, whereas the actinolite schists have a greenish grey appearance due to the green colour of the actinolite.

6.1.1. On contact relationships between calc-schists and calc-gneisses and the country rocks

The contact between the calc-schists and the calc-gneisses is concordant. Lineation and foliation are parallel in the two rock types on either side of the contact. The coarser grained calc-gneisses decrease in grain size towards the calc-schists with significant change in mineralogical composition of the marginal gneisses. These changes begin to develop in gneissic rocks at distances of up to a hundred metres from the boundary contact. In this gradational zone, diopside of the calc-gneisses gradually gives way to amphibole. The amphibole in the marginal gneisses may be pale green weakly pleochroic actinolite or tremolite or bluish green strongly pleochroic hornblende depending on the original composition of the adjacent rocks. The calc-schists also show similar changes in gradational zones towards the diopside gneisses in that diopside may become an important constituent along with amphiboles in the marginal schists. These observations are based on carefully selected samples from exposed outcrops showing the sharp boundary contacts between the calc-schists and the calc-gneisses. Unfortunately, much of the contact is covered by soil and little opportunity was available for detailed sampling. Nevertheless, a gradual change in textures of the two rock types was commonly noticeable in the field when approaching the contact. These gradational features are much more prominent in the coarse grained gneisses where a change in texture is quite significant on mesoscopic scale and also involves a larger distance, whereas in the schists textural variations are limited to a short distance and abrupt changes in composition are common.

Where the calc-gneisses come into contact with quartzo-feldspathic gneisses and migmatites (country rocks) they again show gradational and compositional changes on various scales. These changes are limited to a zone ranging in width from less than 30 cm to about 10 m. These field observations have been made in the northern end of the large gneissic body where the quartzo-feldspathic rocks interfinger with the diopsdde gneisses or in the south-western part of the area where the diopside gneiss is in direct contact with the quartzo-feldspathic gneiss rather than with the calc-schist. The diopside gneisses show a general concordance with the country rock along the strike and their planar and linear structures are conformable. The diopside gneisses grade into the country rock through two successive zones, hornblende gneiss and biotite gneiss. These compositional changes across the contact of the two rocks appear to represent a gradual compositional variation of the original sediments.

The calc-schists in contact with the quartzo-feldspathic gneisses also show significant changes in their marginal zones. These changes are confined generally to a relatively narrow zone (from a few cm to a metre) as compared to the calc-gneisses. In the transitional zone lying between the quartzo-feldspathic rocks and the amphibole calc-schists, diopside is often stable along with hornblende. Sometimes, diopside is absent and hornblende is the only mafic mineral. The amphibole of the normal type of calc-schists is either actinolite or tremolite but near the marginal zone adjacent to the quartzo-feldspathic gneisses it becomes much more strongly pleochroic (olive green to bluish green) and attains the characteristics of hornblende (Sample No. 150; Table 8). The petrographic observations indicate that some amphibole and diopside may have resulted by breakdown of magnesian biotite in the original sediments during the isochemical metamorphism (e.g. Carmichael, 1970; Mills, 1964, p. 142-45). This is shown by relicts of original biotite included or closely associated with hornblende or diopside with occasional K-feldspar as a by-product. Some biotite has escaped complete destruction and occurs as small pale brown flakes indicative of phlogopite.

Occasionally, anthophyllite is found in the marginal part of the calc-schist at the contact with the quartzo-feldspathic gneiss near some small acidic intrusive bodies (e.g. Locations 148037, 149046 etc.). Its rarity in the calc-schists except near the contact with igneous intrusion suggests that probably anthophyllite is stable in the higher temperature conditions (Fisher, 1966, Reaction 6; Butler, 1969).

There is no evidence that metasomatism has played a significant role in the formation of the transitional zone lying in between the calc-schists and the calc-gneisses. The field evidence strongly suggests that the rocks are metamorphosed sediments and that variation in the amount of calcite and chlorite in the original sediments may be responsible for the variations in mineralogy.

6.1.2. Petrography of the calc-schists

The schists are fine to medium grained rocks; the average grain size ranges from 0.2 to 0.5 mm. Occasionally coarser varieties have been seen, in which the porphyroblasts of quartz, plagioclase and amphibole may reach

2.5 mm in diameter. The major components of the schists are plagioclase 40-60%, quartz 25-40% and amphibole 5-20% (Plate 41A) Sphene, apatite, iron ores, rutile, biotite and zircon are the common accessory minerals. Diopside may occur in marginal parts at the contact with the calc-gneiss. Sphene is ubiquitous, by far the most abundant among the accessory minerals and may be present in amounts up to 10% in some rocks. The relative proportion of quartz, plagioclase and amphibole is variable especially in the banded rocks, resulting in quartz-plagioclase rich bands and amphibole rich bands

Quartz occurs as scattered, intersititial grains with curved boundaries. A varying degree of strain is evident in quartz grains. Occasionally it replaces plagioclase, and rarely amphibole (cf. Ramberg, 1947). Some of the porphyroblastic grains of quartz contain tiny inclusions of plagioclase and amphibolite.

Plagioclase is generally untwinned but a few twinned crystals are always present in thin section. It has irregular outlines with granoblastic appearance. Deformation or glide twinning is present. Fading of twin lamellae and strain effects such as undulous extinction are common. Weak normally zoned plagioclases are often found. Bending and fracturing of the twin lamellae are seen in a few thin sections. Generally, the plagioclase is fresh but some are cloudy and highly sericitized grains. Small inclusions of quartz, amphibole or sphene may be present in some grains. The composition of plagioclase varies from An_{15-34}.

Amphibole occurs in variable shapes and sizes. It appears as small isolated euhedral prisms or irregular stumpy porphyroblasts in quartz-plagioclase matrix. It may be colourless and non-pleochroic tremolite or olive green to green moderate pleochroic actinolite or pale green to bluish green with strong pleochroism as hornblende. The bluish green pleochroic colour shown by hornblende is probably due to high sodium contents as suggested by Kranck (1961), Klein (1966) and Butler (1969). Anthophyllite occurs infrequently in the calc-schists (Plate 41B). The larger amphiboles are slightly poikiloblastic and contain inclusions of sphene, quartz and biotite. Diopside is a rare mineral in the normal type of schists except in the marginal zone where it may be present along with amphibole.

It occurs as ovoid porphyroblasts often closely associated with amphibole. The diopside is pleochroic from light green to very faint green.

Sphene is the most abundant accessory mineral as anhedral rounded grains. Generally, it clusters around the mafic minerals. Apatite is ubiquitous, occurs as euhedral prismatic crystals or anhedral grains. Iron ores may be found throughout the rock as subrounded to skeletal grains. It is common in some rocks and in others may be completely absent. Biotite (phlogopite) is a pale yellow weakly pleochroic small flaky mineral. Rutile is a less common accessory mineral. Minute crystals of zircon with subrounded outline are often seen. Small granules of colourless to weakly pleochroic, pale green epidote are found in highly sericitized plagioclases.

6.1.3. Petrography of the calc-gneisses

The mineralogical composition of the calc-gneisses is fairly simple; the major components are clinopyroxene, plagioclase and quartz (Plate 42A). The accessory minerals are sphene, rutile, biotite, apatite and iron ores. The common secondary minerals are epidote and sericite. The normal texture of the rock is granoblastic with average grain size about 1 mm. Generally, the texture in the marginal part of the gneissic body is somewhat finer.

The observed variation in volume proportions of major minerals is : quartz, 20-40%; plagioclase, 40-60%; clinopyroxene, 5-20%; hornblende, 0-15%. The hornblende bearing gneisses are restricted to the marginal parts of the gneissic body or occur as isolated lenses in the calc-schists close to contact boundaries of calc-gneisses. A brief petrographic account of individual minerals is given below.

Quartz is much coarser in the calc-gneisses than in the calc schists. The grain size ranges from 0.5 mm to 1 mm but a few grains reach up to 2 mm. The quartz occurs as clear grains, whose grain boundaries with each other, and with plagioclase, are gently curved. Straining and undulose extinction is present in some grains. Some porphyroblastic grains of quartz contain small inclusions of plagioclase and mafic minerals.

Plagioclase is generally fresh and clear in most rocks. It is distinguished from quartz with difficulty as the twinning is not present in most grains. Some porphyroblastic grains contain small inclusions of other minerals. Zoning is rare, it is always weak and normal type. Because of rarity of twinning, little information about the composition of plagioclase could be obtained from extinction angles. It appears that plagioclase generally has a composition in the oligoclase range but some more calcic compositions (up to An_{35}) were also determined.

Clinopyroxene appears as small anhedral or as large skeletal porphyroblasts (Plate 42B). It is colourless or pale green, with low to moderate pleochroism indicating a low iron content. Small inclusions of quartz, sphene, and biotite may be seen in some larger grains.

Amphibole is olive green to pale bluish green in colour with strong pleochroism. It is rather less common in the typical calc-gneisses than in the calc-schists except in the gradational zone (or a few small lenses in the calc-schists). Three chemical analyses of amphibole indicate that they lie on the boundary line of actinolite and hornblende compositions (Sample Nos. 173A, 173B, 150; Table 8B). The X-ray powder patterns of actinolite and hornblende show slight differences, but all strong lines match well with standard hornblende patterns.

Sphene and apatite are ubiquitous accessory minerals. Sphene is euhedral to anhedral in shape, colourless to slightly brown, either closely associated with or included in diopside and amphibole. Apatite occurs as a small prismatic crystal to subrounded anhedral grains.

Iron ores, rutile, biotite and zircon may be found in some thin sections, but are completely absent in others. Epidote is a secondary mineral and appears as small granules. It occurs as a colourless non-pleochroic mineral, clinozoisite, in some rocks, or as a weakly pleochroic pale yellow epidote in others. Epidote is always found associated with highly sericitized plagioclase, indicating its derivation from alteration of plagioclase. White mica is another secondary mineral which is found in highly altered "dusty" plagioclase.

6.2. CHEMISTRY OF THE AMPHIBOLES

In Tables 8A & B, the chemical analyses of amphiboles representing the minerals from different parts of the calc-schists are presented (see Map 4 for location of samples). Analyses 117, 122, 123 and 109 come from the normal type of calc-schists; Analyses 454, 112, 173A and B* are from the gradational zone between the calc-schists and the calc-gneisses and analysis 150 comes from a gradational zone adjacent to micaceous quartzo-feldspathic gneiss. Analysis 145 is from the calc-schist close to a small granitic intrusive body. The analyses of six amphiboles in Table 8A are similar to analyses of the tremolite-actinolite series as reported by Deer, Howie and Zussman (1963, vol. 2 table 38-39). Using their classification sample 109 and 112 are tremolites, and 117, 122, 454 and 123 are actinolites.

Three amphiboles (sample no. 173-A*, 173-B* and 150)in Table 8B are more highly coloured and strongly pleochroic (X = pale green, Z = 2 = dark bluish green and contain a higher amount of aluminium, iron and sodium. Chemically, the composition of these three amphiboles is similar to those of low alumina hornblendes of Deer, et al., (1963, vol. 2, Table 40).

A rather less common mineral is anthophyllite (Sample No. 145, Table 8B) which generally occurs in marginal zone of calc-schist in contact with micaceous gneiss and close to intrusive bodies.

The composition of calcic amphiboles (tremolite, actinolite and hornblende) and of the anthophyllite has been recalculated on the basis of 23 oxygens (Tables 8 A, B) and the atomic ratios Ca: Mg: Fe (total) have been compared with the data of Deer, Howie & Zussman and others. In Fig. 41, all calcic amphiboles fall within the boundary of tremolite-actinolite and hornblende field and one amphibole in the compositional field of anthophyllite as defined by Deer et al., 1963, Vol. 2, Fig. 68, Klein (1968) and Butler (1969).

In Figs. 42 and 43 the plots of Al^4 against (Na + K) and (Al^6 + Fe + Ti) atoms of the calcic amphiboles show that they fall in the compositional field of tremolite-actinolite and hornblende

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* Sample no. 173 is a marginal zone rock consisting of alternate bands of medium grained schist (173-A) and coarse grained gneiss (173-B) with hornblende as the only mafic mineral in both rock units.

as defined by Deer et al., (1963, Vol. 2, Figs. 71 & 72). Figures 42 and 43 clearly show that amphiboles from the normal type of calc-schists are tremolite and actinolite but those found in the marginal zones have higher amounts of aluminium, iron and alkalies, and are therefore hornblende.

In the light of chemical data, the bulk composition of rocks appears to exert a pronounced control over the chemistry of calcic amphiboles (e.g. Ernst, 1968). This fact is borne out by the uniformity of the composition of amphiboles from the normal type of calc-schists compared with the marginal variants adjacent to calc-gneisses and micaceous gneisses. The scarcity of anthophyllite as compared to the calcic amphiboles may be due to its narrow P-T stability field and the peculiar bulk composition of the rocks in which it occurs (especially the deficiency of calcium). The characteristic occurrence of anthophyllite in the schists near the igneous intrusives suggests that probably higher temperatures than those prevailing during the regional metamorphism of country rock may be the major factor for the genesis of anthophyllite.

6.2.1. Chemistry of Pyroxenes

Four pyroxenes have been analysed chemically (Table 9). Three pyroxenes (Specimen Nos. 121. 118 and 373) were separated from the normal type of calc-gneisses and one pyroxene (Specimen No. 178) was separated from the marginal zone of gneiss grading towards the calc-schist. This marginal rock also contained a small amount of actinolite but it was impossible to separate a clean sample of actinolite for chemical analysis.

The analysis of two pyroxenes (Sample Nos. 121 and 118) show higher percentages of iron and lower percentage of magnesium compared with the other two pyroxenes (Sample Nos. 178 and 373). Similarly, the $Mg/Mg + Fe^{+2} + Fe^{+3} + Mn$ ratios of these two types of pyroxenes are distinctly different from each other (see Table 9). The chemical compositions of more iron rich pyroxenes are similar to salite and the magnesium rich pyroxenes to diopside (see Deer et al., 1963, Vol. 2, Table 5).

The chemical analyses of pyroxenes were recalculated on the basis of 6 oxygens (Table 9). The cation percentages of Ca: Mg: Fe (total) of these minerals are plotted in Fig. 41 which further confirm that two lie in the compositional field of salite and two in the field of diopside as shown by Deer et al., (Vol 2, Fig. 1). Thus optical properties of these clinopyroxenes also match their compositions as the colour and pleochroism become stronger in the salites with the increase in iron whereas diopside remains almost colourless with no pleochroism.

6.2.2. Petrochemistry

Three representative samples, one for calc-gneiss (21), one for calc-schist (123) and one for the rock (178) transitional between the schist and gneiss were chemically analysed (Table 10).

The analysed rocks are notable for the high silica and sodium contents and low potash, iron and magnesium contents. This shows up in the normative minerals - Qz ranges from 36% to 41%, Ab from 42% to 54% and An over 3% to II%. The ferromagnesian minerals represented by normative Di, Wo and Hy constitute a smaller percentage of these rocks.

The trace element data of these rocks is shown in Table 10. Sr is the only element which is present in significant amount; other elements such as Rb and Ba are almost absent. In these analysed rocks the Sr appears to increase with increasing content of normative An, that is the lowest amount of Sr is found in the calc-schist (123) which has the lowest content of normative An, whereas the highest value of Sr was found in the diopside-actinolite gneiss (178) containing the highest An component. Thus Sr is found on mainly Sr in plagioclase because other co-existing minerals like calcic amphibole and clinopyroxene generally contain Sr in the range of 10 to 30 ppm. The depletion of Ba and Rb in these analysed rocks is due to rarity of potash bearing minerals.

6.3. PARAGENESIS OF CALC-SCHISTS AND CALC-GNEISSES

The field relationships suggest that the calc-schists and calc-gneisses have been derived from the re-crystalisation of sediments. The intimate association of calc-schists with calc-gneisses and marked gradational and mineralogical changes between them and the surrounding country rock favour a sedimentary parentage.

The present mineralogy of these rocks shows the predominance of plagioclase over quartz with lesser amounts of ferromagnesian minerals. This suggests that the original sediments were rich in quartz and plagioclase with some chlorite and carbonate minerals. The distinctive occurrence of amphiboles in the calc-schists and pyroxenes in the calc-gneisses indicates that these minerals were derived from compositionally different sediments. A notable feature of schists and gneisses is the complete absence of calcite in the re-crystallised sediments. This indicates that the original carbonate content of these rocks was not very high and may have been present as cementing material in quartz rich sediments.

To get the idea of original carbonate content of these rocks, the analyses have been recalculated. The actual amount of lime present in the form of carbonate mineral in the original sediments has been derived after correcting for plagioclase and sphene. If the calcium used in the plagioclase and sphene is deduced from the total CaO of the rock compositions, then the remaining calcium will represent the original component of carbonate mineral. These corrections have been applied to three analysed rock samples which show that Ca' values (calcium available after plagioclase and sphene is calculated) are 1.62% in the clinopyroxene bearing gneiss (121), 0.89% in the diopside-actinolite schist (178) and 0.27% in the actinolite schist (123). Thus prior to metamorphism the calculated calcites were in the range of 4.1% in the calc-gneiss, 2.2% in the diopside-actinolite schist and 0.7% in the calc-schist.

Thus during the early stage of metamorphism, calcite reacted with quartz and chlorite of the original sediments to form calcic amphiboles by the following reaction:

$$3(OH)_8(Mg,Fe)_5Al_2Si_3O_{10} + 10\,CaCO_3 + 21SiO_2 \rightleftharpoons \underset{\text{actinolite}}{3(OH)_2Ca_2(Mg,Fe)_5Si_8O_{22}} = \underset{\text{epidote}}{2(OH)Ca_2Al_3Si_3O_{12}} + 8H_2O + 10CO_2 \quad \ldots\ldots(1)$$

During the later stage of progressive metamorphism, the remaining chlorite of the re-crystallising sediments reacted with epidote to form actinolite and anorthite as shown below:

$$\underset{\text{epidote}}{6(OH)Ca_2Al_3Si_3O_{12}} + \underset{\text{chlorite}}{(OH)_8(Mg,Fe)_5Al_2Si_3O_{10}} + 7SiO_2 \rightleftharpoons$$

$$\underset{\text{actinolite}}{(OH)_2Ca_2(Mg,Fe)_5Si_8O_{22}} + \underset{\text{anorthite}}{10CaAl_2Si_2O_8} + 6H_2O \quad \ldots\ldots\ldots\ldots\ldots(2)$$

The preponderance of actinolite compared to tremolite in the calc-schists indicates that the above reactions are responsible for the formation of calcic-amphibole. However certain bands of calc-schists do contain tremolite along with actinolite and hence the carbonate may be of dolomitic nature to give rise to tremolite by the following reaction:

$$5 \text{ dolomite} + 8 \text{ quartz} + 1\ H_2O \rightleftharpoons 1 \text{ tremolite} + 3 \text{ calcite} + CO_2 \ldots.(3)$$

The by-product of calcite from the above reaction (3) further reacted with chlorite and quartz to produce actinolite or iron rich tremolite.

The formation of hornblende in the marginal zone of calc-schists adjacent to biotite gneisses may be explained by the following reaction (e.g. Winkler, 1967 p. 83).

$$\text{chlorite} + \text{actinolite/tremolite} + \text{epidote} + \text{quartz} \longrightarrow \text{hornblende} \ldots.(4)$$

The formation of diopside and salite in the coarse grained calc-gneisses is the result of their higher content of calcite relative to the calc-schists. The clinopyroxenes appear to be formed by the destruction of actinolite in the higher grade of progressive metamorphism:

$$1 \text{ actinolite} + 3 \text{ calcite} + 2 \text{ quartz} \rightleftharpoons 5 \text{ diopside} + 3\ CO_2 \pm H_2O \ldots..(5)$$

Fig. 41 shows that the camphiboles and pyroxenes have the same range of Mg/Fe ratios so that the only difference is the availability of $CaCc_3$ to convert the actinolite to diopside (salite).

The paragenesis of primary mineral assemblages of the calc-schists and calc-gneisses has been explained by the above reactions, thus the formation of calcic amphibole in the calc-schists and clinopyroxene in the calc-gneisses is related to original lime contents of these recrystallised rocks. When lime is present in considerable amount then clinopyroxenes form, and when it is present in smaller amounts, only calcic amphiboles form (e.g. Edwards & Baker, 1954). This is true with the clinopyroxene bearing gneisses of the present area which contain considerably higher amounts of CaO than the amphibole bearing schists.

Heitanen (1971) has suggested that the Ca'/(Mg + Fe) ratios of the rocks is the main factor controlling the crystallization of calcic ampbibole or clinopyroxene. If Ca'/(Mg + Fe) is less 0.4, actinolite only forms; if the ratio lies in between 0 4 - 1.0, both actinolite and diopside will form and when this ratio exceeds 1.0 only diopside forms. The ratios in three analysed rocks of the present area are in accord with these ideas. As shown in Table 10; Rock 121, with a Ca'/(Mg + Fe) ratio of 0.97, has diopside (salite) only; rock 178, with a ratio of 0.71, has both diopside and actinolite and rock 123, with a ratio of 0.16 has actinolite only.

6.3.1. The Origin of the Calc-Schists and Calc-Gneisses

The paragenesis outlined above supports the idea that these rocks were originally sediments of an unusual nature. The only possible alternative origin is that they were sodic rhyolites or their tuffaceous equivalents. The major element chemistry shows some similarities with sodic rhyolite, but the trace element data do not support this hypothesis. Van der Kamp (1968) quotes figures for calc-silicate rocks of tuffaceous origin, with Ba 100-60 ppm Rb 10-90 ppm and Sr 300-700 ppm. The Cooke Hill rocks have very little Ba and Rb (of the order of 1 ppm or less) and only Sr lies within the range quoted. On the other hand a greywacke sediment with some calcitic cement could well have a relatively high Sr content but completely lack Ba and Rb. It is notable that the three analysed specimens have only about 0.1% K_2O, which is about the amount which would be expected in detrital plagioclase. Hence the amounts of Ba and Rb are what would be expected from a quartz-plagioclase sediment. Sr can be both in plagioclase and in the calcite matrix.

Similar potash deficient rocks have been reported from other areas of the Kanmantoo Group. White (1956) reports anthophyllite and chlorite schists in the Rockleigh area, south of Palmer and Chinner (1955) found potash deficient gneisses east of Mt. Kitchener. Mills (1964) also met potash deficient rocks in the Springton area, only a few kilometres north of Cooke Hill.

TABLE 8A

Chemical analysis of amphiboles

Sample No.		117	122	109	454	112	123
SiO_2		52.83	53.40	54.69	52.93	55.01	53.75
TiO_2		0.37	0.41	0.25	0.35	0.69	0.28
Al_2O_3		3.35	2.71	3.10	3.63	3.08	3.11
Fe_2O_3		1.89	1.38	0.59	1.41	0.55	1.15
FeO		5.51	5.65	2.85	8.19	4.35	5.54
MnO		0.28	0.24	0.14	0.27	0.25	0.24
MgO		21.93	22.40	24.48	19.60	23.50	22.95
CaO		12.18	11.82	12.51	12.13	12.64	11.11
Na_2O		0.80	0.92	0.63	0.70	0.59	0.77
K_2O		0.16	0.09	0.08	0.12	0.14	0.11
L.o. Ignit.		1.60	1.69	1.49	1.50	n.d.	1.86
Total		100.90	100.71	100.81	100.83	100.80	100.87
		Mineral formula on the basis of 0 - 23					
Si		7.33	7.41	7.45	7.40	7.43	7.43
Al		0.55	0.44	0.45	0.60	0.49	0.51
	Z	7.88	7.86	7.90	8.00	7.92	7.94
Fe^{+3}		0.20	0.14	0.06	0.15	0.06	0.12
Ti		0.04	0.04	0.03	0.04	0.07	0.03
Mg		4.54	4.64	4.97	4.09	4.73	4.73
Fe^{+2}		0.64	0.66	0.32	0.96	0.49	0.64
Mn		0.03	0.03	0.02	0.03	0.03	0.03
	Y	5.45	5.51	5.44	5.27	5.38	5.55
Ca		1.81	1.76	1.82	1.82	1.83	1.64
Na		0.21	0.25	0.17	0.19	0.16	0.21
K		0.03	0.02	0.01	0.02	0.02	0.02
	X	2.05	2.03	2.00	2.03	2.01	1.87
$Mg/Mg+Fe^2 +Fe^3 +Mn$		83.9	84.8	92.5	78.2	89.1	85.7

Al in 6 fold coordination is not present in any of these samples

TABLE 8B

Chemical analyses of amphiboles

Sample No.	173-A	173-B	150	145
SiO_2	49.63	49.79	47.31	56.65
TiO_2	0.65	0.69	0.62	0.09
Al_2O_3	6.29	5.87	6.45	0.97
Fe_2O_3	1.33	3.36	5.94	0.16
FeO	7.91	6.53	10.25	14.86
MnO	0.20	0.20	0.40	0.24
MgO	19.10	19.69	14.22	25.40
CaO	12.04	12.00	11.21	0.34
Na_2O	1.33	1.28	1.70	0.17
K_2O	0.19	0.20	0.60	0.02
L.o. Ignit.	1.06	n.d.	1.24	0.96
Total	99.73	99.61	100.67	99.86
Mineral formula on the basis of O = 23				
Si	7.03	6.98	6.90	7.83
Al	0.97	0.97	1.10	0.16
Z	8.00	7.95	8.00	7.99
Al	0.08	0.00	0.01	0.00
Fe^{+3}	0.14	0.36	0.65	0.02
Ti	0.07	0.07	0.07	0.01
Mg	4.03	4.12	3.09	5.24
Fe^{+2}	0.94	0.77	1.25	1.72
Mn	0.02	0.02	0.03	0.03
Y	5.28	5.34	5.10	7.02
Ca	1.83	1.80	1.75	0.05
Na	0.37	0.35	0.48	0.05
K	0.03	0.04	0.11	0.00
X	2.23	2.19	2.34	0.10
$Mg/Mg+Fe^2+Fe^3+Mn$	78.3	78.2	61.5	74.8

TABLE 9

Chemical analyses of clinopyroxenes

Sample No.	121	178	373	118
SiO_2	52.79	53.38	53.53	50.81
TiO_2	0.21	0.27	0.23	0.18
Al_2O_3	0.98	1.18	0.68	2.15
Fe_2O_3	2.87	1.94	0.20	1.58
FeO	5.50	3.20	3.88	7.21
MnO	0.28	0.17	0.19	0.31
MgO	14.53	16.34	17.35	14.33
CaO	22.22	22.27	23.98	22.38
Na_2O	1.20	1.09	0.52	0.94
K_2O	0.04	0.05	0.04	0.05
L.o. Ignit.	0.17	0.21	n.d.	0.32
Total	100.79	100.10	100.60	100.26
Mineral formula on the basis of O = 6				
Si	1.95	1.96	1.96	1.90
Al	0.04	0.04	0.03	0.10
Z	1.99	2.00	1.99	2.00
Al	0.00	0.01	0.00	0.00
Fe^{+3}	0.08	0.05	0.01	0.04
Ti	0.01	0.01	0.01	0.01
Mg	0.80	0.89	0.94	0.80
Fe^{+2}	0.17	0.10	0.12	0.23
Mn	0.01	0.01	0.01	0.01
Y	1.07	1.07	1.09	1.09
Ca	0.88	0.87	0.94	0.91
Na	0.09	0.08	0.04	0.07
K	0.00	0.00	0.00	0.00
X	0.97	0.95	0.98	0.98
$Mg/Mg+Fe^2+Fe^3+Mn$	75.6	85.1	87.9	74.1

TABLE 10

Chemical analysis and C.I.P.W. norms of calc-gneisses and clac-schists

Sample No.	121	178	123
SiO_2	76.13	77.64	78.12
TiO_2	0.56	0.28	0.30
Al_2O_3	12.19	12.60	11.91
Fe_2O_3	0.20	0.23	0.05
FeO	0.53	0.40	0.48
MnO	0.08	0.07	0.08
MgO	0.82	0.57	0.88
CaO	3.51	3.47	1.17
Na_2O	5.67	4.99	6.40
K_2O	0.14	0.09	0.12
P_2O_5	0.08	0.03	0.05
L.o. ignit.	0.10	0.11	0.19
Total	100.01	100.48	99.75
Q	36.13	41.12	36.97
Or	0.83	0.53	0.71
Ab	47.98	42.22	54.16
An	7.40	11.72	3.42
C	0.00	0.00	0.00
Di	4.47	3.46	1.66
Wo	1.58	0.38	0.00
Hy	0.00	0.00	1.90
Mt	0.29	0.33	0.07
Il	1.06	0.53	0.57
Ap	0.19	0.07	0.16
Ca/Fe+Mg	0.97	0.71	0.16
Trace elements (ppm)			
Sr	339	443	79
Rb	1	1	1
Ba	0	0	0

Diopside gneiss	(121)
Diopside-actinolite gneiss	(178)
Actinolite schist	(123)

CHAPTER 7

COOKE HILL INTRUSIVES

7.1. INTRODUCTION

There are a number of small intrusive bodies mapped in the present area. They range in composition from tonalite* to granodiorite with the exception of some intrusives which are aplo-pegmatitic rocks. Generally, there are two distinct varieties of intrusive rocks; one type is weakly to moderatley deformed and the other type is massive in nature. Mineralogically and texturally, both are very similar except that their general appearance in colour and small-scale fabric makes them distinguishable from each other. Deformation features in one type and the lack in the other, clearly indicate that they have been intruded at two different times. Therefore, they have been classified in two groups based on their relative age:

1. ** Cooke Hill tonalite-granodiorite
2. Massive granodiorite

The Cooke Hill tonalite and granodiorite are indistinguishable in the field due to similarity in colour and textural properties. Petrographically, they can be divided into tonalite and granodiorite on the basis of their relative proportions of potash feldspar.

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* The term tonalite has been used here to express the composition of granite in which the amount of potash feldspar is less than 10% and hornblende is absent. This follows with slight modification, the igneous rock classification of Streckeisen (1967).

** The name Cooke Hill has been used here for the deformed intrusive rocks which occur around the Cooke Hill trig station, and elsewhere.

7.2. COOKE HILL TONALITE-GRANODIORITE

Occurrence and Field Relations

The Cooke Hill intrusive bodies appear in various shapes and sizes, from small plugs and dykes, a metre or so wide, to large lenticular bodies more than 800 metres long (Map 1). They occur as whale back out-crops in the quartzo-feldspathic schists and gneisses, generally covered by tor-like boulders (Plate 48A). The contact with the country rock may be sharp in the case of the big intrusive bodies. These bodies are generally conformable to the foliation of the gneisses but cross-cutting structures have developed due to injection of late residual liquids from the intrusives into the wall rock. Some of the small bodies have irregular shapes and their contacts with the country rock are discordant.

The forceful injection of magma is commonly suggested by distortion of bedding and foliation in the country rocks adjacent to the Cooke Hill bodies.

There is an abundance of xenoliths in many of the Cooke Hill intrusives (Plate 48B). The shape and size of xenoliths varies from angular fragments of metasediments to clots of biotite which are presumed to be highly basified and recrystallized small xenoliths. The internal structures (bedding, foliation etc.) of the xenoliths are randomly oriented.

The Cooke Hill bodies rarely have chilled margins, and there is little evidence of thermal metamorphism in the country rocks adjacent to them. At a few places, the coarsening of texture in schists and gneisses is apparent for up to 100-150 cm from the contact. Mineralogical changes which could be attributed to the intrusion of the Cooke Hill bodies are very uncommon. At one locality (Location 173072), the gneissic host rock has been altered to a creamy-white rock in a thin zone (about 20 cm wide) along its contact with a small tonalite dyke of the Cooke Hill suite. One sample (285/294c) of this altered rock is mainly composed of quartz and plagioclase, most of the biotite having been destroyed. The remaining biotites are highly altered to chlorite and the plagioclase is extensively sericitized with production of secondary white mica (Plate 49A).

7.2.1. Texture

Generally, there are two textural types of Cooke Hill intrusives, one medium grained and the other coarse grained, the latter one being much more common. Mineralogically, the coarse grained variety is always tonalitic in composition whereas the medium grained variety may be tonalite or granodiorite. In some bodies, a medium grained margin grades into a coarse grained core. But such features are limited to a few bigger intrusive where a narrow margin of about 3 to 6 m thick is seen rimming the thick core. Some marginal zones frequently show a porphyritic texture in which quartz phenocrysts are embedded into a medium grained matrix consisting of quartz-plagioclase-biotite minerals (e.g. Location 173062).

Although the general appearances and physical properties of the medium grained and coarse grained rocks may be slightly variable, the colour of the former rock is generally light and uniform while the coarser rock shows a greyish aspect due to an abundance of biotite. Most of the igneous rocks are weakly to moderately foliated and lineated as defined by preferred orientation of biotite. Occasionally, the marginal rock shows better lineation and/or foliation than the core rock. These observations lead to the conclusion that the Cooke Hill bodies were emplaced during an active phase of the metamorphism (cf. White, Compston & Kleeman, 1967).

7.2.2. Petrography

In thin section, the common minerals are quartz, plagioclase, potash feldspar and biotite. The proportion of potash feldspar is extremely variable ranging from nil to 20% of the total rock. Because of the variation in proportion of potash feldspar, the Cooke Hill igneous intrusives have been subdivided into tonalite and granodiorite.

The average grain size of the granodiorite is 1-2 mm, whereas that of the tonalite is 2-3 mm (Plate 50A,B). Generally, both have a hypidiomorphic texture, but in the strongly foliated and lineated rocks they have a granoblastic texture. If the degree of deformation is strong, the hypidiomorphic texture changes to a crystalloblastic one.

The most common mineral is plagioclase, which ranges from 50 to 70% in the tonalites, and from 30 to 60% in the granodiorites, whereas potash feldspar ranges from 0 to 10% in the tonalite, and from 10 to 20% in the granodiorite. Biotite comprises up to 15% of the tonalite and up to 10% of the granodiorite. Among the accessories, muscovite, apatite, sphene, zircon, monazite, opaques, tourmaline and epidote are the more common.

Plagioclase is euhedral to subhedral in shape, ranging in grain size from 0.5 mm to 5 mm. Twinning is well developed as closely spaced thin lamellae to well spaced broad lamellar planes. Most common types of twin laws are albite. Carlsbad-albite and pericline twinning is also seen. Some of the strongly deformed rocks show bending and fracturing of twin planes in plagioclase grains. Zoning in plagioclase is common and is present in both medium and coarse grained rocks of tonalitic to granodioritic composition. Normal and oscillatory zoning (rarely reverse zoning) may be found both in the small and the bigger grains (Plate 51A,B). The normal-zoned plagioclases have calcic cores with compositions as high as An_{60}, then gradually become more sodic towards their rims, where the composition may be as low as An_{20}. The composition of unzoned grains is An_{25-30} in the granodiorite to An_{30-35} in the tonalite. Some plagioclases contain irregular blebs of quartz and/or potash feldspar as inclusions. Sericitization and alteration of plagioclase with extensive clouding by fine dust, especially in the core of plagioclase is not uncommon. Secondary muscovite is found in the altered plagioclases, either as long plates along the cleavage planes or as fine grained white mica.

Quartz ranges from small ($<$ .1-2 mm) aggregates of matrix to large porphyroblastic grains up to 5 mm. The quartz grains are normally anhedral. Strong straining and fracturing of the larger grains is very common. Small inclusions of plagioclase are often found in quartz. Sometimes, wormlike quartz forms a myrmekitic structure at plagioclase-quartz boundaries, apparently a replacement phenomena as described by Shelley (1964, 1966, 1967, 1969, 1970) for igneous rocks.

Potash feldspar commonly shows cross-hatch twinning. It ranges from small grains to 5 mm megacrysts. Frequently, it forms embayments in plagioclase, which it appears to replace. Myrmekitic structures at the boundaries of plagioclase and potash feldspar are often seen. Exsolution lamellae are generally absent.

Biotite ranges in colour from brown to very dark brown. Flakes of biotite up to 2 mm long are common, but some much smaller biotites are often seen as inclusions in plagioclase and quartz. Alteration of biotite to chlorite has been observed in many thin sections. Biotite is frequently embayed by plagioclase.

Muscovite is not common as a primary mineral, and most of the white mica has been formed during alteration of plagioclase cores. Most of these white micas are usually very fine but some have grown to form larger flakes.

Epidote occurs as small grains either in the altered cores of plagioclase, or associated with biotite surrounding highly sericitized plagioclase grains. There is no indication that epidote is a primary mineral. Many petrologists believe that primary epidote is unstable in granitic magma (see Harpum, 1954; Shido, 1958). In the Cooke Hill intrusives its close association with altered plagioclase indicates that it has been formed at the cost of lime released by breakdown of plagioclase. The alteration of plagioclase may be due to the hydrothermal solution which has chloritized the biotite in the same rock. Sphene is common as subhedral to anhedral grains. Apatite is also a common accessory mineral in most of the rocks. Minute prismatic crystals of zircon are commonly found in quartz and plagioclase; in few instances subrounded zircons are also seen. Tourmaline and monazite are found in some thin sections. Iron ores are common in some thin sections and rare in others.

7.2.3. Discussion and conclusions

On the basis of field and petrographical observations, there is little doubt that the Cooke Hill intrusives are magmatic rocks.

The magmatic nature of the Cooke Hill intrusives is further supported by their hypidiomorphic texture and the abundance of normal and oscillation zoning in their plagioclases. Judging from the presence of the marked compositional zoning in plagioclase, the crystallization of magma has progressed more or less under non-equilibrium conditions (e.g. Hall, 1966). The presence of reverse zoning in the core of a few plagioclases may indicate that they are relict crystals (xenocrysts) from a melt (cf. Kato, 1968). The small amount of potash feldspar in most of the Cooke Hill intrusive bodies indicates that most of the potash of magma was consumed by early formed biotites and very little potash was left to form potash feldspar during crystallization (cf. Barth, 1952, p. 230).

The abundance of disoriented xenoliths of schists and gneisses near the margins of intrusive bodies, and the distortion of the country rock around the bodies strongly suggest the forceful injection of granitic magma. The form and shape of igneous bodies indicates that intrusion took place in a stress field during the later stage of tectonic deformation. The peculiar shape and isolation of individual satellite bodies with an abundance of small and large mappable enclaves appear to show that the injected magma squeezed into metasediments which were undergoing deformation at the time of their emplacement. It appears that magma was generated by differential anatexis in pockets of country rocks and rose upward during tectonic movements along the structurally defined weaker zones, viz. the bedding and schistosity planes etc. Mills (1964) has described a similar mechanism for intrusion of granodiorites in the Cambrai area; these are syn-tectonic intrusive bodies and appear to be related in time to the Cooke Hill intrusives.

The presence of weak to moderate foliation and lineation and absence of any contact effect with quartzo-feldspathic schists and gneisses indicates that intrusion took place at the waning stage of metamorphism when the temperature of the country rock was not significantly different from magmatic temperature. The Cooke Hill igneous intrusives are markedly different with respect to chemical and mineralogical compositions from the Palmer granite (White, 1967), the Murray Bridge granites (Kleeman, 1934; Worden, 1965), the Reedy Creek granites (Sando, 1957), the Mannum granites (Goode, 1927 ; Worden, 1965) and the Mannum tonalite (Alderman, 1927), but are similar to the Cambrai granodiorites (Mills, 1964). It is suggested that the Cooke Hill tonalites and granodiorites have been derived from a magma which was generated by anatexis of country rock at greater depth. This fact has been confirmed for the Cooke Hill intrusives by the major and trace element chemistry (see Chapter 8). Mills also has suggested the origin of the Cambrai granodiorites from an anatectic magma.

In summary, it is concluded that the Cooke Hill intrusions took place during the later stage of second phase deformation before the folding was completed (see section 2.6).

7.3. MASSIVE GRANODIORITES*

Occurrence and Field Relations

These granitic rocks generally outcrop in a small area of about 3 km^2 around the vicinity of the Rathjen Gap. There are about 30 small bodies appearing as stocks, dykes and lenses which have clearly discordant contact with the country rocks. Most of these bodies are small and cannot be shown on Map 1. These intrusive bdoies are small, rarely reaching a maximum length of about 150 m and range in thickness from 25 cm to as much as over 30 m.

A few disoriented xenoliths of country rocks are found in the Massive granodiorite. Such xenoliths are highly basified as indicated by an abundance of coarse biotite and low content of quartz and plagioclase. Pre-existing structure (foliation and schistosity) is retained in most of the xenoliths. There is no evidence of chilled margins or contact effects of intrusives with the country rocks.

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* The Massive granodiorite is a field term used to differentiate between the other acidic intrusive bodies in the Cooke Hill area (see Map 1). Petrographically, compositions of these rocks lie at the boundary of granodiorite and tonalite based on relative proportion of potash feldspar and plagioclase.

In hand specimen, they are dark grey, medium grained, dominantly massive rocks. The lack of preferred orientation of biotite in granites suggests that they are post-tectonic intrusive bodies. In some places, a weak lineation is visible in a narrow marginal part of some intrusive bodies, probably due to forceful injection with confining pressure of wall rocks (e.g. Chadwick, 1958). In the field they can be easily distinguished by their darker appearance and more massive texture in contrast to the Cooke Hill tonalite-granodiorite suite which is much coarser and lighter in colour. Tor-like boulders of the Massive Granodiorite are common, a feature which contrasts with the surrounding country rocks in which they seldom form whale back outcrops.

7.3.1. Petrography

The massive granodiorites have an average grain size of about 2 mm. The main minerals, quartz, plagioclase and biotite, with lesser potash feldspar, range in grain size from a fraction of a millimetre to a few millimetres. In thin section, the rocks display a hypidiomorphic texture in which the lath-shaped plagioclases are plentiful (Plate 52A). Although quartz grains exhibit wavy extinction there is little evidence of post crystalline deformation, and a preferred orientation of biotite is absent except a few marginal rocks whose biotite defines a weak foliation.

Plagioclase is the dominant mineral and constitutes about two-thirds of the total rock. Almost all plagioclase shows albite twinning, Carlsbad, pericline and Carlsbad/albite twin laws may also be seen in some thin sections. Zoning is very common and both normal and oscillation zoned plagioclase are present (Plate 52B). Generally, the thick sodic rims have anhedral outlines, but they surround euhedral cores. The composition in zoned crystals ranges from An_{50} in the cores to An_{22} at the rims.

Quartz is the next most abundant mineral after plagioclase. It is subhedral to anhedral, and is extremely variable in grain size. The smallest grains are minute drop-like inclusions; the largest may reach 3 mm in diameter. Graphic intergrowths with plagioclase are not uncommon.

Potash feldspar is less common, but in a few thin sections it makes up about 20% of the total rock. Some grains have cross-hatch twinning. Biotite (X = yellowish brown, Y = Z = dark brown) is the only mafic mineral present in the rock. It commonly alters to chlorite. Small inclusions of biotite are often found in plagioclase and quartz. Muscovite as a primary mineral is less common than in the Cooke Hill intrusives. In some thin sections long plates of muscovite occur in the cores of plagioclase grains. Epidote is a colourless to pale yellow, weakly pleochroic secondary mineral, usually associated with "sericitized" plagioclase. Iron ores are subhedral and may range from minute grains up to 1 mm in diameter. Euhedral prismatic zircon crystals are common. Other accessory minerals are apatite, sphene, tourmaline and monazite.

7.3.2. Conclusions

The magma appears to have been emplaced during the last (brittle) phase of deformation, as suggested by elongation of bodies parallel to the axis of a major F_2 synform.

The intrusion of magma was forceful as indicated by warping and shouldering of country rock structures near the boundary and abundance of disoriented xenoliths in the intrusive bodies. They also show cross-cutting relationships with the country rock as the injection of the late residual fluids in the adjacent quartzo-feldspathic rocks. Furthermore, the magmatic nature of these rocks is exemplified by hypidiomorphic texture, and by normal and oscillation zoning in plagioclase (e.g. Leedal, 1952; Phillips, 1956). Euhedral shaped zircons are common. This is considered to be an indication of magmatic history for granitic rocks (Poldervaart, 1950, 1956; Larsen & Poldervaart, 1957; Mills, 1964).

The Massive granodiorite intrusives constitute a suite of rocks that are different from the Cooke Hill suite of granodiorites and tonalites. This is evident from the difference in physical properties, such as colour and texture, the difference in mineralogy and the complete geographical isolation of the two types of intrusives from each other.

The genesis of the Massive granodiorites has been discussed in the light of major and trace element chemistry for one selected sample (208), which is presented in Chapter 8. The K/Rb ratio of total rock is 219 which is close to the K/Rb ratios of acid crustal rock 230-240 as compared to differentiated magmatic granites 175 (Heier & Adams, 1964). Therefore, it is considered that the Massive granodiorites were formed by an anatectic magma which resulted by melting of country rock at greater depth.

7.4. APLITE-PEGMATITE INTRUSIVES

Introduction

There are hundreds of pegmatitic bodies occurring in many shapes and sizes throughout the whole area. The pegmatites and also aplites tend to concentrate around the main plutonic masses of the Cooke Hill tonalite-granodiorite and the Massive granodiorite, some transecting such plutons or segregated as marginal facies. Most commonly, they occur as thin stringers, lenses, pods and tabular masses ranging in thickness from 10 cm to more than 30 m. They also vary in their attitude and commonly occur along structural features of the country rocks such as pre-existing joints and cracks, bedding and schistosity planes. Some pegmatites and aplites carry xenoliths of the country rocks. Pegmatites are much more common than the aplites. Fine grained aplitic bodies are generally associated with the tonalites and granodiorites either as marginal zones or discordant veins. Occasionally, composite layered bodies have been seen having a core of aplite rimmed by pegmatite or vice versa (c.f. Jahns & Tuttle, 1963).

These pegmatites and aplites are related to the main plutonic masses of the tonalites and granodiorites. Some pegmatites are quite remote from the main intrusive bodies. A remarkable concentration of such pegmatites can be noted cutting the Strangway Hill beds in the south-eastern part of the area (see Map 1). They display intrusive characters, i.e. cross-cutting structures, discordant occurrences and the presence of xenoliths etc., similar to those displayed by the pegmatites closely associated with the large intrusive bodies.

Texturally, there are two types of pegmatites; one is medium grained (average grain size 1-2 mm) and the other is coarse to very coarse grained (average grain size 3-4 mm) with some of the megacrysts of alkali feldspar reaching 3-4 inches in length. The coarse grained variety is more common. On a small scale, there is evidence of deformation in the form of a weak to moderate foliation and lineation in some pegmatites. These are defined by a preferred orientation of mica and elongated quartz grains. Such deformation features may also be found near the margin of the pegmatitic bodies whose cores are generally massive. This wall rock foliation and lineation in pegmatites has also been reported by Mills (1964) in the Cambrai area and Offler (1966) in the Pewsey Vale area.

It is interesting to note that pegmatites are quite common in the quartzo-feldspathic schists and gneisses, marbles and calc-silicate rocks but are less common in calc-schists and calc-gneisses. There are some pegmatites in the calc-schists and gneisses, but they are of different mineralogical composition from the more common types. Like other intrusive pegmatites, they also show transgressive relationships but generally lack xenoliths. The rareity of pegmatites in the calc-schists and the calc-gneisses is possibly due to the incompetent nature of these rocks.

Mineralogically, the pegmatites are extremely variable in their constituent minerals. Quartz, plagioclase and microcline are the major components. Some pegmatites are extremely rich in plagioclase with potash feldspar as a minor constituent. However, the most common pegmatites are rich in microcline; they may contain small to significant amounts of plagioclase. Pegmatites which contain amphiboles are always impoverished in potash feldspar and are most common in the calc-schists and the calc-gneisses.

7.4.1. Petrography

The petrographic features of aplites and pegmatites are described below.

APLITES

Generally, the aplites are medium grained with a 1-2 mm average grain size. They consist of quartz, potash feldspar and plagioclase, together with a little biotite and muscovite. Hypidiomorphic texture is well developed even in moderately foliated rocks.

Quartz occurs as discrete anhedral grains, in which undulose extinction and fracturing is common. Small inclusions of quartz in plagioclase and potash feldspar are frequently seen.

Plagioclase is characterized by well developed twinning; albite, Carlsbad, pericline and Carlsbad/albite are the most common twin laws. Weak normal zoning is frequently present. The composition of plagioclase ranges from An_{10-21}.

Potash feldspar occurs as non-perthitic grains, usually with well developed tartan twinning. Myrmekitic structure is common; worm-like quartz grows inside microcline near the boundary with quartz.

Biotite (X = pale brown, Y = Z = deep brown) occurs as small flakes. Pleochroic halos are rare.

Muscovite is a minor constituent as a primary mineral.

Secondary epidote, opaques, apatite, zircon and sphene are the common accessories.

PEGMATITES

The pegmatites have an average grain size ranging from 0.5 mm to 5 mm. Plagioclase is generally well twinned, occasionally zoned, and ranges in composition from An_{3-21}. Replacement of plagioclase by quartz is common. It is generally fresh but may be extremely dusty by incipient alteration. Microcline is characterized by well developed cross-hatch twinning. Vein perthite occasionally develops. Myrmekite may be common in plagioclase and at microcline-plagioclase contacts.

Quartz is generally anhedral in shape and commonly shows undulose extinction and fracturing. Graphic or vermicular intergrowths of quartz in microcline are common in some pegmatites. Muscovite is generally less common and may occur as small flakes.

Biotite is also a minor constituent and occurs as pale brown to dark brown flakes. Amphibole is absent in most pegmatites, but is a common mineral in some. It is usually weakly pleochroic, light green to olive green. It appears to be one of the first minerals to crystallize and is extensively replaced by later quartz. Epidote is common as an alteration product. Sphene, apatite, tourmaline, zircon and pyrites are other common accessories.

7.4.2. Conclusions

Most of these pegmatites and aplites are genetically related to two magmatic events which gave rise to the Cooke Hill tonalite-granodiorite suite and the Massive granodiorite suite as a residual phase of the parent magmas. The forceful injection of pegmatites and aplites is evidenced by the presence of xenoliths of the country rocks. The emplacement of magmatic liquids took place along the pre-existing joints, cracks, bedding and schistosity planes of the country rocks, which were widened by forceful intrusions. The occurrence of the wall rock foliation and lineation in the marginal part of some pegmatites is considered by Chadwick (1958) to indicate their forceful injection.

The rarity of the aplites and the presence of plagioclase in most of the pegmatites suggest that the crystallization of residual liquid took place under high water pressure conditions (e.g. Jahns & Burnham, 1958, 1969; Jahns & Tuttle, 1963).

7.5. QUARTZ VEINS

There are a number of large and small quartz veins occurring as thin veins filling the fracture planes in the country rock or as variously oriented reefs up to 100 m long. Such quartz reefs are particularly common around the Cooke Hill trig station.

In hand specimen, the quartz veins may be whitish or pinkish in colour, and are essentially composed of anhedral quartz. Rarely, large creamy-white potash feldspar and euhedral cubes of pyrite may also be seen. The veins commonly cross-cut bedding and foliation, some filling a-c joints in the country rocks. They are generally distinguishable from the quartz veins which have formed parallel to bedding and schistosity. Possibly, most of the cross-cutting veins are related to late stage magmatic fluids, and the majority of the conformable veins have arisen by metamorphic segregations.

7.6. FOLIATED PEGMATITES

. Introduction

These are strongly deformed, lineated and foliated rocks of pegmatitic appearance. Generally, they are conformable with the country rock and range in thickness from 10 cm up to 100 m. In hand specimen, they are pink to flesh coloured, medium grained rocks with prominent planar and linear structures defined by orientation of mica. In some cases a lineation is defined by elongated quartz. These structures are parallel to similar structures in the country rock. They do not display cross-cutting structures with the country rock and xenoliths are absent. In rare cases, they may grade lithologically through a narrow zone into the adjacent schists and gneisses.

7.6.1. Petrography

The most common minerals are quartz and plagioclase with small amounts of potash feldspar, biotite and/or muscovite (Plate 53A). The grain size ranges from 0.5-2 mm.

In thin section, a granoblastic texture is commonly evident. In many cases, xenoblasts of quartz and plagioclase form a porphyritic texture.

Quartz is generally subrounded to amoeboid in shape with marked undulose extinction. Fracturing and deformation lamellae are common in bigger grains. Quartz often replaces plagioclase and potash feldspar.

Plagioclase is the most abundant mineral, and has well-developed twinning. Bending and fracturing of twin planes are often seen. The composition of plagioclase ranges from An_{4-14}. Usually, it is fresh but may be slightly altered to fine grained sericite or white mica. It is the next abundant mineral after quartz. Potash feldspar is characterized by its typical cross-hatch twinning but may be untwinned in some thin sections. It is generally fresh, but some altered grains have a dusty appearance due to the formation of fine grained white mica. Potash feldspar is variable in proportion from less than 5% to 15% of the total rock. Biotite is pale brown to dark brown in colour with small flakes .3-.5 mm in length. These have a planar preferred orientation, defining a schistosity. Biotite never exceeds more than 5% of the total rock. Muscovite is less common as a primary mineral. Small flakes of muscovite are often associated with biotite. Secondary muscovite is usually found in the altered plagioclase. The common accessories are irregular shaped iron ores, subhedral sphene, tourmaline zircon and apatite; chlorite forms as an alteration product of biotite. Epidote is often found as small granules in altered plagioclase.

The field evidence does not support that the foliated pegmatites are of igneous origin. Their general concordance with the country rocks and parallelism of internal structures both in pegmatites and adjacent host rock, suggests a metamorphic origin; some metasomatic activity may also have been involved in the genesis of the foliated pegmatites.

7.7. METADOLERITES AND AMPHIBOLITES

Introduction

There are a large number of metadolerites and amphibolites occurring throughout the area. Most of them are small bodies - dykes, lenses and pods ranging in diameter from 30 cm to more than 5 m; some may reach over 30 m. They appear to have intruded during the regional metamorphism and tectonic deformation; therefore, most of them are aligned parallel to the axial trend of major structures and foliation of the metasediments. They have suffered a great degree of recrystallization and reconstitution of the original igneous texture. The majority of the metadolerites and amphibolites are strongly lineated and schistose rocks with preferred orientation of amphiboles visible by naked eye. Generally, contact effects and xenoliths are rare, but some reaction zones have been noticed where dolerites are in contact with the marbles and calc-silicate rocks. In hand specimen, they are fine to medium grained schistose rocks with prominent phenocrysts or porphyroblasts of amphibole and plagioclase embedded in the fine grained groundmass. In the field, they are characterized by their dark green and black colour with strong planar and linear structure. The degree of recrystallization during regional metamorphism varies depending upon the size of intrusives; the thick lenses and dykes show strong lineation and schistosity at the margin but the central part is less recrystallized and weakly deformed. Similar observations have also been reported by White (1956), Mills (1964) and Offler (1966) in the metamorphosed dolerites of the Mt. Lofty Ranges.

7.7.1. Petrography

The major constituents of the metadolerites and the amphibolites are amphibole and plagioclase with minor amounts of biotite and scapolite. The common accessories are iron ores, sphene and apatite. Epidote, sericite and white mica are common as alteration products of plagioclase. Potash feldspar and tourmaline are rare.

In thin section, the degree of recrystallization ranges from a subophitic texture to a completely recrystallized granoblastic texture (Plate 53B). Some of the phenocrysts of plagioclase retain their original igneous habit even in a completely recrystallized groundmass composed of amphibole and plagioclase. The strong deformation feature is indicated by bending and fracturing of plagioclase twins (and occasional elongation of plagioclase along the schistosity plane).

Amphibole is pleochroic X = pale olive green, Y = olive green or bluish green, Z = deep green. The average grain size is 2-2.5 mm but some may reach up to 5 mm as a recrystallized granoblast. Small inclusions of iron ores and plagioclase are commonly seen. It forms about 40-60% of the total rock.

Plagioclase occurs either as elongated phenocrysts up to 4 mm in length or small aggregates of groundmass. Twinned and untwinned crystals are equally common. Normal zoning is well developed in some grains; occasionally, a weak oscillatory zoning may also be seen in the phenocrysts. Alteration occurs in variable degrees, starting as a general clouding and sericitization leading to the formation of small granules of epidote in some cases. The common range of composition is An_{30-60}.

Biotite is found in most of the rocks in amounts varying from less than 1% to as much as 10% of the total rock. The common biotites are X = pale yellow, Y = Z = golden brown but some are light brown (X) to chocolate brown (Y-Z). Biotite is closely associated with amphibole.

Scapolite is a less common mineral. It generally occurs as replacement of plagioclase especially in those metadolerites which are intruded in calc-silicate rocks. Most of the metadolerites which occur in quartzo-feldspathic rocks are devoid of scapolite. Mills (1964) has recorded similar observations and concludes that the genesis of scapolite is related to availability of chlorine and other volatiles which were lacking in the original basic magma but were locally derived from the adjacent country rocks mainly from the calc-silicate rocks.

Iron ores occur as skeletal or irregular shaped aggregates, often aligned along the schistosity plane. In most rocks it is common and may make up about 5% of the total rock.

Sphene is always present as an accessory mineral. Generally, it occurs as subhedral to anhedral grains in aggregates associated with amphibole.

Apatite is also a common accessory mineral. Potash feldspar is rare in most of the rocks. Euhedral short prisms of zircon are frequently seen in plagioclase.

Epidote occurs as small granules with close association of sericite mica in altered plagioclase. It may be found as a weakly pleochroic (colourless to pale yellowish green) mineral or less commonly, as an iron free or iron poor clinozoisite mineral. There are epidote rich veins intruded along the fracture or joint planes in the metadolerites. These are thin veins of a few centimetres wide and mainly composed of weakly pleochroic epidote with small amounts of plagioclase. The field evidence indicates that they have probably resulted as a late stage residual fluid of basic magma. Mills (1964) has also cited several examples of such veins in the metadolerites from the Cambrai area.

7.7.2. Conclusions

The metadolerites and amphibolites are considered to be the igneous rocks which were intruded during the regional metamorphism. The term amphibolite is used here for those igneous basic rocks which have more than 50% amphibole mineral. Although both metadolerite and amphibolite are completely recrystallized rocks, there is no evidence of original augite being converted to hornblende. Generally, relict igneous texture has been completely destroyed to give a granoblastic metamorphic texture but some relict ophitic texture is indicated by lath-shaped phenocrysts of plagioclase.

In the field, most of them are concordant with the country rock but the blasto-porphyritic texture is very good evidence of original igneous character of metadolerite. A notable feature of the metadolerites and amphibolites of the present area is the absence of garnet. This is in contrast to metadolerites described by Wiseman (1934), Wilcox and Poldervaart (1958) and Mason (1962). This absence of garnets in metadolerites of the Cooke Hill area (see also White, 1956; Offler, 1966) is attributed to pressure conditions prevailing during the original metamorphism. The stability of sillimanite as the only alumino-silicate polymorph in pelitic rocks of the area studied is indicative of low pressure intermediate type metamorphism (Miyashiro, 1958, 1961). These facts suggest that the pressure conditions prevailing during the regional metamorphism were not suitable for crystallization of garnet in the metadolerites and amphibolites.

CHAPTER 8

PETROCHEMISTRY OF THE COOKE HILL GRANITIC INTRUSIVES

8.1. INTRODUCTION

Eight samples of the Cooke Hill intrusives were analysed for the major elements. Of these, six can be classified as tonalites and two as granodiorites. One sample (208) of "the Massive granodiorite" from north of the Rathjen Gap was also analysed (Table 11). Trace element determinations were made for Rb, Sr and Ba in five samples of the Cooke Hill tonalites, two samples of the Cooke Hill granodiorites and one sample of the Massive granodiorite (Table 12).

Biotites from each of the above Cooke Hill intrusives were also analysed both for major and trace elements (Tables 5B & 13).

In addition, two xenoliths from the Cooke Hill tonalites and the biotites separated from them were also analysed (Table 14).

8.1.1. Major elements

All six tonalite analyses are chemically similar, with slight variation in silica and alumina contents. When, as in Table 16, the average of 6 analyses is compared with average of the biotite tonalites (Nockolds, 1954), it is found that they are similar except for higher alumina and soda and lower total iron contents in the Cooke Hill tonalites. C.I.P.W. norms of the tonalites are close to average biotite tonalites of Nockolds (1954) except for some variation in Q and Ab values due to the variation of alumina and soda. Likewise, the Cooke Hill granodiorite can be matched with muscovite-biotite granodiorite of Nockolds (1954).

The Cooke Hill granodiorite has higher potash and lower calcium contents than the Cooke Hill tonalites. A plot of Na_2O and K_2O shows two distinctly separate fields of tonalites and granodiorites (Fig. 30). There is some variation in MgO and FeO (total) and in Fe (total) and Ti of the two rock types which is shown in Figures 31 and 32 respectively. The other major elements do not show any significant differences.

The genesis of tonalites and granodiorites has been considered in terms of the Ab-Or-SiO_2-H_2O system (Tuttle & Bowen, 1958; Luth et al., 1964), and the Or-Ab-An-SiO_2 system (Kleeman, 1965). The composition of magma is generally expressed in terms of C.I.P.W. normative quartz, orthoclase, albite and anorthite, and for that reason the C.I.P.W. norms have been used for the Cooke Hill magmas.

The normative Q, Or and Ab for the Cooke Hill intrusives are plotted in Figure 33. The normative composition of these granitic rocks shows they are relatively rich in Ab and consequently they fall close to the Q-Ab sideline and some distance away from the minimal melting compositions of the "synthetic granite" system Ab-Or-SiO_2 of Tuttle and Bowen (1958) and Luth et al. (1964). When the normatives Or, Ab and An are plotted in Figure 34, all the intrusives lie away from the 5000 bar low temperature trough of Kleeman (1965). It is interesting to note that the field of composition of the Cooke Hill tonalite and granodiorite shows great similarity to granitic rocks of Norway (Barth, 1955, Fig. 4; 1966, Fig. 30D), synkinematic granitic rocks of Finland (Eskola, 1956, Fig. 1) and autochthonous granitic rocks of Lewisian (Bowes, 1967, Figs. 2-3).

8.1.2. Trace elements

Trace element determinations were made for Rb, Sr and Ba using X-ray fluorescence methods and are presented in Table 12. The granodiorites are richer in Ba and Rb (as well as K) than the tonalites; Sr does not show any significant difference but Ca is low in the granodiorites. The average ratios of K/Rb are constant in all the rocks, but Ba/K ratios are variable in the tonalites; Sr/Ca and Ba/Rb ratios are generally higher in the granodiorites than the tonalites, The Massive granodiorite has lower absolute abundances of Rb, Ba and Sr than the Cooke Hill tonalites and granodiorites, and the K/Rb, Sr/Ca, Ba/Rb and Ba/K ratios are generally lower.

8.1.3. Chemistry of biotites

The major element chemistry of biotites from the two granitic rock types does not show any significant differences between the samples (Table 5B). Also, the Fe/Fe+Mg ratios of biotites (Table 6) show no characteristic differences in the two rock types indicating

TABLE 11

Chemical analyses and C.I.P.W. norms of the Cooke Hill intrusives

Sample No.	74	140	199	233	295	511	137	232	208
SiO_2	69.62	66.65	68.71	68.71	69.67	68.83	69.82	71.31	68.59
TiO_2	0.34	0.43	0.25	0.37	0.36	0.37	0.26	0.26	0.40
Al_2O_3	16.45	17.15	17.23	16.19	15.81	16.29	16.08	16.01	15.49
Fe_2O_3	0.52	0.59	0.44	0.35	0.38	0.54	0.44	0.52	0.98
FeO	1.73	2.02	1.72	1.89	1.53	1.93	1.34	1.20	2.11
Mno	0.09	0.09	0.04	0.14	0.10	0.10	0.06	0.08	0.08
MgO	1.12	1.26	0.96	0.96	1.23	0.97	0.78	0.68	1.06
CaO	01	3.02	2.94	3.05	2.75	3.12	1.97	1.81	3.66
Na_2O	5.52	5.93	5.47	5.52	5.32	4.69	5.22	4.11	4.[illegible]
K_2O	1.20	1.57	1.71	1.40	1.73	2.10	3.49	3.44	1.08
P_2O_5	0.18	0.13	0.10	0.18	0.08	0.20	0.11	0.15	0.16
L.o. Ignit.	0.42	0.63	0.77	0.75	0.57	0.40	0.51	0.47	0.76
Total	100.20	99.47	100.34	99.51	99.53	99.54	100.08	100.04	99.30
Q	24.05	16.80	21.59	22.27	23.57	24.59	19.97	28.42	25.63
Or	7.09	9.28	10.11	8.27	10.22	12.41	20.63	20.33	6.38
Ab	46.71	50.18	46.29	46.71	45.02	39.69	44.17	34.78	41.72
An	14.18	14.47	14.38	14.48	13.32	14.48	9.62	9.55	16.95
C	0.87	0.39	1.11	0.29	0.30	1.00	0.19	2.03	0.00
Di	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.31
Hy	5.14	5.82	4.85	5.22	5.15	5.09	3.72	3.19	5.04
Mt	0.75	0.86	0.64	0.51	0.55	0.78	0.64	0.75	1.42
Il	0.65	0.82	0.47	0.70	0.68	0.7[illegible]	0.49	0.49	0.76
Ap	0.42	0.30	0.19	0.42	0.16	0.46	0.26	0.36	0.37

TABLE 12

Major and trace elements in the Cooke Hill intrusives

Sample No.	7	140	199	233	295	137	232	208
Na%	4.09	4.40	4.06	4.09	3.95	3.87	3.05	3.66
K%	1.00	1.27	1.42	1.16	1.44	2.90	2.85	0.90
Ca%	2.15	2.12	2.10	2.18	1.96	1.41	1.29	2.61
Rb ppm	35	35	52	49	46	93	100	41
Sr ppm	664	699	542	653	551	665	647	342
Ba ppm	1198	675	588	1759	785	1906	1667	565
K/Rb	286	363	275	237	313	312	285	219
Sr/Ca.10^3	31	33	26	30	28	47	50	13
Ba/Sr	1.80	0.96	1.08	2.69	1.42	2.87	2.58	1.65
Ba/Rb	34	19	11	36	17	20	17	14
Ba/K,10^3	119	53	41	152	54	66	58	63
Na/K	4.09	3.56	2.85	3.52	2.74	1.33	1.07	4.06

Cooke Hill tonalites	(75, 140, 199, 233, 295)
Cooke Hill granodiorites	(137, 232)
Massive granodiorite	(208)

TABLE 13

Trace elements in biotite

Sample No.	75	199	233	140	137	232	208	142	294-A
Sr ppm	12	9	8	6	10	5	9	2	2
Rb ppm	331	488	367	364	533	719	375	443	532
Ba ppm	3120	2408	3197	3911	1348	1818	2200	843	1878
K/Rb	223	155	201	220	147	104	181	183	150
Ba/Rb	9.4	4.90	8.70	10.7	2.52	2.52	5.90	1.90	3.40
$Ba/K.10^3$	41.2	31.7	43.4	48.9	17.2	24.2	32.4	10.5	23.5

Cooke Hill tonalites	(75, 199, 233, 140)
Cooke Hill granodiorites	(137, 232)
Massive granodiorite	(208)
Xenoliths	(142, 294-A)

equilibrium conditions with respect to oxygen fugacity and temperature at the time of their crystallizations (cf. Wones & Eugster, 1965).

The biotites of the granodiorite have mean values of 626 ppm Rb and 1583 ppm Ba, whilst the biotites of the tonalite have 390 ppm Rb and 3159 ppm Ba. Similarly the biotites of the granodiorite have mean values for K/Rb, Ba/Rb and Ba/K of 125, 2.52 and 20.69 compared with the same ratios in the biotites and the tonalite of 200, 8.45, and 41.53.

The behaviour of trace elements in biotite of the Cooke Hill intrusives clearly follows the magmatic trend with a relative abundance of Ba in the biotites of the high temperature tonalite and a relative abundance of Rb in the biotites of the lower temperature granodiorite as is shown by systematic decrease of K/Rb, Ba/Rb and Ba/K ratios from basic granite (tonalite) to acid granite (granodiorite).

The distribution of trace elements in biotite of the Massive granodiorite shows a more or less similar trend to the biotite of the Cooke Hill tonalite.

8.1.4. Chemistry of xenoliths

Two xenoliths from the Cooke Hill tonalites have been analysed both for total rock and biotite (Table 14). These samples are highly basified by enrichment of mica and low content of quartz and feldspar. The basic nature of xenoliths is shown in the chemical analyses by low silica and high total iron. On the basis of major element chemistry, it is suggested that the composition of the xenoliths has been changed by the reaction of granitic magma. The abundance of mica in the xenoliths is related to potassium metasomatism with exchange of some materials as an addition or removal between the magma and the original xenoliths (Chao, 1951; Goodsp 1948; Eskola, 1956; Reynolds, 1956; Hietanen, 1954). Despite this, some of the primary structures, such as relict bedding and schistosity are not greatly changed, and resemblance to country rocks is apparent.

The biotites of the xenoliths have the same chemical composition as the biotites from granitic rocks. The trace element data for

TABLE 14

Chemical analysis of xenoliths and biotites

Sample No.	Total rocks 142	Total rocks 294-A	Biotites 142	Biotites 294-A
SiO_2	58.52	58.05	35.55	34.30
TiO_2	0.95	1.32	2.33	2.95
Al_2O_3	18.13	15.72	17.70	16.91
Fe_2O_3	4.20	3.26	4.02	4.35
FeO	5.51	7.19	17.93	17.07
MnO	0.12	0.19	0.20	0.23
MgO	2.01	2.03	10.75	10.06
CaO	0.22	0.96	0.07	0.00
Na_2O	0.31	2.05	0.14	0.11
K_2O	6.87	6.76	9.76	9.63
P_2O_5	0.19	0.34	0.00	0.00
L.O. Ignit.	2.28	1.59	1.60	4.97
Total	98.98	100.26	100.05	100.58

	Barth mesonorm			Structural formula on anhydrous basis of O = 22	
Q	30.97	18.21	Si	5.28	5.25
Or	4.77	18.78	Al	2.72	2.75
Ab	2.97	19.21	Z	8.00	8.00
An	0.05	4.17	Al	0.38	0.30
Bi	14.25	17.99	Fe^{+3}	0.45	0.50
Ms	41.44	14.33	Ti	0.26	0.34
Mt	4.68	3.56	Mg	2.38	2.30
Il	1.41	1.92	Fe^{+2}	2.23	2.19
Ap	0.42	0.74	Mn	0.02	0.03
			Y	5.72	5.66
			Ca	0.01	0.00
			Na	0.04	0.03
			K	1.85	1.88
			X*	1.91	1.93

biotites is shown on Table 13. Rb concentration in biotite of the xenolith is the same as that in biotite from the granitic rocks, but absolute abundance of Ba is variable, ranging from 843 to 1878 ppm, much lower than the average of Ba in tonalites. The K/Rb ratio is similar, but the Ba/Rb and the Ba/K ratios are lower than in the biotites of the tonalites. On the basis of trace element data for biotites, it may be concluded that a complete equilibrium was not achieved between the xenoliths and the tonalite magma.

8.2. DISCUSSION

From the field evidence alone, the granitic rocks of the Cooke Hill area have all the signs of having been intruded into their present position. The presence of numerous dis-oriented xenoliths of country rock in the granitic bodies strengthens the argument for magmatic origin. Most of the granitic bodies have conformable contacts with the country rocks indicating that magma chose the easiest way of intrusion parallel to the pre-existing structure, bedding or foliation planes and had enough force to break locally the structures of the country rocks during magmatic intrusions.

The structural data demonstrate that the Cooke Hill tonalite granodiorite suite was emplaced as late kinematic intrusives earlier than the emplacement of the undeformed Massive granodiorite. Thus the granitic intrusions in the present area took place in two different periods, one in the stress field of tectonic deformation and the other after the cessation of ductile deformation. The first phase of igneous activity started towards the end of second deformation period and resulted in the emplacement of numerous small to medium size bodies of tonalite to granodiorite composition during the waning stage of metamorphism. Therefore, most of these intrusives are mildly deformed and have weak to moderately developed linear and planar structures. Since both tonalites and granodiorite intrusives have a similar degree of deformation and contain numerous inclusions and xenoliths of earlier deformed country rocks, it thus appears that two magmas of slightly different composition (tonalite and granodiorite) were emplaced during the same period of intrusion.

A characteristic feature of these intrusives is that they tend to concentrate in the hinge zone of the major F_2 synform and are elongate parallel to axial trace of synform. The presence of rather weak linear and planar structures in the intrusives compared to strongly deformed country rocks indicate that the Cooke Hill tonalite-granodiorite magmas crystallised in dying stage of deformation when the stress field was minimal. At the end of second ductile deformation, the second period of igneous activity produced the undeformed "Massive Granodiorite". The massive nature and presence of xenoliths of country rocks having F_1 and F_2 structures suggests that these intrusives are younger than the deformed Cooke Hill tonalite-granodiorite suite.

Thus on field and structural relationships, granitic rocks of the present area have been divided into two groups, the older rocks as a late kinematic tonalite-granodiorite suite and the younger as the post-kinematic massive granodiorite. The variations in chemical composition of the these rocks and distribution of trace elements within each rock series was studied to resolve the problem of origin of the magmas which gave rise to these rocks and to resolve the question whether the granitic rocks of late-to-post kinematic Cooke Hill suite crystallized from one parent magma or were two separate magmas which formed at different places and times.

The tonalites of the Cooke Hill suite differ from the granodiorites both in mineralogical and chemical composition. The tonalites contain less potassium and more calcium and thus have less potash feldspar and more plagioclase (Fig. 35). Biotite is generally more abundant in tonalites than in the granodiorites as is also evident by the plots of Mg, Fe and Ti ratios of these rocks (see Figs. 31, 32). Despite a limited composition range of major elements in the analysed tonalite and granodiorite rocks (Figs. 33, 34), they display some variation in their trace elements (Figs. 27 to 40).

The variations of trace elements in magmatic rocks have been discussed by other workers (for example see Nockolds & Allen, 1953; Kolbe & Taylor, 1966; Hietanen, 1963; Hall 1967; Han-Weinheimer & Ackermann, 1967 and Stephenson, 1972 etc.). The observed variations of trace elements-Rb, Ba and Sr in the Cooke Hill tonalites and granodiorites appear to be related to variable mineral proportions of potash feldspar, mica and plagioclase in the rocks comprising these rock series. The absolute abundance of trace elements is slightly variable within the rocks of tonalite and granodiorite types but the average values of Rb and Ba are fairly distinct for both rock types. The average amounts of Rb and Ba in the Cooke Hill tonalite are 43 ppm and 1001 ppm respectively while those in the granodiorite are much higher at 96 ppm and 1786 ppm respectively. The average Sr values in two rock series, however, do not show any characteristic difference being 622 ppm in tonalite and 656 ppm in granodiorite. The Massive granodiorite has fairly distinct trace element data with Sr 342 ppm, Ba 565 ppm and Rb 41 ppm rocks.

If the Cooke Hill igneous suite are the products of fractional cyystallisation of a parent magma, these rocks would be expected to show decreasing values of the K/Rb, Ba/Rb and Ba/Sr ratios from tonalite to granodiorite (cf. Butler, et al. 1962; Heier and Adams, 1964; Butler, 1964; Taylor, 1965 and Tauson, 1965 etc.). Turikian and Kulp (1956) have found that Sr/Ca ratio decreases with increasing fractionation of granitic magma. It is an established fact that during the progressive differentiation and crystallisation of magmas the Rb and K become enriched in the late stage fractionated acid granites as compared to earlier fractionated "basic granites". In the analysed rocks of the Cooke Hill area, the absolute abundance of Rb and K is higher in the granodiorite than in the tonalite but the K/Rb ratio which should be lower in the late fractionated granodiorite, has an almost identical value to that of the earlier formed tonalite. The average K/Rb ratios of the Cooke Hill intrusives are 295 for tonalite and 299 for granodiorite. This reverse trend found in the granitic rocks of the present area does not uphold the idea of progressive differentiation and fractionation processes in these magmas. The

Ba/Rb ratio is a very good indicator of fractional crystallization of magma (Taylor and Heier, 1960) because Ba concentrates in the early formed potassium minerals and Rb in the later minerals. Therefore, the Ba/Rb ratio must show a marked decrease with progressive fractionation of magma. Although the Ba/Rb ratios of the Cooke Hill "granites" are more variable due to variable amounts of potash feldspar and biotite present in these rocks some of the analysed tonalites have lower Ba/Rb ratios than the granodiorites, again showing the reverse of the differentiation trend. The Sr/Ca ratio which shows marked decrease with increasing fractionation has the reverse order in the Cooke Hill intrusives, having the average Sr/Ca ratio of 30 in tonalite and 49 in granodiorite. Similarly, the Ba/Sr ratio falls with fractionation (Heier and Taylor, 1959). In the Cooke Hill Suite the average for this ratio is higher in the granodiorite (2.7) than in the tonalite (1.6). This aberrant behaviour of trace elements in the Cooke Hill tonalite-granodiorite suite does not support a simple thesis of magmatic differentiation as a major cause in the evolution of these magmatic rocks.

In the light of the above discussion, the trace element distribution in the Cooke Hill tonalite-granodiorite suite shows that they have not resulted by fractional crystallization of single magma. The present variation of trace elements could be suggested as evidence in favour of anatectic origin. The possibility for the evolution of "granitic" magmas of varying composition by anatexis has been investigated by comparing the compositions of the Cooke Hill granitic rocks with the experimental data of Tuttle & Bowen (1958) and Luth et al. (1964) in the Ab-Or-Q-H_2O system. Apart from some scatter, it is evident that the majority of the Cooke Hill granitic samples plot on the Ab-Or-Q diagram in a restricted area (Fig. 33) in the isobaric thermal trough for water vapour pressure around 5 Kb. Since the presence of normative An in significant amount affects the critical boundary lines for different water vapour pressures on the Ab-Or-Q diagram, the assessment of water vapour pressure from this diagram will be only a rough estimate. It must be noted that the analysed rocks of the present area contain An component in the range of about 14% in tonalite and 10% in granodiorite. However, von-Platen (1965), von-Platen & Holler (1966) and James & Hamilton (1969) have taken into account the presence of An in the Ab-Or-Q-H_2O system, and comparison of the Cooke Hill intrusives with their diagrams, a water vapour pressure of about 7 Kb could be suggested for the present rocks. Kleeman (1965) has discussed progressive melting in the system An-Ab-Or-Q-H_2O with reference to projection of the SiO_2-saturated surface on to the An-Ab-Or diagram. In this system the Cooke Hill granitic rocks plot in a continuous elongate field in the plagioclase region adjacent to, and trending towards the thermal valley (Fig. 34). This is consistent with an origin involving crystal-liquid equilibria. The data of von-Platen & Holler (1966) and Winkler (1971, p. 178-199) suggest that melting of gneisses could yield a large amount of magma approximating to the composition of granodiorite and tonalite at a temperature around 700°C at 7 Kb H_2O, provided sufficient water was available to saturate the melts. The continuous transition from tonalite to granodiorite suggests a progressive melting of crustal rocks. It is therefore possible that during tectonic deformation, a series of magmas were generated at different levels from metasediments with a range of granitic compositions. The high temperature tonalitic magma would thus be expected to have evolved at greater depths whilst the granodiorite magma, involving partial melting at lower temperature, could have originated at a slightly higher level. The water to facilitate anatectic melting in underlying metasediments could be derived from breakdown of hydrated minerals such as micas, though the water content of these minerals may not be over 5%. Therefore additional water required for generation of saturated or nearly saturated granitic magmas must be aided by a steady influex of water either from the mantle (Baily, 1970) or from the dehydration of hydrous minerals during metamorphism of surrounding and underlying crustal rocks.

Major and trace element chemistry of the Cooke Hill tonalite and granodiorite is compatible with greywacke type sediments (Tables 16 and 17. The average weight percentages of CaO, Na_2O and K_2O in the Cooke Hill tonalite and granodiorite is presented in Figure 36. They clearly show that the compositions of the Cooke Hill suite of rocks plot at the edge of the various groups of igneous rocks as compiled by Nockolds (1954), being rather more sodic than Nockolds average, but within the field of greywacke as defined by Condie.

To test the theory that magmas of the Cooke Hill originated from melting of greywackish type sediments, the major and trace elements are compared with some of the greywackes from the U.S.A. and Germany. A not unreasonable comparison is found except that the greywackes contain more Fe and Mg. Fe and Mg are generally contained in mafic minerals which require a much higher temperature for melting than does quartz-feldspathic material. The Na/K ratio is higher in tonalite of the present area than in Nockolds average tonalite (see Table 16). This may be related to the anatectic process resulting in enrichment of Ab component in plagioclase. The experimental work of Wyllie & Tuttle (1961) and Orville (1963) demonstrate that if granites have been derived from melting of sediments in the presence of high water vapour pressure, they become significantly richer in sodium than the original sediments available.

The trace element values of the Cooke Hill tonalite-granodiorite are somewhat closer to values reported for greywackes than to those of common igneous rocks (Table 17). The distribution of Rb in diorite granodiorite, and granite is reported as 70 ppm, 170 ppm and 250 ppm respectively (Tauson & Stavrov, 1957). These values are too high for the Cooke Hill intrusives. The high Ca-granites (quartz/diorite/granodiorite) of Turekian and Wedepohl (1961) have Ca and K values of 25300 ppm and 25200 ppm respectively and comparable with the Cooke Hill tonalite-granodiorite but their reported trace elements, Rb, 110 ppm is higher, and Sr 440 ppm and Ba 420 ppm are lower than the Cooke Hill suite. Therefore, the trace element concentrations and ratios are more nearly comparable with those of sedimentary materials rather than with the common igneous rocks. For comparison the greywackes of Precambrian and Paleozoic age have Rb values of 88 ppm (Condie, 1967) and 51 ppm (Whetten, 1966) and their K/Rb ratios are 230 and 285 respectively (see table 17). The average value of Rb in the Cooke Hill tonalite is 43 ppm and in the granodiorite is 96 ppm and their K/Rb ratios are 295 and 299 respectively. It thus appears that the Cooke Hill intrusives having the high K/Rb ratios are more likely to be derived by melting of greywacke type sediments but with variable degree of anatexis. The average K/Rb, Ca/Sr and Rb/Sr ratios of common igneous rocks have been estimated by Condie (1967) from the published results of Turekian & Wedepohl (1961), Heier & Adams (1964), Ahrens (1954) Rooke (1964) and Kolbe & Taylor (1966). These values are shown in Table 17. The Cooke Hill suite have significantly higher K/Rb ratios than the other acid igneous rocks. The K/Rb ratio is close to that of greywacke (Table 17; 7). The Rb/Sr ratio is much lower than Condies estimates for normal granodiorites and tonalites and again relates to greywackes. Similarly the Ca/Sr ratio is lower than Condies averages for granodiorite and tonalite. Normal tonalitic rocks have Ba/Rb ratios in the range 4-5 and Ba/Sr ratios in the 6-7 range (Gribble, 1969) whereas in the Cooke Hill suite the Ba/Rb ratios cluster around 18-23 and the Ba/Sr ratios are low (1.6-2.7). This clearly shows that trace element distributions and the respective ratios of K/Rb Ca/Sr, Rb/Sr, Ba/Rb and Ba/Sr in the Cooke Hill granitic rocks do not stand comparison with those of the common igneous rocks and thus an anatectic origin from melting of country rocks (greywacke) may be suggested.

TABLE 15

Major and trace elements in the metasedimentary rocks

Sample No.	138	139	205	201	184	149
Na%	3.27	3.12	2.52	2.81	2.73	3.26
K%	0.89	3.34	3.48	1.37	3.53	2.07
Ca%	0.66	0.44	0.30	n.d.	0.68	0.79
Rb ppm	36	113	168	32	134	76
Sr ppm	139	108	45	203	69	312
Ba ppm	238	894	524	270	535	1202
K/Rb	272	295	207	428	263	272
Sr/Ca.10^3	21	24	14	0	10	39
Ba/Sr	1.71	8.28	11.64	1.33	7.75	3.85
Ba/Rb	7	8	3	8	4	16
Ba/K.10^3	24	27	15	20	15	58
Na/K	3.33	0.93	0.72	2.05	0.77	1.57

TABLE 16

Average composition of Cooke Hill intrusives, common igneous rocks and greywackes

	1	2	3	4	5	6	7	8	9
SiO_2	68.69	70.56	68.59	72.08	66.88	66.15	64.43	67.67	66.70
TiO_2	0.35	0.26	0.40	0.37	0.57	0.62	0.62	0.57	0.54
Al_2O_3	16.52	16.05	15.49	13.86	15.66	15.56	15.48	13.41	n.d.
Fe_2O_3*	2.27	1.75	3.09	2.72	4.21	5.16	6.54	6.27	6.12
MgO	1.08	0.73	2.11	0.52	1.57	1.94	3.12	3.23	n.d.
CaO	2.98	1.89	3.66	1.33	3.57	4.65	2.22	3.04	3.09
Na_2O	5.40	4.66	4.93	3.08	3.84	3.90	3.74	2.99	1.90
K_2O	1.61	3.46	1.08	5.46	3.07	1.42	2.44	2.02	1.75
Na:K	3.20	1.20	4.10	0.50	1.10	2.50	1.40	1.30	1.00

* Expressed as total iron

1. Average Cooke Hill tonalite
2. Average Cooke Hill granodiorite
3. Massive granodiorite
4. Average granite and adamellite (Nockolds, 1954)
5. Average granodiorite (Nockolds, 1954)
6. Average tonalite (Nockolds, 1954)
7. Average Wyoming Pre-Cambrian greywacke (Condie, 1967)
8. Average Pre-Cambrian greywacke (Pettijohn , 1963)
9. Composite of German greywacke (Palaeozoic) (Wedepohl in Pettijohn, 1963)

TABLE 17

Trace elements in the Cooke Hill intrusives, common igneous rocks, greywackes and metasedimentary rocks of the Cooke Hill area

	1	2	3	4	5	6	7	8
Rb ppm	43	96	41	170-250	75-110	88	51	32-168
Sr ppm	621	656	342	80-100	200-400	424	120	45-312
Rb/Sr	0.06	0.14	0.11	2.00-3.00	0.30-0.50	0.23	0.43	0.15-3.75
K?Rb	295	299	219	160-230	210-250	230	285	207-428
Ca/Sr	34	20	76	50-100	40-80	37	184	25-73

1. Average Cooke Hill tonalite
2. Average Cooke Hill granodiorite
3. Massive granodiorite
4. Granite - adamellite
5. Granodiorite-tonalite

)Estimated by Condie (1967) from the data
)of Turekian & Wedepohl (1961), Heier & Adams
)(1964), Ahrens (1954), Rooke (1964) and
)Kolbe & Taylor (1966)

6. Average Wyoming Pre-Cambrian greywacke (Condie, 1967)
7. Composite of German greywacke (data of K.H. Wedepohl in Pettijohn, 1963)
8. Average Cooke Hill metasedimentary rock.

Greywackes to provide source material are not uncommon in the Kanmantoo Group sediments. The geochemistry of surrounding country rocks (Table 15) is extremely variable due to variations in their bulk compositions (Table 3D), but some of the country rocks occurring in the immediate vicinity of the Cooke Hill intrusives (for example, see sample No. 138 & 139 in Table 15 show comparable trace element chemistry. It is therefore assumed that the Cooke Hill granitic magmas were generated by differential anatexis of country rocks in deeper parts during regional metamorphism. Subsequently, tectonic movements forced the anatectic magmas to move upward, intruding them as small bodies in the form of lenses, dykes and sills. This mechanism of generation of magmas may explain the chemically variable composition of the resultant rocks. Therefore, the process of differential anatexis appears to be operated independently, producing different batches of tonalitic to granodioritic magmas in selective bands or zones of underlying country rocks. This scheme of magmatic evolution has been proposed to explain the variation in composition of magmas and the resultant rocks. In the first stage of magma formation the melt would have had a minimum melting temperature composition, but it would have incorporated increasing amount of higher melting temperature constituents as the degree of melting increased. It will be assumed, in view of their orogenic association and lack of associated basic intrusives, that the magmas originated within the crust and were not derived from the mantle.

8.3. CONCLUSIONS

The petrochemistry of the Cooke Hill intrusives suggests that tonalite-granodiorite suite crystallised from an anatectic magma. The nature of the rocks in the region of magma generation is not known but there is a strong presumption that they could be the downward extension of the surrounding country rocks. The tonalites contain more calcium, iron and magnesium than the granodiorites, and their magmas may be presumed to represent a range of temperature as having the highest degree of melting in the region of tonalitic magma formation and lower degree of melting for granodioritic magma formation. It is therefore possible to predict the sequence of magmas which were produced during an episode of partial melting. The first low temperature magmas would have had a composition near the minimum in the system Ab-Or-SiO_2-H_2O at the appropriate water vapour pressure, and have given rise to granodiorite with richer quartzo-feldspathic material and lower content of ferromagnesian minerals. Further melting at a higher temperature would have produced a magma containing more calcium, iron and magnesium, giving rise to a tonalite with relatively less quartzo-feldspathic material and more ferromagnesian minerals. Because of the degree of melting, the differentiation into two rock series, one relatively rich in potassium and the other poor in potassium, could have occurred during the formation of the magmas, rather than by fractional crystallisation. During the cooling period, each type of magma may have formed by more or less complete crystallization.

8.4. SOME COMPARISONS OF THE COOKE HILL INTRUSIVES WITH GRANITES OF THE EASTERN MT. LOFTY RANGES, SOUTH AUSTRALIA

There are numerous bodies of granite in the eastern Mt. Lofty Ranges and contributions have been made by Goode (1927) for the Mannum granite, Alderman (1927, 1929) for the Reedy Creek tonalite and the Mannum aplite, Kleeman (1934) for the Murray Bridge granite, Rattigan and Wegener (1951) for granitic rocks of the Palmer area and Sando (1957) for the Reedy Creek granodiorite. More recently, detailed studies have been carried out by Mills (1964) for the Cambrai granodiorite, Worden (1965) for the Murray Bridge and Mannum granites, Offler (1966) for the Mt. Kitchener granite and White (1956, 1966b; White et al., 1967) for the Rathjen-Palmer granites. Unfortunately, little work has been done on the trace element chemistry of these granites except that by White (1966b, 1967) and Worden (1965). They have presented some trace element data for petrogenesis of the Rathjen granite gneiss, the Palmer granite, the Murray Bridge and the Mannum granites.

In this section, the major and trace element chemistry of present tonalite and granodiorite is compared with that of these other granites. The Na_2O/K_2O ratios of the Cooke Hill tonalites and the Massive granodiorite are characterized by high values (3.35 and 4.56 respectively) as compared to the Palmer granite (0.92), the Swanport granite (0.72), the Mannum granite (0.65), the Monarto granite (0.89), the Mannum aplite (1.02), the Mannum tonalite (2.27) and the Cooke Hill granodiorite (1.34) (see Table 18).

A plot of CaO, K_2O and Na_2O is presented in Figure 36, which clearly shows that all granitic rocks fall in three distinct fields. The most striking fact that comes out of this plot is that only Cooke Hill tonalites and granodiorites fall in the field of greywacke (Condie 1967), thus supporting the present suggestion for genesis of the Cooke Hill intrusives by anatexis of greywacke type metasediments. In chemical and mineralogical composition the other granites ocorrespond well to the fields of common igneous rocks. The Mannum aplite falls in the field of granodiorite, but Alderman (1929, p. 250) has stated that it is dominated by plagioclase rather than potash feldspar and that the chemical composition shows slightly higher soda content than potash content. The Reedy Creek tonalite near Mannum (Alderman, 1927) and the Cooke Hill tonalites both fall in the field of common igneous rock representing the composition of tonalites (see Fig. 36); however, the Reedy Creek tonalite is richer in calcium than the Cooke Hill tonalites due to the presence of hornblende in equal amount with biotite in the former rock. On the basis of major element chemistry of granites from the Mannum, Monarto and Swanport areas and the Reedy Creek tonalite near Mannum, Alderman (1929) suggested that all these granites and tonalites have been derived from a parent magma and represent the small "cupolas" at the roof of a large batholith. The presence of such a batholith underlying the Murray Plain has been confirmed by mapping by Sando (1957) and White and Thatcher (1957) on the Mannum Sheet (S.A. Geol. Survey). Detailed studies of granitic rocks in the

TABLE 18

Chemical analyses of some granites from the Eastern Mt. Lofty Ranges

	1	2	3	4	5	6	7	8	9
SiO_2	74.12	74.50	70.77	72.42	73.49	63.88	68.69	70.56	68.59
TiO_2	0.31	0.27	0.72	0.22	0.25	0.86	0.35	0.26	0.40
Al_2O_3	13.49	12.80	13.69	15.49	14.14	16.37	16.52	16.52	15.49
Fe_2O_3	1.13	0.86	1.97	0.44	1.26	1.99	0.47	0.48	0.98
FeO	0.87	1.11	0.97	1.03	0.69	2.96	1.80	1.27	2.11
MnO	0.02	0.04	0.00	0.00	0.00	0.00	0.09	0.07	0.08
MgO	0.45	0.28	0.34	0.20	0.44	2.24	1.08	0.73	1.06
CaO	1.39	1.14	0.94	1.44	1.60	5.18	2.98	1.89	3.66
Na_2O	3.80	3.45	3.70	4.30	3.75	3.66	5.40	4.66	4.93
K_2O	4.09	4.75	5.68	3.78	3.67	1.61	1.61	3.46	1.08
P_2O_5	0.13	0.06	0.11	0.19	0.11	0.23	0.14	0.13	0.16
L.o.ignit. or H_2O	0.34	0.42	0.81	0.30	0.47	0.66	0.59	0.49	0.76
Total	100.14	99.70	99.70	99.85	99.87	99.64	99.72	100.04	99.30
Na_2O/K_2O	0.92	0.72	0.89	0.89	1.02	2.27	3.35	1.34	4.56

1.	Average Palmer granite (White et al. 1967)
2.	Swanport granite (AMDL)
3.	Mannum granite (Goode, 1927)
4.	Monarto granite (Chapman, 1923)
5.	Mannum aplite (Alderman, 1929)
6.	Reedy Creek tonalite near Mannum (Alderman, 1927)
7.	Average Cooke Hill tonalite
8.	Average Cooke Hill granodiorite
9.	Massive granodiorite

Palmer region have been carried out by White (1966b, 1967) involving the major and trace element chemistry. He proved that the Rathjen granite gneisses are isochemically metamorphosed sediments while the Palmer granite is an intrusive body derived by melting of the Kanmantoo Group sediments at greater depth. The age of the syn-tectonic Palmer granite is approximately 490 ± 15 m.y. as determined by Rb/Sr methods.

Trace element data for intrusive granites is available only for four samples of the Murray Bridge granites and one sample of the Mannum granite (Worden, 1965), and seven samples of the Palmer granites (White et al., 1967). Some data for granites of the Keith-Tintinara region are also available (Henstridge, 1970). The concentration ranges for Sr, Rb and Ba in the Cooke Hill intrusives, the Massive granodiorite of the Rathjen Gap and other granites from different areas, are presented in Table 19. Unfortunately, no data is available for Ba in the Murray Bridge granites and the Mannum granites.

The petrochemistry of the Cooke Hill intrusives (including the Massive granodiorite) shows that they fall into a separate field of the Na_2O-CaO-K_2O diagram (see Fig. 36) by virtue of their high soda content as compared with other granitic rocks. This composition can be better gauged from the trace element data and in particular Ba and Sr. The granitic rocks, having a high K/Rb ratio in the range 200-300, are considered to be anatectic granites (see Table 19).

A comparison of K/Rb ratios for all granites is presented in Figure 37 and shows correlation between the Palmer granites, the Mannum granites and the granites of the Cooke Hill area. There is a much lower absolute abundance of K and Rb in the Cooke Hill tonalites and the Massive granodiorite than in the rest of the granites. This close relationship of K/Rb between granites of these three areas supports their anatectic origin as proposed by White (1967) for the Palmer granite and Worden (1965) for the Mannum granite. The Rathjen

TABLE 19

Trace element data in parts per million

Sample No.	Sr	Rb	Ba	K/Rb	Ba/Rb	Ba/Sr	Sr/ Ca.10^3	Ba/ K.10^3
			Cooke Hill Tonalites					
75	664	35	1198	286	34	1.8	31	119
140	669	35	675	363	19	1.0	33	53
199	542	52	588	275	11	1.1	26	41
233	653	49	1759	237	36	2.7	30	152
295	551	46	785	313	17	1.4	28	54
Mean	662	43	1001	295	23	1.6	30	84
			Cooke Hill Granodiorites					
137	665	93	1906	312	20	2.9	47	66
232	647	100	1667	285	17	2.6	50	58
Mean	656	96	1786	299	18	2.7	49	62
			Massive Granodiorite					
208	342	41	565	219	11	1.6	13	63
			Palmer Granites					
P7	52	151	259	247	1.7	5.0	8	6.9
P6	78	153	360	251	2.4	4.6	13	9.4
P4	74	142	338	236	2.4	4.6	7	10.1
SA22	82	179	419	212	2.3	5.1	9	11.0
P5	70	144	391	208	2.7	5.6	5	13.0
SA23	77	151	365	213	2.4	4.7	6	11.4
P2	103	138	388	208	2.8	3.8	9	13.5
Mean	76	151	360	225	2.4	4.8	8	10.7
			Murray Bridge Granites					
TR1	86	283	n.d.	149	n.d.	n.d.	11+	n.d.
TR2	87	267	n.d.	158	n.d.	n.d.	n.d.	n.d.
TR3w	75	268	n.d.	151	n.d.	n.d.	n.d.	n.d.
MB1	73	313	n.d.	130	n.d.	n.d.	n.d.	n.d.
Mean	80	283	-	147	-	-	11+	-
			Mannum Granite					
TR1	142	193	n.d.	219*	n.d.	n.d.	21*	n.d.
			South East Granites					
	12-93	216-521	134-913	99-190	0.3-2.0	8-12	5-15	3-12

*Recalculated as correct ratio from the Worden's data

+Calculated assuming CaO = 1.00%

granite gneiss which is a para-gneiss shows similar correspondence with the anatectic granites. The Murray Bridge granites show strong fractionation trends and have lower K/Rb ratios (Table 19). The K/Ba values are plotted in Figure 38; there is no correspondence between the Palmer granite, the Cooke Hill granites and the Rathjen Gneiss, suggesting that although these are all of anatectic origin they were derived from compositionally different sources. It should be noted that although the absolute abundance of K and Ba varies within the Cooke Hill intrusives, they appear to show correspondence with each other indicating their genetic relationships. The Palmer granite stands out alone quite different from all other rocks. The Ba/Rb ratio of the Palmer granite is the lowest and the Cooke Hill granodiorite is highest as shown in Figure 39. There is no correlation between Ba/Rb of the Palmer granite, the Rathjen Gneiss and the Cooke Hill granites; but a linear correlation between the Cooke Hill granodiorites and the Cooke Hill tonalites is apparent in Figure 39. The absolute abundances of Ba and Sr are higher in the granites of the Cooke Hill area, higher than normal. This shows up particularly in the two ratios Ba/K and Sr/Ca (Table 19). The Cooke Hill granites have two $Sr/Ca.10^3$ ratios ranging from 30 for tonalite to 50 for the granodiorite and 32 for the Massive Granodiorite, compared with 5 to 13 for the Palmer granites, 11 for the Murray Bridge granites, 21 for the Mannum granites and 5 to 15 for the South East biotite granites. This reveals an absolute enrichment in Sr in addition to the enrichment of Ca in the Cooke Hill granites. This agrees with the results of Turekian and Kulp (1956) who show that granites with a high Ca content always have high Sr values. A comparison of Ca/Sr values for all granites is presented in Figure 40 showing no linear relationships between granites of different areas. The $Ba/K.10^3$ ratio of the Cooke Hill granites (mean 76), is much higher than the Palmer granites (mean 11) and the even lower ratios for the South east granites.

It is interesting to note that the Cooke Hill metasediments have a mean Sr/Ca.10^3 ratio of 21 and a mean Ba/K.10^3 ratio of 26, while the quartzo-feldspathic schists and the Rathjen granite gneiss of the Palmer area (White, 1966b, Table VII) have mean Sr/Ca.10^3 ratios of 29 and 10, and mean Ba/K.10^3 ratios of 34 and 27 respectively.

Thus it is concluded that the granites of the eastern Mt. Lofty Ranges and adjacent areas having different compositions have been derived by different geochemical processes. The anatectic origin for the Palmer granite, the Cooke Hill intrusives and the Monarto granite is well established in the light of major and trace element chemistry; these granites are unrelated to the batholith believed to underlie the Murray Plain. The Murray Bridge granites are strongly differentiated magmatic rocks and probably derived from a parent batholith. It is suggested that a detailed study, including the major and trace element chemistry, of granites from other areas of the Mt. Lofty Ranges could add greatly to the understanding of the petrogenic history of acid igneous rocks.

CHAPTER 9

CONCLUDING DISCUSSIONS

In the preceding chapters various aspects of the geology of the Cooke Hill area which bear on the origin of the migmatites and other granitic rocks, have been described. These studies have been undertaken to determine the petrographic and geochemical nature of and relations between, the migmatites and granitic rocks (Chapters 5, 7 and 8), their regional setting (Chapter 1) and their relation to the history of deformation and metamorphism (Chapter 2).

As the principal result of this work it has been possible to show that the migmatites and intrusives are not genetically related. As a by-product of the work, the structural studies have also demonstrated superposition of fold episodes on the macroscopic scale, for the first time in the Kanmantoo rocks. The final chapter brings together the principal arguments and conclusions on the structural evolution and petrogenesis of migmatites and granitic rocks in the Cooke Hill area. More detailed discussions have already been given at the end of each chapter.

Regionally, the migmatite zone, which occurs within the sillimanite-K-feldspar sub-facies of the amphibolite facies, is confined to rocks of the Cambrian-Ordovician Kanmantoo Group which was deposited in a trough immediately before the Delamerian orogeny (Chapter 1). The migmatite belt and the metamorphic belt with fibrolitic sillimanite lie oblique to the main fold trend (figs. 2 and 4, Map 3), and affect all stratigraphic units. Thus, migmatites occur in the uppermost Bruckinga Formation and in view of the conclusion that they are not the consequence of partial melting, the experimental evidence indicates temperatures of near 650°C and pressures of 3 kb during their formation assuming P total = P fluid (fig. 15B). There is no evidence to indicate what rock unit may have provided the necessary load pressure, and in any case high geothermal gradients must be postulated. On the other hand the migmatite zone must be underlain by a very great thickness of Kanmantoo and possibly Adelaidean rocks.

The Cooke Hill area lies on the western margin of the migmatite zone, which occupies a major synclinal structure (the Burns Syncline) in the Buckinga Formation (Map 3). In the Cooke Hill area the migmatites are best developed within quartzo-feldspathic units of the Inman Hill Formation (Chapters 2 and 4) and are approximately concordant with the west limb of the Cooke Hill Synform; but the boundary of the migmatites cuts across the stratigraphic and structural trends in the east limb of the synform. Thus the NE boundary of the migmatites is broadly parallel to the general NNW trend of the migmatite amd metamorphic belt. Other original rock types of the Inman Hill Formation in the migmatite zone did not give rise to migmatites but produced calc-schists and calc-gneisses (Chapter 3 and 6 and Map 1). Some lineated granitic gneisses are also interpreted as of metasedimentary origin (Chapter 4.7).

The structural evidence (Chapters 2 and Fig. 9) shows that the Cooke Hill synform is a second generation structure. It refolds an earlier penetrative layer-parallel schistosity and is associated with the local development of a new axial plane structure, mainly in the hinge zones, as a crenulation cleavage in the micaceous rocks and as a secondary schistosity in the quartzo-feldspathic rocks. Moreover, it is a downward facing structure so that the migmatites of the Inman Hill Formation lie in the over-turned western limb of a major first generation anticline (here called the Somme Anticline) which was identified by Mills (1964). The Cooke Hill tonalites and granodiorites lie within the same quartzo-feldspathic portion of the Inman Hill Formation in the Cooke Hill Synform and exhibit some control by its axial plane trend. However, at Cooke Hill itself these intrusive rocks lie outside the belt of migmatites (Map 1).

Within the migmatite zone there is a general increase in grain size from NE to SW and the migmatitic veins themselves are much coarser grained than the host rocks (Chapter 5). The abundance of migmatitic veins appears to be related to the intensity of mesoscopic folding. Both host and veins are essentially plagioclase-quartz rocks with only small and highly variable amounts of potash feldspar. The veins carry a more constant and smaller amount of quartz (ca 30%) than the host rocks. They are generally concordant and folded by F_2 mesoscopic folds but they can be observed parallel to both F_1 and F_2 axial planes and it is inferred that their development began during F_1 and that the waning phases of development overlapped F_2 (Chapter 5). Their development thus followed immediately the main peak of metamorphism (Offler and Fleming, 1968). The Cooke Hill tonalites and granodiorites and the massive granodiorites are similar in their general mineralogy to the the migmatites but they differ in their detailed geochemistry and they are essentially discordant post-tectonic intrusives and thus post-date the main development of migmatite veins (Chapter 5). The intrusives retain clear evidence of their magmatic origin in their abundance of disoriented xenoliths, their hypidiomorphic texture, oscillatory feldspar zoning and euhedral zircons. Pegmatites and aplites associated with the intrusives are usually rich in microcline and are quite distinct from the migmatitic veins (Chapter 5).

Geochemically the migmatite veins are notable in containing a higher proportion of soda and lower total iron than either the metasediments or the Cooke Hill intrusives (figs. 18 and 19). This can be correlated mineralogically with increased plagioclase feldspar and lower biotite in the veins. Sr is relatively enriched and Rb and Ba depleted in the veins. The Ba/Rb ratio for the veins is lower and the K/Rb ratio higher, on the average than for the Cooke Hill intrusives and therefore supports the structural evidence in arguing against an origin of the veins as a highly fractionated residue of the Cooke Hill magma. The study of the chemistry of the biotites indicates that biotites from the metasediments, the migmatitic veins and the Cooke Hill intrusives have all crystallised under similar P_{H_2O}-P_{O_2}-T conditions but does not elucidate further the genesis of the migmatites.

Examination of the geochemical data in the context of the experimental systems shows that the Cooke Hill intrusions probably originated by anatexis from graywacke type sediments with enrichment of the magma in soda due to high water vapor pressure (Chapter 8). The structural setting of the intrusives above a thick pile of folded sediments of the Kanmantoo Group and possibly of the Adelaide Supergroup (Chapter 1) supports this interpretation. On the other hand the geochemistry of the migmatite veins (especially the low K and Rb contents and the relatively high K/Rb ratio) shows that they cannot be regarded as a magmatic differentiate nor as an anatectic melt from their host rocks. Fig. 29 shows that in the Ab-An-Or-SiO_2 system of Kleeman (1965) the migmatite veins fall away from the differentiation trend of magmatic rocks. Instead, an origin by metamorphic differentiation and metasomatism is proposed (Chapter 5.3). It is inferred from the field and petrographic evidence that the process was one in which the felsic material was mobilised on a scale of a few inches to a few feet and tended to become concentrated in structurally controlled areas. Following Dietrich (1959) and Starmer (1969) it is supposed that, in the presence of fluids, pressure gradients exist in the rocks; and that the veins are segregated in low pressure sites, where the fluids have facilitated the coarsening of grain size. This hypothesis of metamorphic differentiation accounts for the relationship between the composition of the veins and the composition of the host rock in both the Cooke Hill and Palmer areas (White 1966a). In the Cooke Hill area the metasediments have a high Na/K ratio and an abundance of CaO and were able to produce an aqueous fluid phase richer in soda than the host. On the other hand, the metasediments in the Palmer area are relatively rich in potash and could produce a yet more potash rich fluid (cf. Orville 1963). That metasomatism by aqueous fluids, as well as metamorphic differentiation was operative in the Cooke Hill migmatites is indicated by the high Na/K ratios of the veins compared with the host rock, by the sericitization of the plagioclase in the veins but not in the host rocks and by the coarser grain size of the veins (Chapter 5). This work therefore supports and greatly strengthens White's (1966a) conclusions regarding the petrogenesis of the migmatites and also places them in the metamorphic and structural history of the region.

A clear distinction is therefore drawn between the origin of the migmatitic veins and the origin of the intrusives in the area. The migmatites were formed during deformation and metamorphism by metamorphic differentiation and metasomatism and their composition varies from area to area according to the composition of the host rocks. The intrusives were formed by anatexis at greater depth and were then emplaced at higher levels during and after the final stages of deformation and after the formation of the migmatitic veins.

APPENDIX 1

APPENDIX 1

PETROLOGICAL DESCRIPTION OF ROCKS TAKEN FOR CHEMICAL ANALYSIS

This section contains brief petrographic descriptions of specimens used for chemical and mineralogical analyses. All thin sections were prepared by cutting the rock at right angles to visible schistosity (foliation) and lineation.

A285/372 Leucocratic vein of migmatite

A light coloured, coarse grained rock with weak mica schistosity. The quartz and plagioclase are predominant minerals with small amount of potash feldspar, biotite and muscovite. The grain sizes of constituent minerals ranges as 1-2 mm.

Quartz: occurs in variable shapes and sizes; some occur as small drop-like inclusions in feldspar and others reach up to 2 mm as porphyroblasts. Undulose extinction is visible in some grains.

Plagioclase: Both twinned and untwinned grains are abundant; twinning is predominantly on albite law. Alteration to sericite and flaky muscovite is common. A few grains show diffuse zoning. The composition of plagioclase measured by extinction angle on albite law twinned grains is An_{31}o.

Potash feldspar: It is mostly untinned but a few grains show cross-hatch twinning. The staining test reveals the presence of K-feldspar in a small amount.

Biotite: Light yellowish brown (X) to dark brown (Z). It shows a weak preferred orientation among quartz-feldspar rich granular matrix. A few grains are slightly altered to chlorite along the margin.

Muscovite: It is less common as a primary mineral but secondary muscovite as a fine grained sericite and post tectonic coarser flakes are commonly associated with plagioclase.

Apatite, zircon and iron ores are common among the accessory minerals.

A285/377 Leucocratic vein of migmatite

A light coloured, coarse grained weakly schistose rock. Quartz and plagioclase form the major constituents of rock with small amount of biotite and muscovite. The average grain size is about 1.5 mm.

Quartz: grain size varies considerably, some reaching up to 2 mm. The grain boundaries are generally sutured. Strain or wavy extinction is common.

Plagioclase: Twinned and untwinned grains are equally abundant; twinning is developed on albite law but it is discontinuous or patchy in most grains. Many grains are cloudy with fine "dust" with abundant sericite and flaky muscovite. The composition of plagioclase is An_{28}.

Biotite: Light yellowish brown (X) to dark brown (Z); generally occurs as discrete flakes with poor orientation.

Muscovite: mainly occurs as secondary muscovite intimately associated with altered plagioclase.

Apatite, zircon, iron ores and a few tiny needles of fibrolitic sillimanite are common accessory minerals.

A285/398 Leucocratic vein of migmatite

A light coloured, coarse grained rock (average grain size 1 mm) with weak mica schistosity. Quartz-plagioclase are major components with minor amount of potash feldspar, biotite and muscovite. Among the accessory minerals, apatite, zircon and iron ores are notable.

Quartz: generally even-grained with sutured boundaries. Some grains have undulose extinction.

Plagioclase: albite law twinned grains are common but untwinned grains are also present. Complex twinning on Carlsbad/albite and pericline laws may be seen in a few grains. Some grains are highly altered with abundance of fine "dust" and sericite. The composition of plagioclase is An_{29}.

Potash feldspar: a few grains in quartz-plagioclase matrix; mostly untwinned.

Biotite: pale yellowish brown (X) to dark brown (Z) with weak preferred orientation. A few grains are altered to chlorite along the margin.

Muscovite: few scattered plates of secondary muscovite.

A285/514 Leucocratic vein of migmatite

A light coloured, coarse grained rock; mainly composed of quartz-plagioclase with small amount of potash feldspar, biotite and muscovite. The average grain size is 1 mm.

Quartz: grain shape is variable, some show strain with numerous fine deformation lamellae.

Plagioclase (An_{28}): albite law twinning is common but it varies from finely developed patchy twinning to well developed broad twin lamellae. Untwinned grains are also abundant. Deformational features such as straining, glide twinning, kink bands, fracturing and bending of twin lamellae are seen in a few grains.

Potash feldspar: mostly untwinned and scarce.

Biotite: light brown (X) to chocolate brown (Z); a few tiny grains appear as inclusions in quartz and plagioclase. It occurs as a widely scattered laths with weak preferred orientation.

Muscovite: mainly as a secondary mineral in the form of fine sericite or coarsely crystalline laths in altered plagioclase.

Some myrmekite structures are seen along the boundaries of potash feldspar and plagioclase.

Apatite, zircon and iron ores are notable accessory minerals.

A285/515 Leucocratic vein of migmatite

A light coloured, coarse grained rock (average grain size 1.5 mm) with weak mica schistosity.

Quartz: grain size varies from small inclusion of less than 0.1 mm to coarse grain up to 2 mm. Strain or wavy extinction is common. A few grains have fine deformation lamellae. It is next most abundant mineral after plagioclase.

Plagioclase (An_{28}): twinning present in most grains; mostly on albite law. Deformational features such as glide twinning, kink bands, bent and broken twin lamellae are seen in some grains. All grains show weak zoning.

Potash Feldspar: generally with cross hatch twinning; forms a minor proportion of plagioclase matrix.

Biotite: light brown (X) to dark brown (Z) with weak preferred orientation. A few randomly oriented small grains are present as inclusions in quartz and plagioclase.

Muscovite: mostly as a secondary mineral replacing plagioclase along and across the cleavage and twinning planes. A few primary grains occur as large plates in association with biotite.

Apatite and zircon are common among the accessory minerals and iron ores are rather rare in this rock.

A285/479 Leucocratic vein of migmatite

Light coloured, coarse grained rock with grain size in the range of 1-3 mm.

Quartz: uneven grains with sutured boundaries. Some show strain and undulose extinction.

Plagioclase (An_{28}): mostly twinned on albite law, few grains show rare twinning as Carlsbad/albite and pericline laws. Strong deformation is shown by glide twinning and bent and broken twin lamellae. Bleb-like inclusions of quartz are seen in some grains.

Potash feldspar: generally untwinned; a few scattered grains in quartz-plagioclase matrix.

Biotite: light brown (X) to dark brown (Z) with weak preferred orientation.

Muscovite: as a secondary mineral, ranging from fine sericite to coarse skeletal grains as alteration product of plagioclase.

Chlorite: present along the margin of few altered biotite.

Among the accessory minerals, apatite, zircon and iron ores are notable.

A285/483 Leucocratic vein of migmatite

Like the other veins of migmatite, it is light coloured, coarse grained rock but as well as quartz and plagioclase as the major constituents, this rock is comparatively richer in potash feldspar.

Quartz: grains variable in shape and size, some show strong strain and wavy extinction with fine deformation lamellae.

Plagioclase (An_{30}): generally well twinned on albite law but a few grains show complex gwinning of Carlsbad/albite and pericline laws. Clouding and alteration to fine muscovite and sericite is seen in some grains.

Potash feldspar: has cross hatch twinning; grains variable from less than 0.1 mm to bigger grains of up to 1.5 mm.

Biotite: light brown (X) to dark brown (Z); plates of biotite form a poor mica schistosity.

Myrmekite: frequently myrmeketic structures are seen at the junction of plagioclase and potash feldspar.

Accessory minerals include apatite, zircon and iron ores.

A285/497 Leucocratic vein of migmatites

A light coloured, coarse grained rock with average grain size of 1.5 mm. Quartz and plagioclase form the major constituents of rock with minor amounts of potash feldspar and biotite.

Quartz: is generally present as highly strained grains.

Plagioclase: Twinned and untwinned grains are equally abundant. A weak diffuse zoning is seen in some grains. Grains are generally clouded due to presence of fine "dust" and sericite. It has a composition of An_{27}.

Potash feldspar: untwinned and scarce.

Biotite: is present as light brown (X) to dark brown (Z) platy mineral which forms a weak mica schistosity.

Apatite, zircon and iron ores are seen as accessory minerals.

A285/608 Leucocratic vein of migmatite

Light coloured, coarse grained rock with average grain size 1.5 mm.

Quartz: uneven grain size with curved boundaries; a few grains attain amoeboidal shape. Some grains are strained and show undulose extinction.

Plagioclase (An_{30}): generally well twinned on albite law and rarely show Carlsbad/albite and pericline twinning. Glide or deformation twinning may be seen in few grains. A few weakly zoned grains are present. Most grains are cloudy due to presence of fine, unidentified "dust" and alteration to sericite and flaky muscovite.

Potash feldspar: cross hatch twinning common in most grains; shows irregular distribution in quartz-plagioclase matrix.

Biotite: light brown (X) to dark brown (Z). It shows weak preferred orientation parallel to dominant mica schistosity of host gneisses.

Muscovite: less common as a primary mineral.

Apatite, zircon and iron ores are the main accessory mineral.

A285/705 Leucocratic vein of migmatite

It has a similar colour and texture to other migmatite veins. Unlike the other rocks, it has a higher proportion of potash feldspar which is the most abundant mineral after plagioclase and quartz. Small amounts of biotite and muscovite are also present.

Quartz: variable in size from 0.1 mm to 2 mm. Strong strain and undulose extinction common.

Plagioclase: has a composition of An_{30}. Both twinned and untwinned grains abundant; twinning on albite law most predominant with few grains rarely showing complex twinning of Carlsbad/albite and pericline laws. Alteration to fine sericite and flaky muscovite is common. Few bigger grains contain small inclusions of quartz.

Potash feldspar: typical cross hatch twinning is well developed. It forms interstitially between quartz-plagioclase grains. A few myrmekitic structures are present along the junction of potash feldspar and plagioclase grains.

Biotite: light brown (X) to dark brown (Z). It shows a weak preferred orientation parallel to schistosity of host gneiss.

Muscovite: primary muscovite rare; secondary muscovite in altered grains of plagioclase and less commonly in potash feldspar are common.

Apatite, zircon and iron ores are notable accessory minerals.

A285/474 Leucocratic vein of migmatite

A light coloured coarse grained rock with average grain size 3 mm. The major components of the rock are quartz and plagioclase with lesser biotite and muscovite. Apatite and zircon are the only accessory minerals.

Quartz: ranges in grain size from small bleb-like inclusions to granoblasts up to 3 mm. Strong strain and undulose extinction are very common. Thin deformation lamellae are generally present in bigger grains.

Plagioclase: is generally cloudy due to fine "dust". Most of the plagioclase grains are twinned; albite twinning is most common, but complex twinning on the Carlsbad/albite and pericline laws is also found. Deformational features such as glide twinning, bent and broken twin lamellae and kink bands are common. Sericitization and late stage muscovitization is prominent in most of the plagioclase grains. The composition is An_{31}.

Biotite: occurs as small flakes with X = light brown, Y = Z = chocolate brown. It shows a weak preferred orientation.

Muscovite: is generally a secondary mineral and occurs as small flakes or large stumpy skeletons up to 4 mm. It forms along the edges of plagioclase with further growth penetrating into the plagioclase in all directions. Sometimes, small grains of plagioclase are surrounded by large skeletons of muscovite.

A285/372AG Gneissic host of migmatite

A light grey, equigranular, quartz rich rock with prominent mica schistosity. The average grain size is less than 1 mm. The excellent schistosity (S_1) is defined by strong preferred orientation of biotite and muscovite.

Quartz: is generally equigranular with straight or gently curved boundaries. Strain and undulose extinction is common. Small inclusions of biotite are frequently seen.

Plagioclase: is generally untwinned but a few grains with albite law twinning are present. The composition of plagioclase is An_{31}.

Potash feldspar: is found in small amount. Some grains have cross-hatch twinning.

Biotite: occurs as light brown (X) to dark brown (Z) flakes. It is more abundant than in the leucocratic veins.

Muscovite: is found as a primary mineral associated with biotite.

The accessory minerals include apatite, zircon, sphene and iron ores.

A285/75 Tonalite

A light grey, equigranular rock with average grain size 2 mm.

Quartz: ranges in grain size from less than 1 mm to 3 mm. Generally, it has anhedral shape with undulose extinction.

Plagioclase: is subhedral to anhedral in shape. Albite law twinning is predominant but Carlsbad/albite and pericline law twins are also present. Normal and oscillatory zoned grains are common. The composition of plagioclase is An_{25-35}.

Biotite: occurs as discrete plates to ragged grains. It is very pale brown to dark brown in colour.

The accessory minerals are apatite, euhedral zircon and a few small flakes of secondary muscovite in plagioclase.

A285/140 Tonalite

A light grey coarse grained rock with a hypidiomorphic granular texture. The average grain size is 2 mm, a few grains reach up to 4 mm.

Quartz: occurs as anhedral grains, some show fracturing and undulose extinction. Some grains have small inclusions of biotite.

Plagioclase: is the most abundant mineral. Twinning is well developed on the albite law, but the twinning may be discontinuous or patchy. Strain is common. Oscillatory zoning is common. Some grains have calcic cores with abundant sericite and with small plates of white mica. The composition of plagioclase is An_{25-40}.

Biotite: X = very pale brown, Y = Z = dark chocolate brown. It occurs as interlocking masses in between quartz and plagioclase.

Muscovite: is present both as a primary and secondary mineral. Primary muscovite is generally associated with biotite; secondary muscovite occurs as stumpy skeletons and small flaky grains in altered plagioclase.

Accessory minerals include apatite, euhedral zircon and rare epidote.

A285/199 Tonalite

A light grey coarse grained rock with hypidiomorphic texture. The grain size ranges from 1 mm. to 3 mm.

Quartz: is anhedral, fractured and commonly exhibits undulose extinction. Small inclusions of biotite are often seen.

Plagioclase: is well twinned on albite law, rarely on the Carlsbad/albite and pericline laws. Zoning is common both as simple normal zoning or as oscillatory zoning. Zoned plagioclases have well developed twinning in the sericitized core with an untwinned outer margin. The composition range is An_{30-42}.

Potash feldspar: is found in small amount (5%) as aggregates of small grains with cross-hatch twinning.

Biotite: X - very pale brown, Y = Z - dark brown. Biotite occurs as small grains ranging in size from 0.1 mm. to large plates up to 2.5 mm. long.

Muscovite: primary muscovite is rare, but fine grained white mica is found in altered plagioclase.

Apatite and euhedral zircon are the main accessory minerals.

A285/233 Tonalite

A light grey, coarse grained rock with a grain size of 1 mm. to 3 mm. The rock is composed mainly of quartz and plagioclase with lesser potash feldspar, biotite and muscovite.

Quartz: is generally anhedral, fractured and exhibits strong undulose extinction. Small inclusions of biotite and plagioclase are often seen.

Plagioclase: is twinned predominantly on the albite law and rarely on Carlsbad/albite and pericline laws. Normal to oscillatory zoning is developed in some grains. The compositional range is An_{25-40}.

Potash feldspar: is seen both as twinned and untwinned grains.

Myrmekite: is found as quartz vermicules in plagioclase or intergrowths of potash feldspar and plagioclase near the boundary of the two minerals.

Biotite: occurs as small flakes of 0.1 mm. to large plates of 3 mm. long. It is generally found in clusters in quartz-plagioclase matrix.

Muscovite: is present in small amount as a primary and secondary mineral. The primary muscovite is associated with or interleaved in biotite flakes. The secondary muscovite appears as fine or coarse flakes (up to 0.5 mm.) in altered plagioclase.

The common accessory minerals are apatite, sphene and zircon.

A285/295 Tonalite

A light grey coarse grained rock with hypidiomorphic texture. The grain size ranges from 1 mm to 3 mm. Quartz and plagioclase with lesser potash feldspar and biotite are the major components of the rock. As with other tonalites plagioclase forms about two thirds of the total rock. Accessory minerals are muscovite, apatite, epidote, rutile and euhedral prismatic zircon.

Quartz: is anhedral. All large grains are strained and exhibit strong undulose extinction. Quartz also occurs as inclusions in plagioclase. Two sets of deformation lamellae at right angles to each other are present in some grains.

Plagioclase: occurs as euhedral to subhedral grains, a few with a tabular habit. Twinning is universal and albite law twins are the most abundant; pericline law twins are also present in a few grains. Bent and broken twin lamellae and glide twinning can be seen in a few strongly deformed grains. Normal and oscillatory zoning is common. A varying degree of cloudiness due to very fine grained "dust" and sericite is usual. The composition range is An_{25-38}.

Potash feldspar: is a less common mineral. A few grains show cross-hatch twinning.

Biotite: X = very pale brown, Y = Z = chocolate brown. It ranges in size from small grains (0.1 mm.) to big flakes up to 2.5 mm. long. Biotite appears in clusters in between quartz-plagioclase matrix.

Muscovite: primary muscovite is less common but secondary muscovite is abundant in the altered cores of plagioclase.

A285/511 Tonalite

A light grey coarse grained rock with hypidiomorphic texture. The average grain size is variable and ranges from 0.5 mm. to 3 mm.

Quartz: occurs as small rounded inclusions in plagioclase and as larger grains up to 3 mm. Most of the grains are strained and show undulose extinction. Small inclusions of plagioclase and biotite are frequently seen in larger grains.

Plagioclase: is the most abundant mineral and forms about two thirds of the total rock. It is euhedral to subhedral in shape, commonly with a tabular habit. Twinning is generally well developed, the most common being albite, and rarely pericline laws. Some of the megacrysts show strain. Deformation (glide) twinning is frequently seen. Zoning is present on various scales; it may be well developed as normal or oscillatory zoning or as weak discontinuous patchy zoned crystals. Alteration to fine grained "dust" and fine sericite is occasionally seen. The composition of plagioclase is An_{32-40}.

Potash feldspar: occurs as interstitial grains in groundmass. Twinning is less common but may be present as cross-hatch twinning in some grains. The amount of potash feldspar is less than 5%.

Biotite: is found as very pale yellowish brown to dark brown flakes. It generally appears in clusters in the matrix. Occasionally small rutile needles are seen as inclusions in biotite.

Muscovite: both primary and secondary muscovite are seen. The primary muscovite is associated with biotite while the secondary muscovite is found in altered plagioclase.

The common accessory minerals are apatite, zircon, secondary epidote and rutile.

A285/137 Granodiorite

A light grey coarse grained rock with hypidiomorphic texture. The grain size ranges from 1 mm to 3 mm. The major minerals are quartz, plagioclase and potash feldspar with a lesser amount of biotite and a little muscovite. The accessory minerals include apatite, sphene, epidote and iron ores.

Quartz: occurs as anhedral crystals with strong strain shadows. The individual quartz grains have sutured boundaries and are often fractured. Quartz also occurs as small inclusions in plagioclase and potash feldspar.

Plagioclase: appears as euhedral grains or subhedral aggregates. Twinning is generally well developed on the albite law. Small platy inclusions of biotite are present in a few grains. Weak to strong normal and oscillatory zoning is seen on various scales. The composition of plagioclase is An_{20-28}.

Potash feldspar: occurs as interstitial grains or as megacrysts up to 3 mm. in length. Some megacrysts are zoned. Cross-hatch twinning is strongly developed in most of the grains. Small inclusions of quartz, plagioclase and biotite are frequent. Myrmekitic structure forms near the boundary of plagioclase/potash feldspar or quartz/potash feldspar.

Biotite: is seen as flakes varying in size from 0.1 mm to 2 mm in length unevenly distributed in the interstices between the quartz and feldspar. It has X = yellowish brown, Y = Z = dark brown.

Muscovite: primary muscovite is rare but secondary muscovite occurs as small flakes in altered plagioclase.

Epidote: small granules of secondary epidote are occasionally seen in altered plagioclase.

A285/232 Granodiorite

A light grey coarse grained rock with hypidiomorphic texture. The grain size varies considerably from 1 mm to 6 mm. Quartz, plagioclase and potash feldspar and small amounts of biotite are the major constituents of the rock.

Quartz: ranges in grain size from bleb-like inclusions to large megacrysts up to 6 mm. Frequent large crystals give the porphyritic appearance in the rock. Fracturing, strain and strong undulose extinction are common features of the quartz grains.

Plagioclase: occurs as euhedral laths or subhedral tabular crystals. Twinning is universal, albite, Carlsbad, Carlsbad/albite and pericline laws being common. Weak to strong normal and oscillatory zoning is seen in some grains. Clouding due to incipient alteration to fine "dust" and small platy muscovite is common. The composition of plagioclase is An_{22-32}.

Potash feldspar: has well developed cross-hatch twinning and occurs as interstitial grains in between quartz and plagioclase matrix. Fine hair perthite is seen in some grains. Myrmekitic structure at the boundary of plagioclase and potash feldspar is often seen. It is slightly cloudy and alters to fine white mica. Small inclusions of quartz and biotite are occasionally seen. A few zoned grains are also found.

Biotite: X = very pale yellowish brown, Y = Z = dark brown. It ranges in size from 0.1 mm to 1.5 mm long. The uneven distribution of biotite results from a clustering in the matrix.

Muscovite: is less in amount than biotite. It occurs mainly as secondary muscovite, and is particularly abundant in the altered core of plagioclase.

Accessory minerals are apatite, sphene, zircon and iron ores.

A285/208 Tonalite

A dark grey coarse grained rock with hypidiomorphic texture. The grain size ranges from 1 mm to 4.5 mm. The main minerals are quartz, plagioclase and biotite with a small amount of potash feldspar. Accessory minerals include muscovite, apatite, sphene, zircon, secondary epidote, chlorite and iron ores.

Quartz: has anhedral shape with strong strain and undulose extinction. The grain size ranges from bleb-like inclusions in plagioclase to large megacrysts up to 4.5 mm.

Plagioclase: is well twinned on the albite, law and occasionally on the pericline law. A few grains show glide (deformation) twinning. Zoning is common; both normal to oscillatory type zoned grains are present. The cores of zoned plagioclase are generally sericitized and cloudy. The composition of plagioclase is An_{19-32}.

Potash feldspar: is present in minor amounts. Cross-hatch twinning is developed in most of the grains.

Biotite: X = very pale brown, Y = Z = dark brown. Granular masses of sphene and iron ores are often associated with biotite. Alteration to chlorite is common.

Muscovite: is found as an alteration product of plagioclase.

Some epidote is formed as an alteration product of plagioclase. It appears as a spongy weakly pleochroic pale yellow mineral. A few granules of epidote occur as inclusions in biotite.

A285/138 Quartzo-feldspathic gneiss

A light grey, well crystallized rock with an average grain size of 0.6 mm. Excellent schistosity (S_1) is defined by preferred orientation of biotite.

Quartz: shows variation in grain size. The grains are usually strained and show undulose extinction. The grain boundaries are generally curved.

Plagioclase: is found both as twinned and untwinned grains. Albite law twinning is the dominant one. The composition of plagioclase is An_{30}.

Potash feldspar: is the most abundant mineral after plagioclase and quartz. It shows cross-hatch twinning in some grains, but otherwise it is mostly untwinned.

Biotite: straw brown (X) to dark brown (Z). It has a flaky appearance with a grain size of 0.1 mm to 1 mm.

Muscovite: is found in small amounts. It appears as a primary mineral either associated with biotite or as discrete grains in the matrix. Secondary muscovite is present in altered plagioclase.

Accessory minerals are sphene, apatite, zircon and iron ores.

A285/139 Quartzo-feldspathic gneiss

A grey, medium grained granoblastic rock. The average grain size is 0.6 mm.

Quartz: subrounded to curvilinear grains with slight strain.

Plagioclase: is generally twinned, most commonly on the albite laws. Untwinned plagioclase is also abundant. A few grains show bent and glide twinning. The composition of plagioclase is An_{30}.

Potash feldspar: is present in small amount. It is generally untwinned, but some grains show cross-hatch twinning.

Biotite: is found as a platy mineral, X = pale yellow brown, Y = Z = dark brown. The strong preferred orientation in the S_1 plane is well developed.

Muscovite: as a primary mineral is found associated with biotite.

Accessory minerals are apatite, sphene and zircon.

A285/205 Mica gneiss

A dark grey semi-pelitic gneiss. The abundance of mica produces an excellent - S_1 schistosity. The average grain size is less than 1 mm.

Quartz: appears in varying sizes. Undulose extinction is present in some bigger grains. Some grains are elongated parallel to the S_1 schistosity.

Plagioclase: is found both as twinned and untwinned grains. Albite law twinning is most common. The composition of plagioclase is An_{29}.

Biotite: X = pale brown, Y = Z = dark brown. Strong preferred orientation is parallel to the S_1 schistosity. It makes up about one third of the total rock.

Muscovite: is found as a primary mineral interleaved with biotite flakes. It is less abundant than biotite.

Accessory minerals are apatite, zircon and a few sphene grains.

A285/201 Quartzo-feldspathic gneiss

A light grey, granoblastic rock with average grain size less than 1 mm. Excellent schistosity (S_1) is defined by parallelism of biotite flakes. Quartz-feldspar aggregates form well crystallised mosaics.

Quartz: grains of varying size, some display undulose extinction. Inclusions of biotite are often present in some bigger grains.

Plagioclase: ranges in grain size from 0.2 mm to 1 mm. Twinning is developed on albite law, but untwinned grains are abundant. Slightly bent twin lamellae are in a few grains seen. The composition of plagioclase is An_{33}.

Potash feldspar occurs as xenomorphic grains with cross-hatch twinning. It forms in the interstices between quartz and plagioclase.

Biotite: (X = light brown, Y = Z = dark brown). Strong preferred orientation of biotite lies in S_1, but some grains are oriented parallel to S_2 at a slight angle to S_1.

Fibrolite: occurs in eye-shaped nodules in which the mats of fibrolity are wrapped around the quartz grains. Quite a few quartz grains contain needles of sillimanite.

Accessory minerals are apatite, sphene and rarely subhedral zircon.

A285/184 Quartzo-feldspathic gneiss

A light coloured coarse grained rock with grain size ranging from 1 mm to 3 mm. The main constituents of the rock are quartz, plagioclase, potash feldspar and lesser biotite.

Quartz: occurs in variable sized grains 1 mm to 3 mm. It shows a varying degree of strain and undulose extinction.

Plagioclase: is generally twinned on its albite law, but untwinned grains are also present. Clouding due to fine grained "dust" and sericitization is common. Occasionally, deformation twinning and slight bending in twin lamellae are also seen. The composition of plagioclase is An_{28}.

Potash feldspar: has well developed cross-hatch twinning. It is the next most abundant mineral after plagioclase and quartz.

Biotite: is a light brown (X) to dark brown (Z) platy mineral. It shows moderate preferred orientation parallel to S_1.

Accessory minerals include irregular shaped sphene, apatite and iron ores.

A285/144 Coarse granitic gneiss

A pink coloured coarse grained rock with grain size 0.2 mm to 2.5 mm. It is composed mainly of quartz and plagioclase with only a small amount of biotite.

Quartz: appears as small grains in the matrix or as large granoblasts up to 2.5 mm. Weak deformation lamellae are also seen in a few of the bigger grains.

Plagioclase: is found both as a twinned and untwinned grain (.2-2.5 mm. in size). Coarse albite twinning is common. Generally, it is fresh and unaltered, but some grains are extremely altered to fine grained "dust" and fine sericite. Small inclusions of quartz are often seen in bigger grains. It has a composition of An_{30}.

Biotite: (X = light brown, Y = Z = dark brown). It occurs as flakes from 0.2 mm to 1 mm long. The degree of preferred orientation parallel to S_1 is rather low.

Accessory minerals are apatite and iron ores.

A285/142 Pelitic gneiss (xenolith)

A dark grey coarse grained rock with average grain size 1.5 mm. The rock is mainly composed of mica (55-60%), quartz (30%) and feldspar (10%).

Quartz: is seen as flattened grains (0.5 - 1 mm.) in biotite rich laminations. In less rich biotite areas, it is subrounded to polygonal in shape. Strain is present in most of the grains to a varying degree.

Plagioclase: is untwinned and only a few grains show faint twinning. Some grains are dimensionally oriented parallel to S_1.

Potash feldspar: is generally untwinned grains, some grains have small inclusions of biotite and quartz. Fine hair perthite is developed in some grains.

Biotite: (X = pale brown, Y = Z = dark brown) appears as long flakes. It shows a strong preferred orientation parallel to S_1. Some grains are slightly bent probably due to second deformation.

Muscovite: appears as flakes similar in shape to biotite, but a few stout porphyroblasts with inclusions of quartz and biotite are also seen. It shows a strong preferred orientation parallel to S_1 schistosity.

Among the accessory minerals, apatite and iron ores are most prominent.

A285/294A Quartzo-feldspathic gneiss (xenolith)

A dark grey coarse grained rock. Quartz, plagioclase, potash feldspar, biotite and muscovite are the main constituent minerals.

Quartz: (0.2-2.5 mm.): it has a subrounded to interlobate shape. Fracturing and undulose extinction are common. Small inclusions of plagioclase and biotite are often seen in bigger grains.

Plagioclase (0.5 - 2.5 mm): twinned and untwinned grains in about equal amount. Albite twinning is the most common. Some grains are weakly zoned. Clouding and sericitization are common. The composition of plagioclase is An_{32}.

Potash feldspar (0.5 - 1 mm): is generally seen as untwinned grains. It forms an interstitial matrix between quartz and plagioclase.

Biotite (0.1 - 2.5 mm): X = light brown, Y = Z = dark brown. Long flakes of biotite are strongly oriented parallel to foliaton and schistosity (S_1). Some biotite flakes show a moderate orientation oblique to S_1 schistosity probably indicative of schistosity S_2.

Muscovite: may be both primary and secondary. Primary muscovite has a flaky appearance and is associated with biotite. Secondary muscovite is found in altered plagioclase.

A285/117 Quartz-plagioclase-actinolite schist

A light grey coarse grained rock with granoblastic texture. The grain size ranges from 0.2 mm to 0.8 mm.

Quartz: gently curved or straight grain boundaries. Undulose extinction is common.

Plagioclase: generally twinned, predominantly on albite law. Small inclusions of quartz, sphene and actinolite are frequently seen. Some grains are strongly strained with wavy extinction. The composition of plagioclase is An_{31}.

Actinolite (0.1 - 0.5 mm in cross section): appears as prismatic crystals with pale green colour. Pleochroism is moderate.

Sphene (0.1 - 0,5 mm): is found as a lozenge shaped euhedral crystal to subrounded granular aggregates.

Rutile (0.2 - 0.3 mm): is present as euhedral to subhedral crystals.

A285/149 Quartzo-feldspathic Schist

Grey, medium grained rock, average grain size less than 3 mm. Schistosity (S_1) is well developed, with fine (1 - 3 mm) quartz/mica lamination parallel to schistosity. Average grain size less than 3 mm.

Quartz: straight or gently curved quartz to quartz grain boundaries; grains slightly flattened parallel to S_1 in mica rich bands.

Potash feldspar: scattered grains are occasionally seen in the matrix.

Plagioclase: some grains twinned on the albite law, but untwinned grains are more abundant. The composition of most grains is An_{28}.

Biotite: X = pale brown, Y = Z = dark brown; forms thin, semicontinuous trains parallel to S_1, or well developed discrete plates also parallel to schistosity.

Accessory minerals include chlorite (partly altered biotite), muscovite, apatite and iron ores.

A285/3 Quartzo-feldspathic Schist

Light grey, rather fine grained rock; quartz-feldspar matrix are well crystallised. The grain size is less than 2 mm.

Quartz: a few grains display undulose extinction.

Potash feldspar: a few scattered grains of untwinned feldspar.

Plagioclase: mostly untwinned; some grains have albite law twinning. The composition is An_{28}.

Biotite: X = light brown Y = Z = dark brown. Well developed discrete plates forms a strong schistosity (S_1).

Muscovite: rather scarce.

Chlorite: laths and irregular patches replace biotite.

A285/122 Quartz-plagioclase-actinolite schist

A light coloured coarse grained rock with granoblastic texture. The grain size ranges from 0.2 mm to 0.5 mm.

Quartz: occurs as a subrounded to irregular shaped grain; some grains show triple point junctions. Strain is common.

Plagioclase: is usually twinned on albite law; untwinned grains are also present. Most of the grains are fresh, but a few show extensive sericitization. Strained grains are frequently seen. Small inclusions of quartz, plagioclase and actinolite are present in some grains. The composition of plagioclase is An_{27}.

Actinolite:occurs as prismatic crystals (0.2 - 0.5 mm in cross section). Polysynthetic twinning is well developed in a few grains.

Sphene: is found as colourless subhedral to anhedral crystals.

Rutile: is golden brown euhedral to anhedral crystals. It has a grain size from 0.1 mm to 0.3 mm.

A285/109 Quartz-plagioclase-actinolite schist

A light coloured coarse grained rock with granoblastic texture. The grain size ranges from 0.2 mm to 0.8 mm.

Quartz: occurs in grains of variable size, from tiny bleb-like inclusions (0.1 - 0.2 mm) to strained aggregates up to 0.5 mm. in size.

Plagioclase: is generally untwinned and strained. Clouding and sericitization is present in some grains. The composition of plagioclase is An_{34}.

Tremolite: is found as colourless prismatic grains, some with very weak pleochroism.

Rutile, sphene and apatite are common accessory minerals.

A285/454 Quartz-plagioclase-actinolite gneiss

A light coloured coarse grained granoblastic rock with average grain size 2.5 mm.

Quartz: ranges in size from 0.3 mm to 3.2 mm. The larger grains are strongly strained and show undulose extinction.

Plagioclase: is usually twinned on the albite law. Untwinned plagioclase is abundant. Deformation (glide) twins with bent twin lamellae are seen in some grains. The composition of plagioclase is An_{29}.

Actinolite: pale green, moderately pleochroic prismatic crystals. It ranges in size from 0.2 mm to 0.4 mm in cross section.

Sphene and apatite are common accessory minerals.

A285/112 Quartz-plagioclase-tremolite gneiss

A light coloured coarse grained rock with a granoblastic texture and average grain size 2.5 mm.

Mineralogically, it is similar to A285/109 except for the coarser grain size. The composition of plagioclase is An_{30}.

A285/123 Quartz-plagioclase-actinolite schist

A light coloured, granoblastic coarse grained rock. The average grain size ranges from 0.5 mm to 1 mm.

Quartz: appears as an irregularly shaped interlobate grain with undulose extinction.

Plagioclase: is generally twinned on the albite law, but untwinned grains are not uncommon. The composition of plagioclase is An_{30}.

Actinolite: occurs as a pale greenish, moderately pleochroic, prismatic mineral. It ranges from 0.2 mm to 0.8 mm in cross section.

Sphene, rutile and apatite are the common accessory minerals.

A285/173A Quartz-plagioclase-hornblende schist

A light coloured, granoblastic rock with average grain size 0.5 mm. A strong foliation is defined by the concentration of hornblende prisms in thin bands orientated parallel to S_1.

Quartz: is present in variable shapes and sizes. Its grain size ranges from 0.2 mm to 0.6 mm. Most grains are elongated parallel to the S_1 schistosity. Strain and undulose extinction are most commonly seen.

Plagioclase: is generally untwinned, but some grains are twinned on albite laws. The composition of plagioclase is An_{27}.

Hornblende: X = pale green, Y = Z = bluish green. It has a prismatic to equant shape with a range of grain size from 0.1 mm to 0.5 mm.

Phlogopitic biotite, sphene, apatite and iron ores are common accessory minerals.

A285/173B Quartz-plagioclase-hornblende gneiss

A light coloured granoblastic rock with grain size 0.3 mm to 2 mm. Hornblende is aligned parallel to S_1 schistosity.

Quartz: has variable sizes. The general shape is amoeboidal with undulose extinction.

Plagioclase: is generally untwinned, but a few grains are weakly twinned on the albite law. Most of the twinned grains are strained. Composition cannot be measured due to lack of good twinning and to strain effects in twinned grains.

Hornblende: is a pale green to bluish green mineral. It has a rather skeletal shape, but some grains show a euhedral prismatic habit.

Phlogopitic biotite, sphene, apatite and iron ores are common accessory minerals.

A285/150 Quartz-plagioclase-hornblende gneiss

A pink coloured granoblastic rock with average grain size of 1 mm. Well crystallized quartz generally has curved boundaries. A weak preferred orientation parallel to S_1 is shown by hornblende.

Quartz: grain size varies considerably from 0.3 mm to 1.5 mm. Strong strain is common.

Plagioclase: is found both as twinned and untwinned grains. The composition of plagioclase is An_{32}.

Hornblende: is pleochroic from pale green to bluish green. It has a well crystallized prismatic habit but a few occur as irregular poikilo-blasts.

Apatite, sphene and iron ores are common accessory minerals.

A285/145 Quartz-plagioclase-anthophyllite schist

A light coloured granoblastic rock with an average grain size of 0.4 mm.

Quartz: (0.2 - 0.5 mm.) has straight and curved boundaries. Strain is common.

Plagioclase: (0.3 - 0.5 mm.) is present both as twinned and untwinned grains. The composition of plagioclase is An_{32}.

Anthophyllite: occurs as a euhedral prism or as fibres.

Biotite: is a weakly pleochroic, pale yellow, flaky mineral. Optically it appears to be a magnesian biotite (phlogopite).

Rutile, sphene and apatite are common accessory minerals.

A285/118 Quartz-plagioclase-diopside gneiss

A light coloured coarse grained rock. It has a granoblastic texture with a grain size of 0.5 mm to 2.5 mm.

Quartz: has an irregular shape and shows strong strain effects. Some grains contain small inclusions of sphene and subhedral zircon.

Plagioclase: is generally fresh, except for some grains which are strongly clouded with small granules of weakly pleochroic epidote. It shows twinning on a variable scale, from good albite law twins to patchy, discontinuous twin lamellae. Untwinned grains are also abundant. Some grains are weakly strained. Deformed and gently bent, often displaced twin lamellae, are also seen. A weak normal zoning is present in a few grains. The composition of plagioclase is An_{34}.

Diopside: is found as skeletal shaped xenoblasts (1-2.5 mm.). Small grains up to 0.3 mm. to 0.5 mm. are also abundant. It contains numerous inclusions of sphene and quartz.

Hornblende: is rather scarce occurring as small granular masses, some associated with diopside. It is pale green to green in colour with strong pleochroism.

Sphene: is generally in subhedral to anhedral shaped crystals. It generally clusters around diopside.

Apatite, rutile and zircon are the common accessory minerals.

A285/178 Quartz-plagioclase-diopside-actinolite schist

A light coloured granoblastic rock with an average grain size of 0.5 mm. Good foliation is developed due to concentration of diospde/actinolite and quartz/plagioclase in alternate bands. Preferred orientation of actinolite and lenticular quartz with dimensionally oriented diopside define the S_1 schistosity. The matrix is well crystallized.

Quartz: has straight to curved boundaries. Strain is common in most of the grains.

Plagioclase: is generally untwinned, but some twinned grains are present. Inclusions of quartz, sphene and diopside may be seen in some grains. The composition of plagioclase is An_{34}.

Diopside: occurs as irregularly shaped small aggregates (0.1 - 0.2 mm.) and as large skeletons (up to 0.5 mm.). It has a light greenish colour.

Actinolite: is weakly pleochroic from pale green (X) to green (Z). It appears as interstitial grains between quartz/plagioclase matrix. It is rarely associated with diopside.

Sphene: is generally found as an anhedral shaped colourless mineral. It is more common in the quartz/plagioclase matrix rather than in association with ferromagnesian minerals.

Apatite and rare rutile are the only accessory minerals.

A285/121 Quartz-plagioclase-diopside gneiss

A light coloured granoblastic rock with average grain size ranging from 0.5 mm. to 2.5 mm. The matrix is well crystallized.

Quartz: is extremely variable in grain size ranging from 0.2 mm. to 2.8 mm. Generally, it appears as interlobate shaped grains. The matrix is mainly composed of medium sized (0.5 - 1.5 mm.) quartz grains. Undulose extinction is more common in bigger grains.

Plagioclase: is equally abundant both as twinned and untwinned grains. A few grains show deformational features such as glide twinning or slightly bent twin lamellae. Moderate strain may be seen in a few grains. The composition of plagioclase is An_{34}.

Diopside: is found as large skeletons pale green in colour. It has an average grain size in the range of 0.5 mm to 3 mm, but some grains may be much smaller in size. Small inclusions of sphene and quartz are common.

Sphene: is a brown coloured mineral. It is seen as euhedral lozenge shaped crystals or as large irregularly shaped granules.

Among the accessory minerals, apatite and subhedral zircon are notable.

A285/373 Quartz-plagioclase-diopside gneiss

A light coloured granoblastic rock with grain size in the range of 0.5 mm to 3 mm. The matrix is well crystallized and essentially composed of variable sized grains of quartz and plagioclase. Ragged crystals of diopside appear interstitially between quartz and plagioclase.

Quartz: is extremely variable in shape and size (0.1 - 3 mm.). Undulose extinction is common.

Plagioclase: is generally well twinned on albite law. Some untwinned grains are also present. Glide (deformation) twinning with gently bent twin lamellae is frequent. A few grains are moderately zoned with diffuse boundaries. Inclusions of sphene, diopside and quartz are found in some bigger grains. The composition of plagioclase is An_{34}.

Diopside: is found as a colourless ragged crystal in variable sizes. The most common grain size is from 0.5 mm. to 2.5 mm., except for a few which are smaller. Numerous inclusions of sphene are common in bigger crystals.

Sphene: is a neutral coloured mineral with a subhedral to anhedral shape. Most commonly, it clusters around or is associated with diopside.

Apatite is the only notable accessory mineral.

APPENDIX 2

APPENDIX 2

TECHNIQUE AND RELIABILITY OF MODAL ANALYSES

Modal analyses have been done on the leucocratic veins and the host rock of migmatites. Due to the presence of some untwinned plagioclase and of potash feldspar, all thin sections were stained. The staining was done on uncovered thin sections by etching with hydrofluoric acid and then potash feldspar was stained yellow with sodium cobaltinitrite and plagioclase red with amaranth solution (see Laniz et al., 1964). After staining, thin sections were covered in the normal manner.

Thin sections used for modal analyses were cut normal to foliation and lineation. The counting on thin sections was carried out with a Swift point-counter. Only one thin section was used for point counting. Therefore, the maximum area was covered depending upon the uniformity of the grain size. The leucocratic veins are generally coarse grained with a large variation in grain size. Therefore the number of counts varied from 5,000 to over 10,000. The gneissic host rocks of the migmatite are finer grained and uniform in grain size, and the number of counts ranged from 1,500 to 2,300. Hence, consideration was given to the size of the area with relative variation of grain sizes which could give low counting variations (e.g. Chayes, 1956; Bayly, 1965).

The results of modal analyses are presented in Table 2, sect. 5, 7.2. It is believed that the point-counter analyses give a reasonable estimate of the volume percentage of individual minerals. However, inhomogeneity of the leucocratic vein of migmatite is such that a single random slab represents only the composition of the portion of the vein from which it was cut. However, the values for the various veins are not that dissimilar and it can be asserted that the values quoted truly represent the veins examined.

APPENDIX 3

SELECTION & PREPARATION OF ROCK & MINERAL SAMPLES

SELECTION OF SAMPLES

Because of inhomogeneity it was difficult to analyse the bulk migmatites. Attention was confined to the leucocratic veins. These veins were separately analysed to obtain data as to the origin of the quartzo-feldspathic material (see Chapter 6). The samples which were analysed come from the leucocratic veins of migmatites showing the following characters:

Sample No.	Thickness of leucocratic vein of migmatite	Remarks
372	1.5 inches	Biotite selvedges absent; exposed length of vein 4 feet.
377	6 inches	Biotite selvedges absent; exposed length of vein 4.5 feet.
398	1 foot	Biotite selvedges absent; exposed length of vein 12-15 feet.
515	1 foot	Thin biotite selvedge present; exposed length of vein 3-4 feet.
514	8 inches	Biotite selvedges present; exposed length of vein 3-4 feet.
479	2-3 inches	No biotite selvedges; folded vein
474	.3-2 feet	No biotite selvedges; pinch and swell structure.
483	6 inches	No biotite selvedges; exposed length of vein 2 feet.
497	2-8 inches	No biotite selvedges; pinch and swell type vein.
608	3 inches	No biotite selvedges; exposed length of vein 3 feet.
705	3-4 inches	Thin biotite rich selvedges present; exposed length of vein 2 feet.

SAMPLE PREPARATION

All rock samples used for chemical analyses were slabs cut at right angles to foliation and lineation. Any weathered material on the slab was trimmed off on a diamond saw and then the surface ground with carborundum powder. The sample was washed with water to remove dust and other loose materials. It was soaked in a 10% HCl solution for 10 minutes and finally washed with distilled water. The rock slab was dried at 110°C for 2 hours.

The rock slab weighing about 250-500 grams was broken into small pieces in a fly press with stainless steel plates. Care was taken to avoid the significant loss and contamination of samples during crushing in the fly press. The broken pieces of each sample were ground in a "Siebtechnik" mill for 30 seconds to 1 minute and then passed through a 120 mesh nylon sieve. After repeated grinding in the mill and the last remaining coarse fraction in agate mortar by hand, all samples passed through the 120 mesh nylon sieve. It is believed that during grinding in the "Siebtechnik" mill the contamination was insignificant from the grinding vessels. The chrome steel vessels and annular grinding rings of the "Siebtechnik" mill have the composition 1.7% C, 0.3% Si, 0.35% Mn, 12.0% Cr, 0.12% V and 85.53% Fe.

The -120 rock powder was thoroughly mixed and divided on a sample splitter. The rock was split into two portions; one sample was retained for bulk rock analyses and the other was used for mineral separation.

The rock powder to be used for chemical analyses was again thoroughly mixed and about 5 grams were taken for finer grinding. This sample was ground for about 2-3 hours with acetone in a "Fisher" mill. This fine powder was used for wet-chemical and X-ray spectrographic analyses after drying overnight at 110°C.

MINERAL SEPARATION

Mineral separation was done using -120+260 sieved rock powder. This powder was repeatedly washed with water to remove dust and fine particles sticking to -120+260 sized grains. Finally, it was washed with distilled water and then acetone before drying under an infrared lamp.

Biotite, clinoamphibole and clinopyroxene were extracted from the rock powder using a Frantz Isodynamic Separator. Generally, a concentrate of biotite with a purity up to 99.5% and clinopyroxene and clinoamphibole with a purity from 97-99% were obtained by several runs through the separator. Further purification was done by hand picking when the impurity of quartz, plagioclase, sphene and rutile exceeded more than 2%. All biotite samples had less than 1% impurity (quartz and/or plagioclase) except Sample No. 377 (2% quartz and plagioclase). The clinoamphibole and clinopyroxene samples were about 98-99% pure. The following minerals contained impurities either in the form of inclusions (sphene and rutile) or other grains.

Sample No.	Impurities in mineral
117 (actinolite)	1% quartz and sphene
122 (")	1% sphene and rutile
109 (tremolite)	1% quartz
454 (actinolite)	1% quartz
112 (tremolite)	2% sphene and quartz
123 (actinolite)	0.5% quartz
173A (hornblende)	1% quartz and sphene
173B (")	1% biotite and sphene
150 (")	2% quartz and sphene
145 (anthophyllite)	1% rutile and quartz
121 (salite)	1% quartz and actinolite
178 (diopside)	1% sphene and quartz
373 (")	2% quartz
118 (salite)	1% quartz and actinolite

These represent maximum limits of impurity, and it was not felt justified to adjust the analysis of the minerals on the basis of these impurities.

All mineral samples were ground in a "Fisher" mill with acetone, biotite for about 3 hours per gram of sample and clinoamphibole and clinopyroxene for about 1.5 hours. The mullite mortar and pestle were cleaned with quartz powder for half an hour after each sample.

For trace element work, biotite powders were checked under the microscope to ensure that the grain size was less than 50 microns. In general, the average grain size of powder was less than 25-30 microns.

ANALYTICAL TECHNIQUES

Major elements

The elements Si, Al, Ca, Ti, Fe, P and Mn were determined by X-ray fluorescence spectrography. Fused buttons were prepared of rocks and minerals. Norrish and Chappell (1967) and Norrish and Hutton (1969) have described the technique for preparation of a button (glass disc) by fusing the sample with a mixture of lithium borate and lanthanum oxide. These glass buttons were used as samples for X.R.F. analyses against an artificial standard (FS 11) supplied by Dr. Norrish, C.S.I.R.O., Division of Soils, Adelaide, South Australia. Two samples of standard rocks, GSP I and G_1, were also analysed as unknown samples to compare the results with the published data of Flanagan (1966, 1969) and Carmichael et al. (1968).

The technique for rock and mineral analyses by X.R.F. method was followed after Norrish and Chappell (1967) and Norrish and Hutton (1969) using a standard Philips generator, analysing and counting equipment. The following conditions were used for quantitative determinations of listed elements:

Operating conditions for X-ray fluorescent spectrography

Element determined	Primary radiation	Analysing crystal	Gas used
Si	Cr	P.E.T.	Ar-10% CH_4
Al	"	"	"
K	"	"	"
Ca	"	LiF200	"
Ti	"	"	"
Fe	"	"	"
Mn	Mo	"	"
P	Cr	G.E.	"
Flow proportion counter used for all determinations			

Alkalis (Na_2O and K_2O) were determined by flame photometer. About .5 grams of rock sample and .25 grams of mineral sample was dissolved in hydrofluoric and perchloric acids in platinum crucibles following the scheme of Riley and Williams (1959). Four to five standard solutions of Na_2O and K_2O were used to calibrate the flame photometer with a batch of six unknown solutions. A blank solution was run with each batch of samples.

MgO was determined by atomic absorption spectrophotometry according to the scheme of Nesbitt (1966). The same solutions were used as for the alkali determinations. A set of three to four standard solutions of MgO were used for calibration. Operating conditions for the spectrophotometer were kept as follows during the MgO determinations.

Lamp filament current : 4 mA
Wavelength : 2855 Å
Slit width : 50 microns
Flame : Rich N_2O/C_2H_2

FeO was determined by titration of the standard "Pratt" method. Duplicate determinations were done for each sample. The total iron as determined by X.R.F. spectrography was corrected for the FeO content and the difference gave the Fe_2O_3 value.

LOSS ON IGNITION

Loss on ignition for rocks and minerals was determined by heating the samples in an electric furnace at 1,000°C at constant temperature. The weight of samples taken was about .5-1 grams and it was heated for 2 hours.

TRACE ELEMENT ANALYSES

Rb, Sr and Ba analyses were made both for total rock and biotites by X.R.F. methods (Norrish & Hutton, 1969). A flat circular pellet was prepared by using 1-2 grams of fine grained sample encased by borax powder and compressed at a pressure of 4 tons/sq. in. by a hydraulic press (Norrish & Chappell, 1969).

The following instrument settings were used for Sr and Rb determinations:

Analysing crystal LiF_{220}
Molybdenum tube at 44KV/20mA
Scintillation counter with coarse collimator
Fixed count accumulation 4×10^4 counts for Sr and 2×10^4 counts for Rb peaks respectively
K_α spectral line for each element

The following 2θ positions were measured:

Background	Sr peak	Background	Rb peak	Background
34.61	35.67	36.73	37.79	38.85

The non-linear correction for background was less than 0.5% for Sr and about 2% for Rb. This correction was applied to Rb only.

Rb and Sr determinations were made against the geochemical standard G_1. The following values were used for G_1.

Sr	Rb	
247 ppm	210 ppm	(determined by isotope dilution method, A.N.U)
μ8.59	10.05	(A.U.G.D. figures)
μ8.64	9.93	(measured)

Mass absorption corrections were applied to all samples; some were measured and some were calculated from the analysis of the rocks (major elements) using tables by Champion et al. (1968)*.

Mass absorption (μSr and μRb) were measured on the following samples using the same powder as was used to make the pellets.

Sample No.		μ Sr	μ Rb
514	(total rock)	8.54	9.60
140	"	8.28	9.78
295	"	8.35	9.56
137	"	8.67	9.96
232	"	8.06	9.51
149	"	8.40	9.56
205	"	11.68	12.27
372AG	"	8.89	9.83
139	"	7.69	8.48
138	"	8.65	9.58
497	"	8.18	-
140	(biotite)	16.82	19.66
75	"	17.68	20.67
233	"	17.58	20.09
514	"	15.55	21.99
382AG	"	17.62	20.51
149	"	15.56	24.27

The following instrumental settings were used for Ba determinations.

Analysing crystal LiF_{200}
Chromium tube 60KV/40mA
Flow proportional counter with coarse collimator
Count for fixed time, 20 seconds
Lβ2 spectral line

Correction was applied for non-linearity of background. Mass absorption correction was calculated from the major oxides analysis for all samples.

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* CHAMPION, K.P., HURST, H.J. & WHITTEM, R.N. (1968): Tables of Mass absorption coefficients for use in X-ray spectrochemical analysis. Australian Atomic Energy Commission, Lucas Heights.

GSP I with a value of 1,360 ppm for Ba (Flanagan, 1969) was used as a primary standard. Ba was measured at the $L\beta_2$ peak in preference to K_α because of interference of the Ti peak with the Ba peak. Reproducibility for Ba determination at the $L\beta_2$ peak is reasonably good as is shown below.

Ba in ppm

	This method	Flanagan, 1969	Carmichael et al., 1969
G_1	1067	-	1040
G_2	2031	1950	2030

NORM CALCULATIONS

C.I.P.W. Norm

The C.I.P.W. norms were calculated on a computer programme written at the University of Adelaide and based on the scheme of Kelsey (1965)*. This is the conventional norm calculation modified to be used on a computer and using the 1961 atomic weights.

Mesonorm

The mesonorms were calculated on a computer programme written by Dr. A. W. Kleeman. The rules used in this programme differ from those of Barth (1962) in some respects:

1. If the cation percentages of Na + K + 2Ca exceed Al, then Ti is allotted to sphene - as in Barth's scheme - but where Al is less than Na + K + 2Ca, then Ti is calculated as ilmenite.
2. The excess Al is calculated as sillimanite rather than corundum. In the case of muscovite bearing rocks, muscovite has been calculated as 5Or + 2Al = 7 Musc., and the remaining Al has been allotted to sillimanite.

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*KELSEY, C. H. (1965) : Calculation of the C.I.P.W. Norms. Min. Mag. 34, 276-282.

Following the modified Barth mesonorm of Kleeman, a few discrepancies may result from the actual mineralogical composition of the analysed rocks. The following points are important in comparing the mesonorms with the mineralogy of the rocks:

1. The conversion of normative sillimanite to muscovite generally results in a lower quantity of normative orthoclase. Also the calculated muscovite is always found to be higher in amount than its modal percentage.
2. Generally, sphene is under-estimated and normative ilmenite is over-estimated according to the present method of norm calculations compared with modal analyses.

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